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KATABATIC WIND FORCING OF THE TERRA NOVA BAY POLYNYA

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Abstract. The Terra Nova Bay polynya is a perennial winter feature in the western Ross Sea, Antarctica, which occupies roughly 1000 km^2. It is formed and maintained by the combined influence of persistent katabatic winds, which advect newly formed bay ice eastward, and the Drygalski Ice Tongue, which prevents northward drifting pack ice from entering Terra Nova Bay. Existence of anomalously strong katabatic drainage along this coast is predicted by Parish's (1982) simulation of wintertime airflow which reveals a pronounced confluence of surface winds upslope from the Reeves Glacier where the winds are further focused by local topography. The simulation is strongly supported by regional sastrugi orientations. Average wintertime atmospheric conditions and ice sheet topography which control surface air drainage are stable on a climatic time scale; therefore, persistent wintertime katabatic winds should be an annual phenomenon. Further evidence comes from multi-year Landsat images which consistently show windswept, snow-free areas on the Reeves Glacier. In marked contrast to typical Antarctic katabatic winds, strong persistent winter winds are observed at sea level ~25 km beyond the coastal slope break. Air probably descends as bora-type winds and is likely to be significantly denser than the air at sea level; conditions are not favorable for hydraulic jumps apparently typical of other katabatic regimes. The horizontal density difference is maintained during airflow across the Nansen Ice Sheet because relatively little air mass modification occurs there in contrast to situations where air moves over an ice-laden ocean. Observations thus suggest that katabatic winds maintain their identity for some distance seaward of the coast; qualitative trajectory calculations indicate that for representative geostrophic conditions this distance is on the order of the observed polynya width. Estimated ice freezing rates are ~20 cm day^{-1}, but wind-generated waves and currents prevent ice from consolidating. The blocking effect of the Drygalski Ice Tongue is a consequence of its orientation with regard to western Ross Sea ice drift patterns; ice tongue length controls polynya width. Absence of such blocking along other coasts experiencing strong katabatic outflow partly explains why similar polynyas do not form there.

Introduction

Wintertime open water in Terra Nova Bay, western Ross Sea, has been observed since 1912 when Scott's northern party was stranded at Inexpressible Island [Bromwich and Kurtz, 1982]. It was not until the advent of infrared satellite imagery, however, that the existence and extent of the polynya were realized. Knapp [1972] was probably first to report it, followed by Szekielda [1974]. Polynya behavior during 1979 has been documented by Kurtz and Bromwich [1983], who suggested yearly recurrence. Its area averaged 1300 km^2, ranging from 0 to 5000 km^2.

Several mechanisms have been proposed to explain polynya formation. Knapp [1972] hypothesized that it was one of a class of polynyas that form when strong winds associated with passing cyclones generate net sea ice displacement away from fixed barriers affecting sea ice drift. Analysis of polynya areal fluctuations and concurrent geostrophic conditions [Kurtz and Bromwich, 1983] indicate that synoptic scale motions serve only to modify its extent; they are not a primary cause of its formation or maintenance. Szekielda [1974] postulated that marine upwelling or submarine volcanism causes bay waters to be too warm to freeze, but oceanographic and marine geologic data contradict this. In Terra Nova Bay the summer water column is nearly isothermal at the sea surface freezing point (stations 168 and 169 in Jacobs and Haines [1982]) and will be similar in winter, and there is no evidence of submarine volcanism in piston cores [Anderson and Kurtz, 1980] or 3.5 kHz bottom profiles from this area.

The men of Scott's northern party felt strongly that the "plateau wind" which blew for months on end [Priestley, 1914] was responsible for the open water; possible influence of descending air was postulated by Kurtz and Bromwich [1983] based upon observations in infrared satellite images of a recurring thermal signature in the Reeves Glacier valley west of Terra Nova Bay (Figure 1).

This warm signature was often associated with a large or expanding polynya. A more complete model is proposed and demonstrated in the present paper; it is outlined schematically in Figure 2. The Terra Nova Bay polynya is formed and maintained through the combined influence of persistent katabatic winds, which advect bay ice eastward and prevent it from consolidating in the polynya, and the blocking action of the Drygalski Ice Tongue, which prohibits northward drifting pack ice from entering the bay. No long-term atmospheric or sea ice observations exist for Terra Nova Bay, but using theoretical arguments, explorers' observations, and regional information we demonstrate (1) confluence of surface air on the plateau west of Terra Nova Bay; (2) descent of this air to sea level, as katabatic winds which drain principally through the Reeves Glacier valley; (3) stability of katabatic outflow seaward of the coastal slope break, for distances comparable to observed polynya width; (4) that bay waters maintain a high sea surface heat flux and that wind-generated wave and current activity prevents sea ice from consolidating; (5) blocking action of the Drygalski Ice Tongue; and (6) that the above recur each year.

