Surface airflow around Windless Bight, Ross Island, Antarctica*

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(Received 13 February 1986; revised 10 December 1987)

SUMMARY

Ross Island, Antarctica is located along the Transantarctic Mountains and is subjected to a strong southerly mountain-parallel wind regime. Despite this, Windless Bight on the island's southern coast is a region of anomalous calm. The atmospheric boundary layer dynamics that gives rise to this phenomenon is analyzed both theoretically and observationally. It is the strong static stability of the boundary layer air encountering the high steep topography of Ross Island, that causes a stagnation zone resulting in the calm conditions of Windless Bight.

Direct and proxy observations of boundary layer winds provide a detailed description of airflow around the island. The anomalous (with respect to the synoptic pressure field) but persistent north-easterly winds at Scott Base are due to the deflection of highly stable, low-level air around Hut Point Peninsula. By contrast, the less frequent, strong southerly winds which override the peninsula are associated with the influx of warm maritime air from cyclonic systems to the east. It is inferred that flow of air around the terrain of Ross Island gives rise to locally strong winds; these are responsible for the ice breakout and polynya occurrences in McMurdo Sound.

The surface airflow past Ross Island can be modelled by a two-dimensional, steady, frictionless, irrotational, incompressible flow past an obstacle, with a shape based on an island height contour. The flow is assumed to separate from the eastern and western sides of the island and form a wake downstream. A solution for this flow is presented, based on potential theory for streaming motion past an obstacle. The streamline and isobar patterns clearly depict the stagnation region in Windless Bight. For approaching winds of 20 m s⁻¹ the local pressure field is perturbed by several millibars.

1. INTRODUCTION

(a) History and description

One location that has figured prominently in the history of antarctic exploration is Ross Island. It is located near the western edge of the Ross Sea at the northern boundary of the permanent Ross Ice Shelf. The very steep terrain of Ross Island is dominated by three peaks, Mts Erebus, Terra Nova and Terror. The terrain rises steeply to form a mountain chain over 1500 m high oriented nearly east–west, with the island's southern boundary forming a bight (see Fig. 1). Early in this century the region was explored by the British expeditions of Robert F. Scott (1901–1904 and 1910–1913) and Ernest H. Shackleton (1907–1909 and 1914–1917) who conducted numerous field investigations. At that time it was observed that the southern bight of Ross Island was a region of anomalous calm (Simpson 1919). Here the sledging parties encountered deep soft snow that made travel extremely difficult. This region is called Windless Bight and is known to be one of the calmest places on earth. For this reason it is a preferred location for measuring atmospheric infrasound waves generated by a variety of geophysical phenomena in the polar regions (Wilson 1977; Wilson and Collier 1982).

* Dedicated to the late Professor Werner Schwerdtfeger, for his pioneering work in the dynamic meteorology of the antarctic.
We intend to explain the existence of the anomalously calm Windless Bight as a consequence of the boundary layer airflow past Ross Island. This is shown by the classic diagram due to Simpson (1919 Fig. 37 p. 110, reproduced here as Fig. 2(a)) based on wind observations during blizzards and the orientations of wind-induced ridges on the snow surface (sastrugi). Since the late 1970s, automatic weather stations (AWS) have been deployed in the vicinity of Ross Island reporting surface temperature, pressure, and wind observations (Stearns and Savage 1981; Renard and Thomson 1982). Resultant winds for February through May 1984 from the AWS (Savage et al. 1985) and from Scott Base have been added to Fig. 2(a) for comparison. It appears that Simpson’s streamlines provide an approximate description of both blizzard and time-averaged surface airflow around Ross Island. Similar conclusions were obtained by Stearns (1982), Sinclair (1982) and Schwerdtfeger (1984 p. 97).

Some discrepancies between the streamlines and the resultant winds can be seen. Winds from sites near Hut Point and close to sea level (AWS18 and Scott Base) show that the time-averaged low-level airflow goes around Hut Point Peninsula. AWS09 is at an elevation of 200m and monitors higher level air motions which closely follow the streamline pattern. The orographic influence of Hut Point Peninsula is analysed in greater detail in section 3(a). The lighter resultant wind speed at AWS11 probably arises because it is situated in the main airflow deceleration region. AWS06 (at Marble Point) is subject
to both the southerly regime and episodes of katabatic drainage through the adjacent mountains (Savage and Stearns 1985). The persistent west-south-west winds at AWS 05 indicate a frequent upstream blocking situation south of Minna Bluff, where the boundary layer air turns eastward and goes around this 600 to 1000 m-high obstacle.

(b) Synoptic climatology

Before the boundary layer airflow around Ross Island is considered in detail, the topographic and climatological circumstances which give rise to persistent mountain-parallel winds approaching Ross Island from the south will be outlined. The Ross Ice Shelf extend from 78°S to 85°S in latitude and from 165°E to 155°W in longitude (Fig. 3), with the ice surface at a height of about 60 m above sea level. Fifty kilometres to the west of Ross Island lie the Transantarctic Mountains, which form the western
boundary of the Ross Ice Shelf and Ross Sea. This mountain range rises abruptly to elevations exceeding 2500 m, and extends about 700 km to the north and to the south of Ross Island. Near Ross Island the range has a meridional orientation. West of the Transantarctic Mountains lies the elevated East Antarctic ice sheet.