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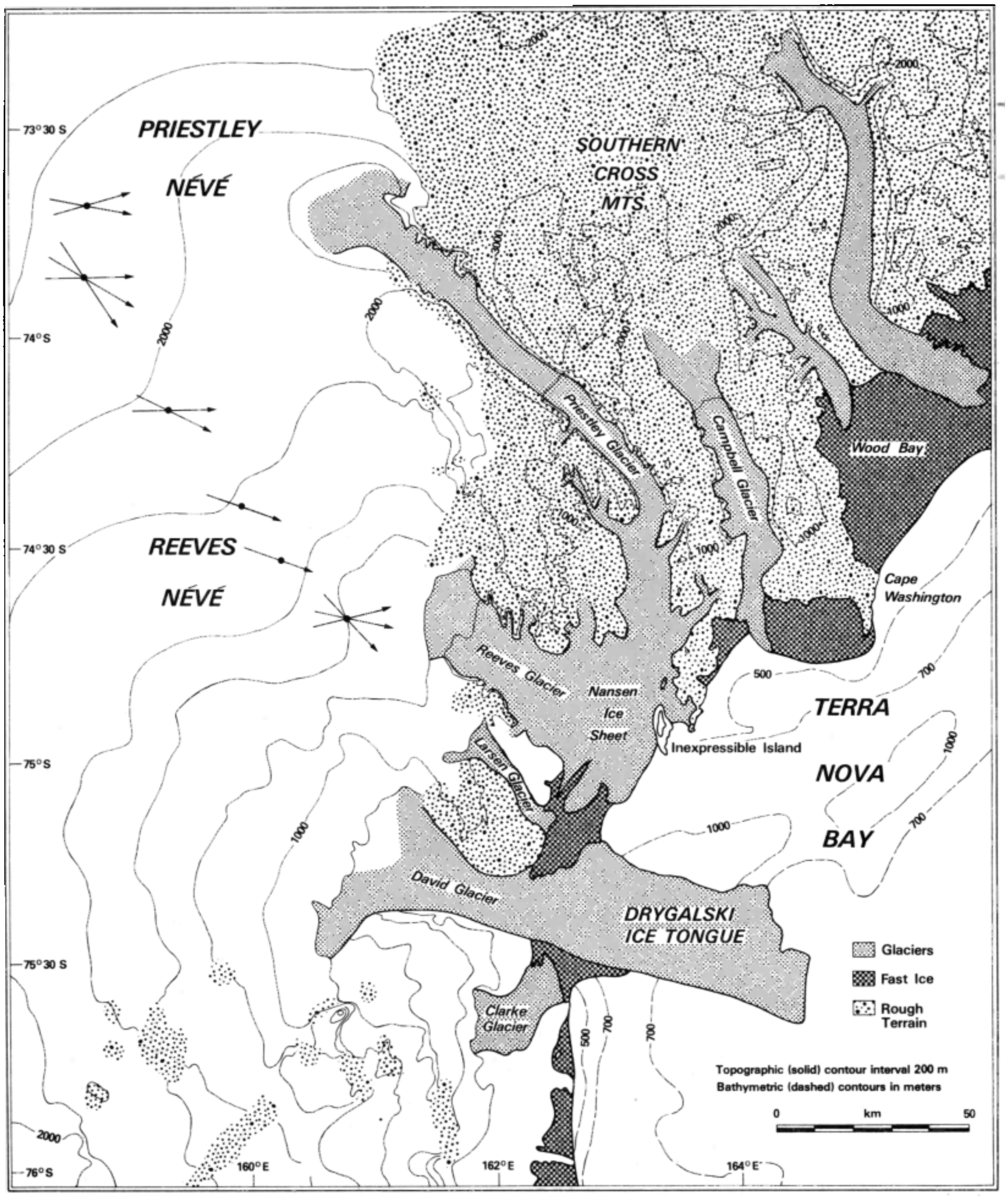

Fig. 1. Location and topographic map. Sastrugi directions (short arrows), indicating the prevailing wind direction, are from David and Priestley [1914].

This is the first instance where a primary role of katabatic winds in forming an Antarctic polynya can be demonstrated.

Katabatic Drainage

Parish [1982] has modeled the mean winter surface airflow over a substantial fraction of the east Antarctic ice sheet above 2000 m elevation. Calculations are based upon integrated boundary layer equations which express a steady state balance between the pressure gradient force generated by the sloped surface temperature inversion, the Coriolis effect, and friction. The influence of free atmospheric pressure gradients upon the time-averaged state was inferred to be of secondary importance. Input data consisted of the boundary layer temperature structure and ice sheet topography. Explicit mass conservation was not required; thus only a qualitative picture of the drainage mass flux is provided. Figure 3 depicts the resulting streamline field for that part of the ice sheet in the vicinity of Terra Nova Bay. Based upon studies of Cape Denison, where very strong katabatic winds are well documented [Mawson, 1915], Parish [1981, 1982] has argued that a prerequisite for strong, persistent coastal katabatic winds is a large cold air reservoir which drains through a relatively narrow outlet.

The confluence west of Terra Nova Bay (draining about 3% of east Antarctica), is postulated to deliver air to the catchment of the Reeves Glacier. To test the local validity of this surface flow field, all available sastrugi information [David and Priestley, 1914; Stuart and Heine, 1961] has been added to Figure 3. Sastrugi are erosional snow surface features, usually observed during the summer, whose long axes are parallel to the prevailing wind [Mather, 1962, 1969; Watanabe, 1978]; they persist for a few months. Although the generating, topographically controlled surface winds weaken between winter and summer, their unidirectional character is maintained (compare, for example, Ohata et al.[1981]). Therefore, sastrugi orientations observed during summer probably reflect the winter surface wind regime. Because these Antarctic surface winds are determined by winter radiation conditions and ice sheet topography [Parish, 1982], sastrugi data from different years can be combined; both controls are likely to be essentially invariant on the decadal time scale [e.g., Jouzel et al., 1983]. The fit in Figure 3 between the two sets of sastrugi directions (collected about 50 years apart) and the modeled surface wind pattern is very close. In particular, the area labeled as the "Parting of the Winds" [David and Priestley, 1914] is well located by Parish's simulation. T. W. Edgeworth David's party, on their treck to the south magnetic pole during the austral summer of 1908-1909, observed that the prevalent wind direction in this area switched from W or NW to the east of the ridge line around to SE on its northwestern side. The streamlines in Figure 3 thus appear to provide a reliable description of the time-averaged winter surface wind on the plateau.

Figure 3 shows that the streamlines tend to cluster in the northern part of the concave elevation contours west of Terra Nova Bay. This result agrees with Ball [1960] who found, for similar topography, that katabatic flow lines concentrate on the left slope of a valley (looking downwind). Westerly winds are calculated along the 1700 m contour to the west of the bay. Examination of the terrain immediately inland of Terra Nova Bay (Figure 1), in conjunction with the modeled drainage, suggests preferred sites for katabatic winds to descend to sea level. The location and width of the entrance to the Priestley Glacier valley, and its orientation with respect to the angle of incidence of the katabatic flow, indicate that it is unlikely to be a major outlet. The steep mountains located southwest of the Priestley Glacier (with individual peaks rising to over 2800 m) probably block the impinging stable cold air and deflect it to the right down the Reeves Glacier. This deflection, the location of the Reeves catchment basin in the region of wind confluence, and valley orientation make it likely that most of the cold air drains through this glacier. Ball [1956] describes the dynamics of a katabatic wind which is constrained by local topography to flow directly downslope. Air tends to pile up against the obstruction, and the variation in the depth of cold air leads to winds which are strongest close to the obstacle and diminish with distance from it; this phenomenon on sloped-terrain is analogous to Schwerdtfeger's [1974] barrier winds. The David Glacier lies on the southern side of the katabatic confluence region and thus is probably exposed to a smaller mass flux than

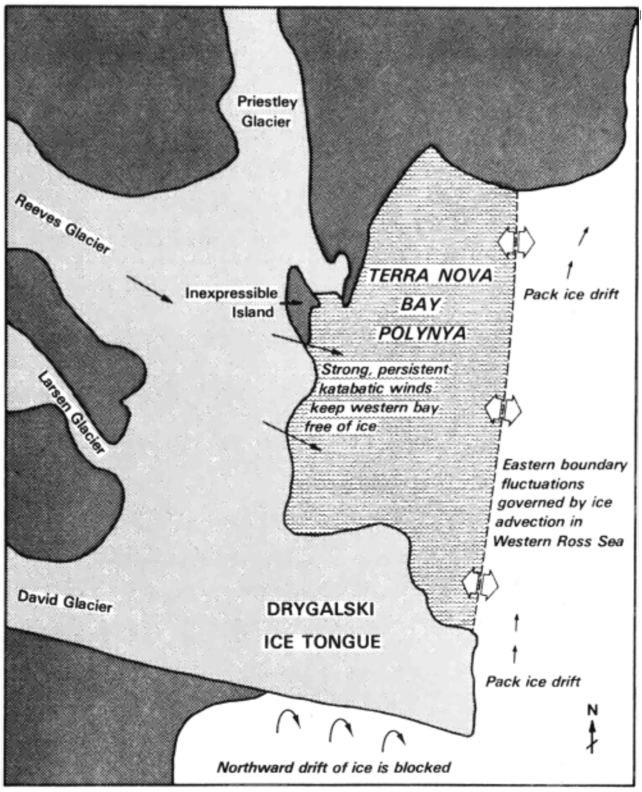

Fig. 2. Schematic illustration of the proposed mechanisms which force the open water in Terra Nova Bay.

the Reeves Glacier. In summary, therefore, strong, persistent winter katabatic winds should blow down the Reeves Glacier. Less intense katabatic flow is expected from the David Glacier.

Strong Katabatic Winds in Western Terra Nova Bay

Our analysis of the experiences of Scott's northern party [Bromwich and Kurtz, 1982], who were stranded on Inexpressible Island (Figure 1) during the winter of 1912, indicates that katabatic winds from the Reeves Glacier cross the Nansen Ice Sheet and enter Terra Nova Bay from the WNW with a representative speed of 15 m s^{-1}. This estimate refers to the unimpeded flow crossing the edge of the ice sheet to the south of Inexpressible Island. No other in situ winter data are available for this area.

Figure 4 is presented as additional evidence supporting the existence of strong katabatic winds blowing across the Nansen Ice Sheet. The feature of interest on this Landsat MSS image is the low albedo (dark) of the sheet ice for solar radiation at these wavelengths (band 7: 0.8 to 1.1 μm). This signature is barely perceptible in band 4 (0.5 - 0.6 μm) and steadily becomes more prominent as the wavelength increases. Such relative wavelength behavior occurs when the spectral albedo for old melting snow or bubbly ice is compared with that for new snow [Dozier et al., 1981; S. Warren, personal communication, 1983]. Various band 7 images between November 28, 1972, and January 14, 1974, show that, although there is considerable variability in the