From a global climatological standpoint there is sinking air over the polar regions, so that the surface pressure increases polewards (compare Taljaard et al. 1969), giving rise to the circumpolar easterlies. This is a relatively shallow atmospheric phenomenon, and the easterlies weaken above the boundary layer and become westerlies aloft. Subsidence into the boundary layer inversion gives rise to the climatological feature of south-easterly geostrophic winds over the ice shelf (Jenne et al. 1971).

These shallow winds over the Ross Ice Shelf have an easterly component, often being from the south-east and sometimes from the east. A strong inversion develops over
the ice shelf, especially in winter (Schwerdtfeger 1970 Fig. 12; 1984, p. 83; Savage and Stearns 1985), so that the air has a high static stability. As the surface air impinges on the high wall of the Transantarctic Mountains it cannot flow over because of the terrain height and the strong static stability of the air. The blocking condition can be described by an internal Froude number of the upwind flow (Manins and Sawford 1982). The depth of the boundary layer increases from east to west as the cold stable air is dammed up against the western boundary and deflected northwards. The dynamics of this situation is described in detail by Schwerdtfeger (1975) as it applies to the cold arctic air sloping up against the north side of the Brooks Range in Alaska. If this situation persists for some time, an approximate balance develops between the pressure gradient force due to the varying depth of cold air piled up against the mountains and the Coriolis force. This geostrophy gives rise to mountain-parallel jet streams below crest height, so-called barrier winds (Parish 1982). Schwerdtfeger (1979a, b) summarizes observational evidence supporting this line of reasoning. In particular there is the persistent climatological feature of the barrier winds at Ross Island. Savage and Stearns (1985) show that the mean sea
level pressure distribution around Ross Island is consistent with a climatological barrier wind regime.

Barrier winds are a ubiquitous phenomenon. The general conditions under which low-level stable air is blocked by extended steep mountains have been studied by Pierrehumbert and Wyman (1985). The resulting mountain-parallel winds are observed at numerous locations outside the polar regions such as to the north of the Alps (Pierrehumbert and Wyman 1985), to the east of the Rockies (Dunn 1987), and along the Sierra Nevada Mountains in the western United States (Parish 1982).

Barrier winds can also be generated or modified by the low-level blocking of synoptic-scale cyclonic circulations (Schwerdtfeger 1979b, 1984 p. 96), especially in the Weddell Sea along the Antarctic Peninsula. Figure 2(b) illustrates such a situation over the northwestern Ross Ice Shelf. The sea level pressure analysis is based upon AWS observations (Savage et al. 1985). A visible DMSP (Defense Meteorological Satellite Program) satellite picture taken four hours later confirms the placement of the two cyclonic centres as well as the easterly geostrophic flow across the ice shelf. This westward moving air is blocked and deflected to the north by the Transantarctic Mountains and then channelled around Ross Island. This last effect is shown by the deceleration of the airflow from 15 m s⁻¹ near White Island (AWS 11) to 1 m s⁻¹ in Windless Bight (AWS 18); over the same 30 km sea level pressure increases by 3 mb. This event lasted for two and a half days with the conditions described by Fig. 2(b) being stationary for about 9 hours.

A major source for cold air over the northern Ross Ice Shelf is katabatic drainage from Byrd Glacier. Parish and Bromwich (1987) modelled the winter surface windfield over the antarctic ice sheets and found that air from a huge section of the continent converges into this glacier valley. Such confluences in the boundary layer windfield are believed to be responsible for downwind regions of greatly intensified cold air transport from the ice sheet (Parish 1984). Winter thermal infrared (TIR) satellite images persistently show a prominent warm signature (a thermal 'plume') streaming onto the shelf from Byrd Glacier (Kurtz and Bromwich 1985; D'Aguanno 1986). Less prominent plumes nearly always emerge from Skelton and Mulock Glaciers. Such signatures are assumed to be generated primarily by snow surface warming in association with turbulent, adiabatically-warmed katabatic airstreams which also carry substantial loads of drift snow (Kurtz and Bromwich 1985). Even though these airflows are probably negatively buoyant in relation to the boundary layer as a whole, they appear warm when compared with snow surface emissions from adjacent areas which are relatively calm and cold. A TIR image presented by Scorer (1986 Fig. 16.2i) shows a similar katabatic signature coming from Kangerdlugssaq Fjord in Greenland, although this is stated to be a kind of foehn wind. Thus satellite images may confirm that large quantities of cold katabatic air are transported onto the Ross Ice Shelf from Byrd Glacier.

It is probable that this katabatic airstream plays a major role in the southerly wind regime affecting Ross Island. Kurtz and Bromwich (1985 Fig. 3c) give a TIR image which shows that katabatic winds from Byrd Glacier can propagate 400 km across the flat Ross Ice Shelf to reach the Ross Sea just east of Ross Island. Continuing analyses of AWS data should help to illuminate the governing atmospheric dynamics.