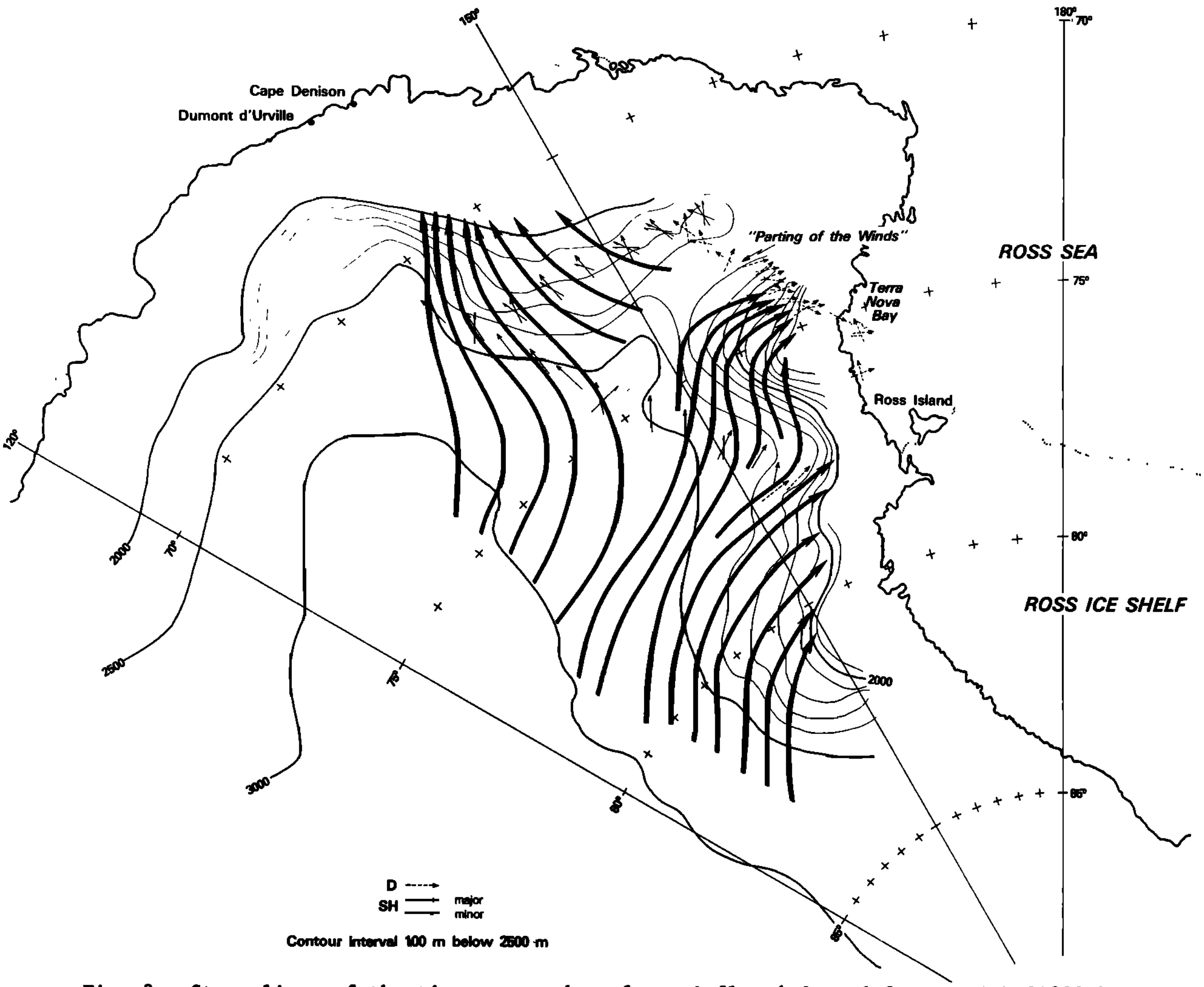

Fig. 3. Streamlines of the time-averaged surface airflow (adapted from Parish [1982]) supplying the strong katabatic winds which blow from the Reeves Glacier. Sastrugi directions labeled with a D were taken from David and Priestley [1914] and with SH come from Stuart and Heine [1961]. The former are also plotted on Figure 1.

surface conditions (depending upon such things as the time since the last appreciable snowfall and the intervening wind and temperature conditions), the main features observable in Figure 4 are still obvious. In addition, D. N. B. Skinner (personal communication, 1983) reports that U.S. Navy air photographs, from several austral summers, show the area of low band 7 albedo to be bare ice with snow to the north and south. The ice/snow boundaries are sharp and nearly coincide with the low/high albedo boundaries shown in Figure 4. The evidence thus indicates that the prominent Nansen Ice Sheet signature is persistent in time.

We suggest that the low albedo of the Nansen Ice Sheet in the near infrared arises because strong katabatic winds from the Reeves Glacier sweep the surface clean of new snow (compare Williams et al. [1983]). The only occasions in the 1912 winter when drifting or blowing snow was observed in the vicinity of Inexpressible Island was in conjunction with snowfall. Substantial winter ablation of the ice sheet surface should occur by sublimation into the (probably) dry, fast moving airstream. Less reflection of incident shortwave radiation means more absorption and perhaps enhanced summer melting. Extensive melting is observed on the Nansen Ice Sheet during summer [Priestley, 1974; D. N. B. Skinner, personal communication, 1981], and meltwater lakes are entered on the USGS map of this area (Mt. Melbourne SS 58-60/9). By contrast, the men of Scott's northern party, throughout their stay at Inexpressible Island, frequently observed clouds of blowing snow over the Drygalski Ice Tongue. Loewe [1974] pointed out that descending air will evaporate blowing snow particles (compare Gosink [1982]), and thus maintain a comparatively high relative humidity with respect to ice. This factor in combination with less intense katabatic drainage suggests that complete snow removal and strong evaporative ablation on the undulating Drygalski Ice Tongue are unlikely. A higher band 7 albedo is the end result. However, no explanation can be offered for the very different character of katabatic drainage down these neighboring glaciers.