The paper is organized as follows. Section 2 reviews existing theoretical and observational studies to demonstrate that low-level air must flow around Ross Island and form a downstream wake. Direct and proxy observations of boundary layer winds near Ross Island are examined in section 3 to characterize the airflow. Section 4 presents a simple model description of the windfield which is based upon potential theory for streaming motion past an obstacle.
AIRFLOW AROUND WINDLESS BIGHT

2. Flow around isolated topography

(a) Theoretical investigations

Brighton (1978), Baines (1979), Hunt and Snyder (1980), and Smith (1980) show that when air flows past an isolated mountain there can be different flow regimes for different regions. The upper-level flow goes over the mountain top and may form lee waves. Recently, Lamb and Britter (1984) and Snyder et al. (1985) have shown that in the presence of strong stratification the flow near the surface lacks the kinetic energy to rise over an isolated obstacle and must flow around it. This is a nearly horizontal flow regime that separates from the sides of the mountain so that a wake is formed downstream (compare Walter and Overland 1982). In particular, Brighton (1978 p. 296 Fig. 3) reports that this flow regime near the surface can be described by two-dimensional, inviscid potential theory. Smith (1980 p. 355 Fig. 4) shows that after the surface streamline separates from the mountain, it has a permanent outward deflection. This is the bounding streamline for an infinite wake that does not close behind the mountain. There may subsequently be sinking air from higher levels into the wake region. Similar conclusions were reached by Arakawa (1973).

These features can be related to observations of airflow near Ross Island. Pileus or cap clouds indicating rising air near the top of Mt Erebus, and upper-level lee wave clouds have been observed since the earliest expeditions (Shackleton 1909 pp. 390–396). This is also one of Antarctica’s active volcanoes, and gives off a steam plume that can serve as a tracer of upper-level wave motions. Although the upper-level airflow is observed to go over the island, the stagnation region at Windless Bight must be the result of low-level airflow going around the island. Ross Island is one place where the persistent boundary layer winds and high static stability make this phenomenon a climatological feature.

The geometry of Ross Island is an important consideration in the flow separation. Windless Bight is nearly symmetrical with respect to a north–south axis, which nearly coincides with the direction of the incoming flow (see Fig. 2(a)). Any quantitative study of the flow generation is greatly simplified by assuming complete symmetry of the bight with respect to the direction of approaching airflow. This assumption, in conjunction with near-surface airflow around the obstacle, permits an examination of the phenomenon at Windless Bight by the analytical means of potential theory for flow past a bluff body. Once again Antarctica is a laboratory for meteorology (Lettau 1971).

(b) Case study of topographically controlled winds

The studies of Dickey (1961) and Kozo and Robe (1986) of the effects of Alaska’s Brooks Range on strong low-level winds are relevant to our investigation. This range runs nearly east–west with a semi-circular knob extending to within a few tens of kilometres of the Beaufort Sea coastline. Dickey noted that arctic air in the lowest layers has a high static stability and therefore tends to flow horizontally around the ends of a topographic barrier rather than over it. This gives rise to orographically enhanced winds at the end of a topographic barrier. The cases investigated were periods of strong easterlies or westerlies blowing parallel to the mountains and then encountering the protruding knob of the range. The 2000-foot height contour of this knob was represented by the arc of a semi-circle. Dickey showed that for reasonable assumptions (discussed later) the boundary layer windflow could be modelled by the dynamics of two-dimensional, irrotational, incompressible flow past a circular cylinder. The textbook solution (see Lamb 1945 p. 77) has a stagnation zone and region of enhanced flow past the obstacle. This resulting flow pattern was able to explain the anomalous strong winds.
observed at Barter Island, just off the coast, as a result of the orographically modified flow past the Brooks Range knob. Dickey suggested that wind anomalies at many places, especially in polar regions where the surface air is highly stable, might be explained by the flow of air around, as well as over, arbitrarily shaped barriers, and could be investigated theoretically or in laboratory wind tunnels.

(c) Satellite observations of wakes

Analyses of flow separation at other islands will be discussed, with a view toward showing that a boundary layer wake downstream of Ross Island is probable. These studies are especially valuable due to the relative scarcity of data from the antarctic. In particular Lyons and Fujita (1968) investigated mesoscale motions in oceanic stratus near the Aleutian Islands. They found that islands that penetrated well above the inversion level had downstream wakes and concluded that the boundary layer flow past small islands is essentially horizontal with patterns resembling the flow past extended elliptic cylinders in the laboratory. They also found direct evidence of flow stagnation in the form of crescent-shaped brighter cloud regions extending several kilometres upstream of the islands, attributing these to deeper layers of stratus with higher reflectance as the air was piled up against the mountains. There was also evidence for downward transport of drier air resulting in clear wakes behind the islands. Satellite photographs clearly showed some cases of vortex shedding and von Kármán vortex streets in the downstream wakes. Other reports of wakes in the lee of islands beneath a boundary layer inversion, as evidenced by satellite photographs of von Kármán vortex street cloud patterns, have been reviewed by Chopra (1973) and Gjevik (1980); Heinemann (1986) gives antarctic examples. These observational studies indicate that boundary layer flow separation and wake formation can occur for islands at low, middle, and high latitudes, and concur with the results of theoretical analyses.

3. Observations of windflow around Ross Island

There are two stations on the south side of Ross Island (Fig. 1) that are manned continuously. The American McMurdo Station has produced an unbroken record of surface and upper air observations since 1955. The New Zealand station Scott Base is located 3 km away and collects surface data.