Figure 4 shows that the change in band 7 albedo

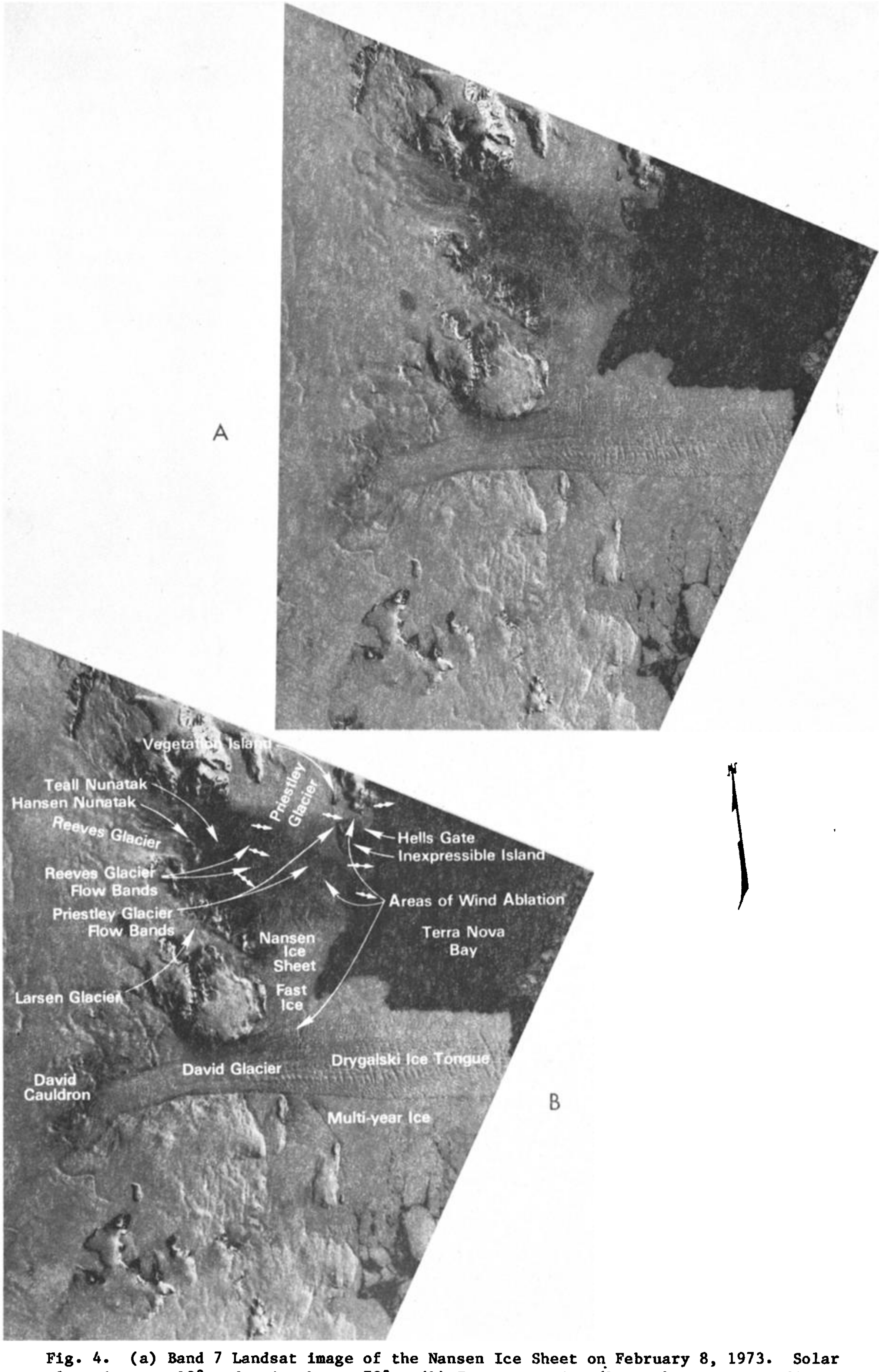

Fig. 4. (a) Band 7 Landsat image of the Nansen Ice Sheet on February 8, 1973. Solar elevation was 18° and azimuth was 79°. (b) Interpretation (bottom). Sastrugi directions (short arrows) have been adapted from Wright and Priestley [1922].

between the Nansen Ice Sheet and the Priestley Glacier takes place in two distinct steps. These two boundaries form acute angled lines emanating from the northern part of the Reeves Glacier. The enclosed triangular area may be present because the northern boundary of the katabatic jet fluctuates in position. However, it cannot be ascertained from available data whether this is a persistent feature. On site observations from three austral summers do indicate that katabatic winds from the Reeves Glacier have a pronounced northern edge. To paraphrase D. N. B. Skinner (personal communications, 1981, 1983), within a direct line of the Reeves Glacier, the wind blows almost continuously with a speed of at least 5 m s^{-1}; north of this line there is little wind but a lot of snow accumulation.

The existence of strong katabatic winds along the western shore of Terra Nova Bay appears to be a notable anomaly. Visual observations by one coauthor (DK) indicated that the Nansen Ice Sheet is a nearly flat plain 10 to 20 m above sea level. Thus strong katabatic winds are present some 25 km beyond the end of the main slope where they lose almost all of their dynamic support. Observations from other parts of east Antarctica show that katabatic winds dissipate over the pack ice within a few km of the shore [Tauber, 1960; Weller, 1969; Schwerdtfeger, 1970], provided the flow is not assisted by a favorable synoptic pressure gradient.

Ball [1956, 1957] analyzed the dissipation of strong katabatic winds by drawing an analogy with the behavior of water flowing down an incline and merging with a quiescent body of water such as a lake. He explained, in terms of hydraulic jumps, the characteristics of abrupt lulls in the nearly continuous katabatic storm experienced by Douglas Mawson's party at Cape Denison. In qualitative agreement with the often-observed rapid dissipation of katabatic winds, the predicted transition between strong downslope winds and comparatively calm conditions over flat terrain takes place near the bottom of the slope. As pointed out by Ball [1956], shooting flow (which applies to drainage down the Reeves Glacier as the Froude number significantly exceeds 1) can be maintained far beyond the slope break if the cold air depth there is less than the critical depth (~700 m). The probable presence of frequent barrier winds [Schwerdtfeger, 1979b; Kurtz and Bromwich, 1983], in which cold air masses pile up against the Transantarctic Mountains and overlie the katabatic outflow, strongly suggests that Ball's model does not apply. However, inadequate information on the boundary layer structure over the western Ross Sea precludes a definitive resolution of this question. The case studies carried out by Gosink [1982] and Mahrt and Larsen [1982] demonstrate that comparatively strong synoptic pressure gradients, a factor neglected by Ball [1956], markedly influence katabatic flows near the foot of the slope. Probably the most important shortcoming in applying Ball's model to average conditions is that it implicitly assumes a negligible density difference between continental air and boundary layer air out to sea. Reynolds [1980] has shown that, over the Gulf of Alaska, the buoyancy contrast between katabatic and marine boundary layers can be eliminated within 20-30 km of the coast by the large upward flux of sensible heat. In general, this buoyancy dissipation rate must be considerably reduced over ice-covered Antarctic coastal waters.

For situations where a significant density difference exists between the air masses, there are two mechanisms by which air on the plateau can descend to sea level [Schwerdtfeger, 1970]: as either foehns or boras. Foehn winds are usually much warmer than the air at the foot of the slope and would require a suitable pressure gradient to remove denser air from the coast of Terra Nova Bay. As demonstrated for 1979 [Kurtz and Bromwich, 1983] and as suggested by climatic maps [Taljaard et al., 1969], westerly geostrophic winds occur infrequently in this area. Thus, the persistent strong winds at Inexpressible Island cannot be foehns. With bora-type conditions, air blowing downslope is denser than low level air. This airflow does not require favorable geostrophic conditions and better explains katabatic winds in Terra Nova Bay.

There are at least three factors that could contribute to the presence of strong bora winds blowing across the eastern edge of the Nansen Ice Sheet. First, the horizontal density difference between the cold air coming downslope and the air over the shelf may be atypically large. This could be due to a markedly colder (in terms of potential temperature) source region or to atypical energy exchanges between the air and the ice preceding its descent to sea level [e.g., Radok, 1973]. Second, for those locations where rapidly dissipating katabatic winds have been observed, the ice slope ends abruptly at the coast and the cold, dry air moves out over ice covered ocean. Here, there is relatively minor air mass modification by energy exchange with the underlying ice sheet, and the horizontal density difference persists undiminished. Third, airflow down the Reeves Glacier valley may stabilize the katabatic outflow [e.g., Scorer, 1952].

A plausible estimate can be constructed for the projected horizontal density difference between the boundary layer air at the coast and that out to sea. Based upon July climatological data given by Taljaard et al. [1969], a nearly isothermal boundary layer is assumed for the point 75°S, 165°E. This is consistent with the isopleths of surface inversion strength given by Phillpot and Zillman [1970]. An average cold air depth of 450 m at the edge of the Nansen Ice Sheet is adopted for reasons discussed in the next section and is compatible with the few winter values available from the writings of Scott's northern party. The layer-average potential temperature of this katabatic current is specified to be 2 K colder than the low level air out to sea; potential temperature differences on isobaric surfaces are directly proportional to density differences. As a result the katabatic layer in the vicinity of Inexpressible Island is about 4 K colder than the air aloft. Corresponding values measured at Mirny (for the layer between the surface and the warmest temperature at 400 m) during "pure" katabatic flow was 3 K [Rusin, 1964]. The resulting July surface temperature (~-30°C) at the edge of the Nansen Ice Sheet is somewhat colder than July averages (-26°C) for McMurdo and Hallett stations [Schwerdtfeger, 1970].

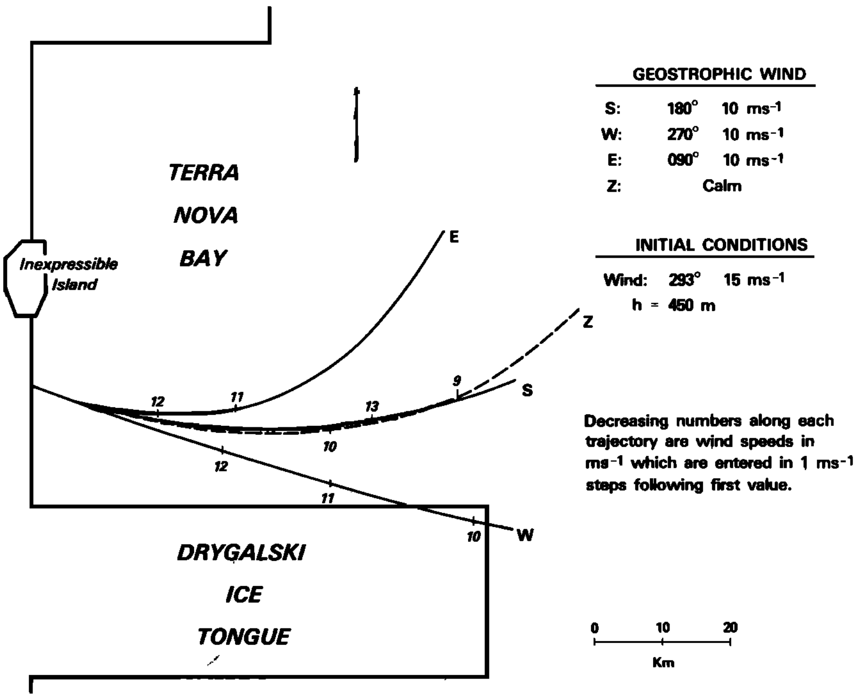

Fig. 5. Approximate trajectories of the dissipating katabatic jet.

Maintenance of the Open Water

Trajectory calculations for the dissipating katabatic airstream provide a basis for scaling the rate at which the horizontal density contrast between the jet and the surrounding air decays. Details of the trajectory modeling are given in the appendix. The layer-average wind speed and the surface value are assumed to be equal. Probably the most serious simplification is the neglect of large fluxes of sensible and latent heat into the cold air (estimated below) once it blows across the open water of western Terra Nova Bay. Thus, simulated trajectories over the western bay are of qualitative value only. However, derived air motions are probably representative of flow across the Nansen Ice Sheet.

Figure 5 shows the derived trajectories obtained for a set of geostrophic winds encompassing most of those encountered during the winter of 1979 [Kurtz and Bromwich, 1983]. The initial condition is the representative surface wind at the eastern edge of the Nansen Ice Sheet which was inferred from Priestley's 1912 observations [Bromwich and Kurtz, 1982]. For all conditions there is relatively little difference between the trajectories within 20-30 km of the coast, a typical polynya width. It should be noted that geostrophic winds from the quadrant between east and south are most favorable for the maintenance of the jet. Such geostrophic winds were observed 90% of the time during the winter of 1979 [Kurtz and Bromwich, 1983]; the majority of these were probably barrier winds. Upstream extrapolation of the trajectories to the mouth of the Reeves Glacier gives a layer-average wind of 23 m s^{-1} and a 16° anticyclonic deflection during airflow across the Nansen Ice Sheet. A wind direction change of this order can also be inferred from Figure 4. A cold air depth of 450 m at Inexpressible Island yields a 300 m deep katabatic layer at the foot of the Reeves Glacier slope; the latter value is typical for Cape Denison [Ball, 1957]. It is interesting to note that 23 m s^{-1} is close to the winter-mean surface wind speed recorded at Cape Denison [Parish, 1981].

The trajectories in Figure 5 also suggest that the energy exchanges across the sea-air interface can be approximated by considering a surface wind speed of 13 m s^{-1}. The resulting surface energy balance calculations are presented in Table 1. Allowance has been made for air mass modification during airflow across the polynya by using climatological, marine air temperatures and dew points. Suppose that the katabatic layer over the polynya has a mean speed of 13 m s^{-1}, an average depth of 500 m (from trajectory modeling), and a horizontal layer-average buoyancy deficit of 2 K. With a sensible heat flux of 630 W m^{-2} into the layer, there is no horizontal density contrast once the air has traversed 29 km of open water. With a combined sensible and latent heat flux of 820 W m^{-2} (this implies complete water vapor condensation), the density contrast has disappeared after the air has travelled 22 km. These results together with those of Reynolds [1980] suggest that the katabatic jet could maintain its identity as it crosses the typical expanse of open water in western Terra Nova Bay. However, the uncertainty attached to this calculation indicates that further research is needed to ascertain whether grease ice in the polynya (discussed in the next paragraph) significantly affects the spatially-averaged surface energy fluxes. Under the present assumptions, conclusions drawn from Table 1 agree with the trajectory analysis in Figure 5.