(a) Boundary layer climatology

In order to characterize the structure of the lower atmosphere, the complete rawinsonde record (00 GMT and 12 GMT when available) from McMurdo Station for 1979 was analysed in some detail. This record, in common with those from many other antarctic coastal sites (Bromwich 1978), is biased in favour of lighter surface winds. For release time speeds of 10 to 14 m s⁻¹, 27% of scheduled 00 GMT soundings were unsuccessful, and both soundings were missed when winds were stronger than 15 m s⁻¹.

The stratification of the airstream approaching Ross Island from the south can be roughly estimated from McMurdo radiosonde data (Schwerdtfeger 1984 p. 83). Table 1 gives the strength, depth and occurrence frequency of the surface and first aloft inversions at McMurdo. In the warmer months when the sun is almost continuously above the horizon (November to February), surface inversions are rarely observed. During the predominantly nocturnal months of March to October, surface inversions were recorded during 56% of ascents, had an average strength of 3.1 degC and were 250 m deep. These results are similar to those obtained by Phillipot and Zillman (1970) from analysis of 1967 data. Both analyses yield an average surface inversion strength for all March to October
<table>
<thead>
<tr>
<th>Month</th>
<th>Strength (deg C)</th>
<th>Depth (m)</th>
<th>n</th>
<th>% of all soundings</th>
<th>Strength (deg C)</th>
<th>Altitude of base (m)</th>
<th>Depth (m)</th>
<th>n'</th>
<th>% of all soundings</th>
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<td>1</td>
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<td>180</td>
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<td>420</td>
<td>160</td>
<td>29</td>
<td>56</td>
</tr>
</tbody>
</table>

n and n' identify the number of soundings respectively with a surface inversion and with at least one aloft inversion below 2 km altitude.

soundings of 1.7 degC; this quantity is calculated by setting the strength to zero when temperatures near the surface are constant or decrease with height. Because temperatures to the south and east are markedly lower (Savage and Stearns 1985) surface air approaching Ross Island is substantially more stratified than indicated by Table 1. Scott Base for example, only 3 km to the east of McMurdo Station but on the ice shelf side of Hut Point Peninsula, averages 3.1 degC colder than McMurdo from March to October (calculated from climatological surface temperatures for both stations given by Sinclair (1982)). The higher surface temperatures at McMurdo have been attributed to maritime influence and to greater vertical mixing (Sinclair 1982; Savage and Stearns 1985). The majority of soundings contained at least one elevated inversion below 2 km altitude. Between March and October, 70% of ascents showed an inversion, averaging 1.8 degC with average base at 690 m and depth of 160 m.

Evaluation of all three-hourly McMurdo surface weather observations between March and October 1979 showed that as the wind shifts from the predominant east and northeast directions to the infrequent southerly direction, both the average temperature and average wind speed increase. As discussed below, this clockwise rotation is associated with the change from surface airflow around to airflow over Hut Point Peninsula. To isolate the concurrent upper air conditions, soundings with surface directions between north and east (representing flow around) and between south-east and south-west (flow over) were separately averaged and the results are presented in Table 2; cases with weak, possibly locally generated, winds have been excluded. East-north-easterly surface winds are associated with a surface inversion, moderate wind speeds in the lower troposphere, a clockwise rotation of the resultant wind direction with height, and a steady altitudinal decrease of the directional constancy. Wind directions are rather variable above 2 km. Southerly surface winds are accompanied by significantly higher tropospheric temperatures and wind speeds, and high directional constancies up to 5 km. The resultant direction turns clockwise up to 2 km then anticlockwise to 5 km. It appears as though the typical easterly surface airflow at McMurdo is generated by a variety of situations while the infrequent southerlies are associated with an influx of warm, moist air probably in conjunction with the approach of a maritime cyclone.
Schwerdtfeger (1984 pp. 98–99) notes that at Scott Base surface winds from the north-east are most frequent when wind speeds are less than 10 m s\(^{-1}\) but winds with speeds greater than 15 m s\(^{-1}\) are generally from the south. This is the result of the boundary layer airflow from the south-east encountering the topography of the 200 m-high ridgeline of Hut Point Peninsula (Fig. 1) which extends from the slopes of Mt Erebus to McMurdo Station. This is best discussed with reference to an internal Froude number based on the height of the terrain obstacle (Kitabayashi 1981). The square of this Froude number is a ratio of the kinetic energy of the approaching wind to the potential energy required to lift an air parcel from the surface to the height of the obstacle. It is given by

\[
Fr = \frac{U_o (gH \Delta T/\bar{T})^{-1/2}}{T}
\]

where \(U_o\) is the approaching wind speed, \(g = 9.8 \text{ m s}^{-2}\) the acceleration of gravity, \(H = 200 \text{ m}\) the height of the ridgeline, \(\Delta T\) the potential temperature difference between the surface and top of the ridge, and \(\bar{T}\) the vertically-averaged potential temperature over this layer.