It is instructive to consider how the open water of the polynya can supply the necessary

TABLE 1. Estimated Surface Energy Balance (in W m^{-2}) of Open Water in Western Terra Nova Bay for a Wind of 13 m s^{-1}.

Month	SW	LW	R	H	LE	G
March	63	-37	26	379	176	-529
April	15	-40	-25	403	179	-607
May	0	-48	-47	575	192	-814
June	0	-46	-46	601	188	-835
July	0	-50	-50	574	192	-816
August	6	-48	-42	625	194	-861
September	36	-42	-6	615	193	-814
October	117	-41	76	391	173	-488

The sensible and latent heat fluxes are directly proportional to wind speed.

SW = absorbed shortwave radiation (equation from Sellers [1965]). LW = net longwave absorption (equation used by Reed [1976] and Gordon [1981]). R = net radiation = SW+LW. H = upward flux of sensible heat (bulk equation with transfer coefficient = 1.25×10^{-3} [Liu et al., 1979]). LE = upward latent heat flux (bulk equation with transfer coefficient = 1.33×10^{-3} [Liu et al., 1979]). G = energy flux gained by the water column as a result of energy exchanges across the air-water interface = R-H-LE.

Sign convention [Munn, 1966]: An energy gain by radiation is positive. Turbulent transports (H and LE) are positive when the surface loses energy. Storage changes and heat advection in the water column (G) which direct energy to the surface are negative.

Input Data: Monthly clear sky solar flux interpolated to 75°S [Rusin, 1964]. Monthly water surface albedo at 70° latitude [Budyko, 1964]. Monthly total cloud cover for McMurdo Station [Schwerdtfeger, 1970]. Water surface temperature of -1.9°C. Saturated vapor pressure at the water surface. Monthly surface air temperature, dew point and pressure at 75°S, 165°E [Taljaard et al., 1969].

energy (term G in Table 1). The winter water column in Terra Nova Bay probably consists solely of Ross Sea Shelf Water. This water is essentially isothermal, with density increasing at depth due to increasing salinity; temperature and salinity ranges are -1.8° to -2.0°C and 34.75 to 35.00‰, respectively [Jacobs et al., 1970]. With the water column being almost at the surface freezing point of sea water, the only significant energy source is latent heat release associated with ice formation (i.e., the horizontal and vertical advection of heat within the water are likely to be small). The highly unstable lapse rate in the lowest layers of the katabatic jet promotes efficient momentum transfer from the air to the water for wave and current generation. Frazil ice crystal suspensions, which form first in wind-agitated seas, are probably herded by Langmuir circulations to form long rows of grease ice parallel to the wind [Martin and Kauffman, 1981; Martin, 1981]. Streaks of ice parallel to the likely katabatic wind direction have been observed in the Terra Nova Bay polynya during late austral spring (D. H. Elliot, personal communication, 1982). Downwind surface drift in these ice convergence zones clears the upwind water surface for new ice formation. To supply an average of 720 W m^{-2} ice production equivalent to a 22 cm solid layer is needed each day. This corresponds to a daily ice transport of 5500 m^3 per m width at the eastern edge of a 25 km expanse of open water. With low ambient air temperatures, it follows that cessation of the katabatic wind leads to rapid freezing of the sea surface, as was observed in 1912 [Bromwich and Kurtz, 1982]. Ice formation and associated brine release will induce haline convection [Foster, 1968; Lewis and Walker, 1970]. Countryman [1970] points out that such downward mixing may occur to great depths due to the absence of a sharp pycnocline; in parts of Terra Nova Bay the water depth exceeds 1000 m (Figure 1).

Blocking Action of the Drygalski Ice Tongue

The wind conditions discussed heretofore prevent sea ice from consolidating in Terra Nova Bay. But, the existence of the polynya also appears to be fundamentally controlled by the Drygalski Ice Tongue because it prevents existing sea ice from drifting into the bay from the south. This blocking effect is reasonable in view of probable northward ice drift in the western Ross Sea. No direct measurements of winter sea ice motions are available in this vicinity, but the track of the Aurora, which drifted helplessly in the western Ross Sea under the influence of winds and surface currents, indicates net northward pack ice drift with a mean speed of 7.6 cm s^{-1} between May 6 and August 2, 1915 [Wordie, 1921]. Numerical simulation of Ross Sea ice drift for the month of April [Baranov et al., 1977] indicates that sea ice in the western Ross Sea drifts to the north with a small component toward the coast. Comparable atmospheric circulation during all winter months in the Ross Sea allows this drift pattern to be extrapolated for the entire winter. Following the reasoning of Schwerdtfeger [1979a], barrier winds should enhance equatorward ice advection within 100-200 km of the coast. Northward sea ice motion may not occur there during summer. Ahlnaes and Jayaweera [1982] observed that ice flows near the Drygalski Ice Tongue drifted slowly to the southeast between December 3, 1981, and January 3, 1982.

Observations of sea ice conditions in the vicinity of the Drygalski Ice Tongue also support this. In 1909, David and his party [David and Priestley, 1914] observed numerous pressure ridges and extensive areas of old ice along the southern border of the ice tongue. One of the present coauthors (DK) also observed numerous pressure ridges there during the austral summer of 1980; and multi-year ice is visible there in Figure 4a. These features are consistent with the compressive sea ice stress field predicted by Baranov et al. [1977]. Early Antarctic explorers appear to have believed that sea ice was deflected by the ice tongue. Priestley [1974] wrote that, in 1912, northward drifting sea ice was known to accumulate south of the Drygalski Ice Tongue, and to move slowly eastward until it passed the eastern end of the barrier. There, the ice resumed its northward drift. Priestley's assertions imply that these drift patterns were general knowledge at that time, but no earlier reference is given.