Kitabayashi (1977) investigated the upstream blocking of a boundary layer flow by a 200 m-high ridgeline in Japan. From analyses of meteorological and wind tunnel data he concluded that the critical Froude number is 2.3. For higher values the flow is over the ridgeline, while for lower values the flow is blocked up to some height on the ridge. If this partial blocking occurs, the air near the surface flows around the end of the ridgeline, as described by Baines (1979 p. 7836 Fig. 4). Using average values of \(\Delta T\), \(\bar{T}\) and \(U_o\) derived from Table 2 for east-north-east winds (2-9 K, 253 K and 6-8 m s\(^{-1}\)) and for south-east winds (1-2 K, 258 K and 8-3 m s\(^{-1}\)) at McMurdo in Eq. (1) yields Froude numbers of 1.4 and 2.8, respectively. From Kitabayashi’s criterion, it appears that the predominant airflow (from the north-east at Scott Base and east at McMurdo) is associated with blocking of the lowest level air and deflection around Hut Point Peninsula. This conclusion is consistent with the discussion in section 1(a). For the atypical southerly wind events, surface air goes over the peninsula.

<table>
<thead>
<tr>
<th>Altitude (m)</th>
<th>Mean air temperature (°C)</th>
<th>Mean wind speed (m s(^{-1}))</th>
<th>Resultant wind</th>
<th>Mean air temperature (°C)</th>
<th>Mean wind speed (m s(^{-1}))</th>
<th>Resultant wind</th>
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<td>67°/6.8</td>
<td>q</td>
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</table>
(b) Sea-ice observations

Strong wind occurrences are of some significance for the annual ice breakout in McMurdo Sound. During the austral winter the sound is usually frozen as far north as Cape Royds (Fig. 1), but by the end of February there may be open water at McMurdo Station, facilitating resupply by ship. Historically, this feature allowed early expeditions to move into the sound. The ice breakout has been studied by Heine (1963) and Prebble (1968) who concluded that after the sea-ice has been broken up due to thermodynamic processes and heavy sea swells from the open Ross Sea to the north, it is the strong southerly wind squalls that remove the ice from the sound. Significantly, only the eastern side of McMurdo Sound close to Ross Island opens early, while along the western shore the sea-ice may remain year-round. This might indicate the effect of anomalous strong winds resulting from the air being forced around the topography of Ross Island, and causing a flow separation from the side of the island and a wake downstream.

According to Stonehouse (1967), the existence of a penguin breeding colony at more southerly latitudes indicates that open water occurs anomalously in early spring, because these birds need a food supply. Stonehouse indicates that there is a recurring polynya (area of open water surrounded by ice) forming in the austral winter or early spring, close to Ross Island between Cape Royds and Cape Bird. This may be the reason for the penguin colonies along the western side of the island (Fig. 1). There are no penguin colonies on the western shore of McMurdo Sound. Again this might indicate the effect of windflow around Ross Island. The existence of penguin colonies at other places in the Antarctic where there are anomalous high winds and year-round open water has been noted by Bromwich and Kurtz (1982). There are also penguin colonies at Cape Crozier, at the easternmost extent of Ross Island, where we would also expect topographically enhanced winds.

(c) Descriptive evidence of a wake

An outstanding example of topographically controlled winds was observed by Scott's last expedition (Wright and Priestley 1922 pp. 12–13) in the vicinity of Cape Evans on the western side of Ross Island. Between about 10 am and 7 pm on 1 October 1912, a south-easterly blizzard with very strong surface winds and dense blowing snow buffeted the station, probably well in excess of 20 m s\(^{-1}\) although no measurements are given by Simpson (1923). The wind speed then gradually decreased until the wind recorder registered a dead calm around 10 pm. Upon emerging from the hut, the situation sketched in Fig. 4 was noted. Cape Evans was situated in a narrow calm belt with the blizzard still raging only 0.8 km away to the south-west. A procession of small vortices was observed a short distance to the north-east and farther away a moderate north-westerly breeze prevailed.

The above description is offered as evidence for a boundary layer flow separation of the strong south-easterly blizzard winds from the western edge of Ross Island. The presence of easterly geostrophic winds near the island (in conjunction with the probable eastward propagation of a cyclone across the northern Ross Sea (Kidson 1947)) and comparatively high and stable sea level pressures at Cape Evans during 1 October (Simpson 1923) are consistent with a barrier wind event. The lighter winds from the north-west may have resulted from backflow around the island in a relatively calm wake. The small whirlwinds could have been mechanical turbulence produced in the strong horizontal wind shear zone at the edge of the wake.
4. DYNAMICAL MODEL

(a) Basic assumptions

It is desired to model two southerly wind situations: a well developed barrier wind event with an approaching wind speed of $20 \text{ m s}^{-1}$; and a climatological average situation with a wind speed of $5 \text{ m s}^{-1}$. For both cases the flow may be assumed steady and the local time derivatives set to zero. Furthermore, the effects of friction will be neglected. Any coefficient of surface friction for stable airflow over the flat smooth ice shelf must be small, and friction is not the dominant consideration in determining the lateral flow separation from the sides of the mountain.

We neglect Coriolis terms. The length scale $L = 32 \text{ km}$ represents half the east–west width of the island, and the absolute value of the Coriolis parameter, $a$, is $1.4 \times 10^{-4}\text{s}^{-1}$. Then for the well developed barrier wind case, the Rossby number, $U_r(aL)^{-1}$, equals 4.5, indicating that advective acceleration dominates the Coriolis acceleration. However, a complete description of the extended wake would require that these terms be retained (Boyer and Davies 1982). For the case of climatological winds the Rossby number is near unity, but the solution can still be used to describe the flow regime on the smaller scale of the stagnation zone in Windless Bight, if not the complete flow past the island.

As was mentioned in section 2(a), in order for an airflow to rise over an obstacle it encounters, it must possess sufficient kinetic energy to do the work necessary to overcome the stratification. These considerations are encompassed by the Froude number (Eq. (1)) based on the height of the obstacle. Hunt and Snyder (1980), Snyder et al. (1985) and Snyder (1985) show that for strongly stratified flow past three-dimensional obstacles ($0 < Fr < 1$), there will be two layers: the flow in the lower layer is around the obstacle in a nearly horizontal plane, while the flow in the upper layer can be over the obstacle.
Their laboratory experiments with flows of linear density stratification past bell-shaped hills show that the height of the dividing streamline, \( H_d \), separating these layers can be approximated by the formula

\[
H_d / H = 1 - Fr \quad 0 < Fr < 1.
\] (2)

If we assume dynamic similarity of the flow past Ross Island with these laboratory flows, we can calculate the dividing streamline heights for our southerly wind cases. For strong wintertime barrier winds, based on data for March–October 1979 (Table 2), we have the approximate conditions \( U_o = 20 \text{ m s}^{-1}, \Delta T = 8.0 \text{ K} \) and \( T = 262 \text{ K} \). If the average height of Ross Island is taken to be \( H = 1500 \text{ m} \), then from Eqs. (1) and (2) we calculate the values \( Fr = 0.94 \) and \( H_d = 90 \text{ m} \). For the climatological wind case after allowing for the surface temperature gradients measured by Savage and Stearns (1985), the values are \( Fr = 0.16 \) and \( H_d = 1260 \text{ m} \) (\( U_o = 5 \text{ m s}^{-1}, \Delta T = 16.4 \text{ K}, T = 257 \text{ K} \)). Formula (2) for the dividing streamline height was obtained for bell-shaped hills, whereas Ross Island resembles two intersecting conical hills. However, the work of Castro et al. (1983) indicates that the cross-stream width of an obstacle has little effect on the dividing streamline height. As the quantitative theory is only approximate, representative dividing streamline heights were taken to be 400 m for the strong barrier wind case, and 1000 m for the climatological case. The island shape for each case will be taken to be the topographic contour at the dividing streamline height. The nearly horizontal low-level airflow around Ross Island will be modelled as two-dimensional potential flow past a bluff body with downstream wake formation. This certainly describes the flow just above the surface, if not of the entire boundary layer, and it is the surface wind monitored by AWS, sastrugi, and polynyas that we wish to model.

(b) Dynamical equations

The horizontal momentum equations give a steady state balance between the pressure gradient and advective acceleration terms:

\[
U \frac{\partial U}{\partial X} + V \frac{\partial U}{\partial Y} = -\frac{1}{\rho} \frac{\partial P}{\partial X}
\] (3)

\[
U \frac{\partial V}{\partial C} + V \frac{\partial V}{\partial Y} = -\frac{1}{\rho} \frac{\partial P}{\partial Y}.
\] (4)

The continuity equation is that for incompressible flow:

\[
\frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} = 0.
\] (5)

This is the case of irrotational, incompressible flow. If there is uniform flow far upstream of the island (\( U = U_o, V = 0, P = P_o \)), the relative vorticity of an air parcel is zero for all time. A Bernoulli equation for the conservation of energy of an air parcel along a streamline can be evaluated for the undisturbed conditions far upstream of the island, with the result that

\[
\{(P - P_o)/\rho\} = \frac{1}{2}(U_o^2 - (U^2 + V^2)).
\] (6)

A representative air density for cold southerly winds is \( \rho = 1.4 \text{ kg m}^{-3} \). This expression will enable the pressure increase in the bight, due to the flow stagnation in advance of the island, to be determined.
The coordinates and velocities can be nondimensionalized so that \( x = X/L, \) \( y = Y/L, \) \( u = U/U_0, \) \( v = V/U_0. \) The problem has been simplified by assuming that the island is symmetric with respect to the direction of the oncoming flow. Here the uniform southerly flow approaches parallel to the \( x \) axis, as is the usual convention in engineering applications. The equation representing the shape of Ross Island in contact with the oncoming flow between the separation points must be specified. It would not be realistic to account for every wiggle in the 400 m contour, and so we use the equation

\[
F = -1.120y^2 + 2.394y^4 - 1.071y^6 = 0
\]  

(7)

determined by a least-squares fit. The separation point \( y_s = 1.0 \) was chosen so that the flow separates smoothly as the easternmost and westernmost capes are approached.