Maximum polynya width closely correlated with the length of the Drygalski Ice Tongue during 1979. The eastern boundary of the polynya during its maximum extent was always marked by a well-defined band of cold sea ice which separated polynya waters from the Ross Sea pack ice. This band extended northeastward from the eastern end of the ice tongue. A remnant of the band persisted even when less open water was present [e.g., Kurtz and Bromwich, 1983, Figure 2a]. The persistence of this zone, regardless of polynya behavior, suggests that it was influenced by factors in addition to those controlling the area of open water. One possibility is that the band reflects compressional stresses at the boundary between eastward drifting sea ice from the polynya and the northward drifting Ross Sea pack ice.

The ice tongue blocking effect explains in part why similar polynyas do not form along other Antarctic coasts where strong offshore winds blow (e.g., Cape Denison). Despite favorable wind conditions, those coastlines are not similarly protected from large-scale movements of the pack ice. Wintertime open water originating due to wind forcing is local, nearshore, and temporary, occurring only when favorable synoptic conditions prevent it from being overwhelmed by regional sea ice advection [Knapp, 1972]. Unprotected coasts with strong average offshore winds may, however, be sites of early ice breakup or late ice formation.

Conclusions

The Terra Nova Bay polynya is formed and maintained through the action of katabatic winds which drain into the bay through the Reeves Glacier valley. Open water is shielded from northward drifting pack ice by the Drygalski Ice Tongue, and polynya width is governed by the length of this barrier.

Wintertime katabatic winds which collect over a large area of the east Antarctic ice sheet flow in topographically controlled drainage patterns. As long as the present ice sheet topography exists, and a wintertime surface temperature inversion forms, katabatic drainage patterns will remain the same. Under the present circumstances, strong, persistent winds blow down the Reeves Glacier almost continually during winter. Those winds do not undergo turbulent "jumps" at the base of the slope but, rather, blow out over the bay for great distances. Simulations indicate that the zone of wind action, even under unfavorable synoptic conditions, is consistent with observed polynya dimensions. When conditions are favorable, these winds may extend beyond the end of the Drygalski Ice Tongue. Winds passing over the polynya remove heat from the sea surface, probably forming sea ice which is advected to the east; the water column contains at least the latent heat of freezing. Great depths and the absence of a pycnocline enable waters to be brought to the surface through wind-driven upwelling or haline convection.

The Drygalski Ice Tongue plays a less dynamic, though equally important role in maintaining the polynya through its blocking action. The absence of similar sea ice blocking agents along other coasts experiencing strong katabatic outflow may well explain why persistent polynyas do not form there.

The model is strongly supported by existing information, but in situ observations are needed to confirm and expand it. During the 1983/1984 austral summer, an automatic weather station will be installed on the southern tip of Inexpressible Island and a current mooring will be deployed in Terra Nova Bay (the latter is part of the 1984 Ross Sea Heat Flux Experiment) to obtain the necessary data. Simultaneous monitoring of the forcing functions, polynya fluctuations, and water column behavior will contribute to a greater understanding of the interactions between the polynya and the atmospheric and oceanic circulation in the western Ross Sea.

Appendix: Trajectory Modeling

Trajectories [Gordon and Taylor, 1975; Parish, 1977; Rao et al., 1978] have been derived by solving the horizontal momentum and mass conservation equations for a stationary ($\partial/\partial t = 0$) katabatic layer of constant density.

$$\frac{du}{dt} = f(v - v_g) - KVu/h \qquad (1)$$

$$\frac{dv}{dt} = -f(u - u_g) - KVv/h \qquad (2)$$

$$\frac{d}{dt}(hV) = 0 \qquad (3)$$

u and v are the zonal (x) and meridional (y) wind components which are assumed to be independent of height; u_g and v_g are the corresponding geostrophic values; f is the invariant Coriolis parameter; $V = (u^2 + v^2)^{1/2}$; h is the layer depth; K, the drag coefficient, is set equal to 5×10^{-3} [Ball, 1957; Bromwich, 1976]. Apart from the addition of acceleration terms and the deletion of the downslope buoyancy force, the present approach is the same as that used by Ball [1960]. With straightforward manipulation (1) and (2) transform into two simultaneous ordinary differential equations in u and v. Solution in the Lagrangian sense [e.g., Rao et al., 1978] was carried out by marching in time (along the trajectory) with the Runge-Kutta algorithm.

Parish [1977] has carried out similar trajectory calculations to examine the behavior of barrier winds from the east side of the Antarctic Peninsula once they outrun the dynamic support of the mountains. Both that situation and the present one are very similar, and constitute examples of pseudo-inertial flow within the atmosphere [Parish, 1983]; the main differences concern the depth and width of cold air.

A number of important, simplifying assumptions have been invoked to obtain the solutions. First, most of the effects of air entrainment into the katabatic layer have been neglected. The bulk drag terms in (1) and (2) parameterize the combined influence of surface friction and entrainment drag. The former was calculated using the ocean-surface drag coefficient adopted by Overland et al. [1983], and the latter was estimated from the entrainment expression given by Reynolds [1980] for strong convection. On average for the cases considered in Figure 5, the sum of these two terms approximates the impact of the adopted parameterization with $K = 5 \times 10^{-3}$. In neglecting mass entrainment (3) underestimates the katabatic layer growth during transit across the polynya by about 40%. The incorporation of potentially warmer air increases the heating rate of the cold katabatic air mass by about 30%. These partly offsetting factors will be included in later refinements to the analysis. Second, the instability phenomena (hydraulic jumps) discussed by Ball [1956] do not occur with the present formulation of the problem. Third, the oppositedly directed, local barotropic and baroclinic modifications to the pressure gradient force have been omitted from the equations of motion (compare Overland et al. [1983]). The former is due to the increasing downstream depth of the katabatic jet and decelerates the flow. The baroclinic term reflects the heating of the cold air by the large upward surface energy fluxes over the polynya. Scale analysis demonstrated that these terms almost completely cancel.

Observational studies [Rusin, 1964; Bromwich, 1976] have demonstrated that, for Antarctic katabatic winds, the speed at 10 m and the layer average value are fairly similar. For simplicity and consistency equality of these values is assumed here.

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