(c) Flow separation problem

A realistic representation of the boundary layer flow past Ross Island is that of discontinuous or streaming motion past an obstacle, where the flow separates from the side and forms a wake or cavity downstream. The description of the flow regime resulting from a wake has been studied since the earliest works of Helmholtz and Kirchhoff in the last century. In these problems the flow domain is bounded partly by the obstacle and partly by the wake. For simplicity it is assumed that the wake is a stagnant zone where the velocity vanishes, but the pressure is constant across the free streamline forming the boundary of the wake. If the flow separates from the edges of the island, then usually the wake extends to infinity in the mathematical representation of the problem, and the pressure in the wake can be assumed to be equal to that of the undisturbed flow, \( P_0. \) Application of Eq. (6) to motion on the free streamline bounding the wake yields the result that the nondimensional flow speed is constant:

\[
u^2 + v^2 = 1.
\]  

(8)

In only a few cases where the obstacle is composed of straight line segments, can a closed form representation of the transformation equations be found. The case for separation at a finite flat plate oriented perpendicular to the oncoming flow, with a resulting wake behind the plate, is solved in Lamb (1945 pp. 99–102) and would give a first approximation to the surface streamlines near Windless Bight if realistic topography were neglected. The general method for solving flow separation problems for bodies of arbitrary curved shape was developed by Levi-Civita and is summarized in Milne-Thomson (1968 pp. 338–348) and Birkhoff and Zarantonello (1957 pp. 130–152, 215–217). The specific approach presented here was developed by Vanden-Broeck (1983).

First consider the flow separation in the \( z = x + iy \) plane (Fig. 5). The stagnation point \( C \) is at the origin, and the point at infinity for the upstream flow is point \( E. \) From symmetry considerations we have the requirement that

\[
v = 0 \quad 1 > u \geq 0 \quad \text{on} \quad y = 0 \quad -\infty < x \leq 0
\]  

(9)

and so only the upper half plane need be considered. Points \( D \) and \( G \) represent the points at infinity on the wake downstream of the island, and the flow separation is taken to be at points \( A \) and \( B. \) Because the two-dimensional motion is irrotational and incompressible, the velocities can be represented by a streamfunction \( \psi \) and velocity
potential \( \phi \), both of which satisfy Laplace's equation. The complex potential is defined as \( f(x, y) = \phi(x, y) + i\psi(x, y) \) so that the complex velocity is

\[
b \frac{df}{dz} = u - iv
\]  

(10)

where the nondimensional constant \( b \) will be determined by the requirement that \( \phi = 1 \) at the separation points. The symmetric streamline ECBG is taken to be \( \psi = 0 \), and that part BG is the free streamline. A series of conformal transformations is needed. It is necessary to find an intermediate transformation that maps the \( z \) plane to the \( t = r \exp(i\sigma) \) plane, so that the flow regime in the upper half plane is mapped to the interior of a semicircle (Fig. 5). This complex \( t \) plane is subsequently mapped to the complex \( f \) plane by the Levi-Civita transformation

\[
f^{1/2} = (t - t^{-1})/(2i).
\]  

(11)

The major difficulty of the problem is finding the appropriate transformation from the \( z \) plane to the \( t \) plane. This required transformation must satisfy the velocity condition (9) on the symmetric streamline in advance of the island, the Bernoulli equation (8) on the free streamline of the wake, and expression (7) for the shape of that part of the island in contact with the flow. It is useful to introduce another expression for the complex velocity

\[
u - iv = \exp(\tau - i\theta).
\]  

(12)
The appropriate general form of the required transformation is

$$\tau - i\theta = -\ln \left( \frac{1 + t}{1 - t} \right) - \sum_{n=1}^{N} A_n t^{2n-1}$$

(13)

which may be expressed in terms of its real and imaginary parts to give $\tau$ and $\theta$ as functions of $r$ and $\sigma$. On the symmetric streamline in advance of the island $\theta = 0$, $-\infty < \tau < 0$ and so (12) shows that the symmetry condition (9) is satisfied. On the free streamline bounding the wake $\tau = 0$ and so (12) shows that the Bernoulli condition (8) is satisfied. The logarithmic term is the necessary transformation for the problem of uniform flow impinging perpendicularly on a finite flat plate, and the arbitrary constants $A_n$ of the truncated infinite series are necessary to account for more general obstacle boundaries and will be determined presently.

The succession of complicated transformations makes it impossible to obtain an explicit expression for the transformation from the $z$ plane to the $\tau$ plane and obtain $x, y$ directly in terms of $r, \sigma$. However, this can be done by numerical integration:

$$z = \int \frac{dz}{df} df$$

(14)

so that along a streamline $\psi = \psi_0$

$$x = \int \frac{\partial x}{\partial \phi} \left( \frac{\partial \phi}{\partial r} dr + \frac{\partial \phi}{\partial \sigma} d\sigma \right) d\sigma$$

(15)

$$y = \int \frac{\partial y}{\partial \phi} \left( \frac{\partial \phi}{\partial r} dr + \frac{\partial \phi}{\partial \sigma} d\sigma \right) d\sigma$$

(16)

All expressions in the integrands can be evaluated (when the $A_n$ and $b$ are known) for a given value of $\sigma$.

The $N + 1$ coefficients $A_1, \ldots, A_N, A_{N+1} = b$, must be determined numerically from the requirement that the streamline $\psi = 0$ must coincide with the shape of Ross Island in contact with the oncoming flow. In the $\tau$ plane this portion of the flow boundary is represented by the first quadrant arc of the unit circle. The equations governing the transformation of this part of the dividing streamline are a special case ($r = 1$) of the general streamline transformation equations (11)–(15) which is worked out in detail by Vanden-Broeck (1983), using co-location and finite differences. The largest transformation coefficients are given in Table 3.

In order to proceed with the integration along a general streamline, the initial values of $x, y$ that correspond to $r, \sigma$ must be obtained. The equipotential $\phi = 0$ can be integrated numerically (Eq. (14)) from the origin to give the values of $x, y, \psi$ along its path, in an analogous manner to the procedures used for integrating along a streamline. Then the general streamlines can be integrated upstream and downstream from where they cross the zero equipotential. The velocities along a streamline are found from Eq. (12) and the pressure change from Eq. (6). The streamlines, isobars, and actual 400 m contour are shown in Fig. 6.

The climatological case with monthly mean wind speed $U_0 = 5$ m s$^{-1}$ and island shape corresponding to the 1000 m contour was also considered. The same length scale was used, but a different position was selected for the origin, because the stagnation point
TABLE 3. TRANSFORMATION COEFFICIENTS FOR THE FLOW SEPARATION AROUND THE ISLAND BOUNDARY REPRESENTED BY (a) 400 m AND (b) 1000 m ELEVATION CONTOURS

<table>
<thead>
<tr>
<th>N</th>
<th>(a) $A_N$</th>
<th>(b) $A_N$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-0.82188</td>
<td>-1.07517</td>
</tr>
<tr>
<td>2</td>
<td>0.13196</td>
<td>0.22918</td>
</tr>
<tr>
<td>3</td>
<td>0.29147</td>
<td>0.28034</td>
</tr>
<tr>
<td>4</td>
<td>0.13684</td>
<td>0.17123</td>
</tr>
<tr>
<td>5</td>
<td>0.080393</td>
<td>0.084894</td>
</tr>
<tr>
<td>6-20</td>
<td>&lt;0.040</td>
<td>&lt;0.042</td>
</tr>
<tr>
<td>21 (=b)</td>
<td>0.59674</td>
<td>0.72670</td>
</tr>
</tbody>
</table>

Figure 6. Streamlines (solid lines) and isobars (dashed) indicating flow separation for strong barrier wind case ($U_o = 20$ m s$^{-1}$). Separation points indicated by arrow heads. Maximum pressure increase at the origin is 2.8 mb.

has to be at the origin for each problem. The equation for the 1000 m contour island boundary is taken to be

\[ F = -1.188y^2 + 2.303y^4 - 0.739y^6 = 0 \]  

(17)

and the flow separation at the edge of the island is taken to occur at $y_s = 0.94$. The coefficients necessary for this case are given in Table 3. The streamlines and isobars are shown in Fig. 7.
(d) Discussion

The flow patterns in Figs. 6 and 7 are qualitatively similar. The salient feature of the pronounced stagnation zone in Windless Bight is clearly evident, and qualitatively similar to the stagnation zone for flow separation in advance of a flat plate. The maximum pressure occurs at the stagnation point, and is proportional to the square of the background wind speed. As is the case with two-dimensional potential flow, the loss of kinetic energy of an air parcel approaching the island along a streamline is manifested as work done against the pressure field.

For the dynamical model assumed here, the pattern of the streamlines and associated isobars is altered only by the shape of the island, and not by the background wind speed. Then for each island shape, we may change the value of the approaching wind by a constant factor, and then the values of the isobars shown will be increased by the square of this factor. In the case of the climatological winds, the presence of Ross Island does not measurably influence the ambient pressure field. However, for the case of a well-developed barrier wind regime, the island's presence can alter the synoptic pressure field by as much as 2 mb in the vicinity of McMurdo Station.

The calculated pressure rise given in Fig. 6 for Windless Bight is very similar to that observed during the synoptically forced barrier wind event analysed in Fig. 2(b). Also, the observed and modelled approaching wind speeds are nearly identical (19 and 20 m s⁻¹); the former is obtained by correcting the AWS 11 wind speed for frictional slowing. However, as noted in section 1(a), the AWS 11 site may be subject to the upstream influence of Ross Island. These findings indicate that the present modelling results provide a good approximation to actual conditions.

Two aspects of the applicability of the model to the real world should be discussed. For the linear stratification used here with strong barrier winds, the Froude number will be greater than unity for wind speeds in excess of 21.5 m s⁻¹. Then according to the model represented by Eqs. (1) and (2), the approaching wind has sufficient kinetic energy to flow over Ross Island and there will no longer be a stagnation zone in Windless Bight. Although the dividing streamline theory is qualitatively correct, the effects of terrain steepness may become important for higher wind speeds. When the obstacle extends well above the boundary layer, the flow in the synoptically forced upper layer becomes a consideration. For very strong wind speeds (say greater than 25 m s⁻¹) the size of the stagnation zone may be reduced, but a study of this phenomenon would require a three-dimensional model.

The other consideration is the amount by which the wind is increased as it flows around Ross Island in the vicinity of Capes Royds and Crozier. The recurring winter polynyas in these regions are indications of enhanced winds due to orographic modification. However, in the model the maximum wind speed occurs on the free streamline of the wake, and has the same magnitude as the wind speed of the undisturbed flow far upstream of the island. Thus, the region near the flow separation does not have a marked increase in wind speed. We shall discuss this discrepancy between the model results and the pronounced orographic enhancement shown by the observations. In our model the flow separates from the island and an infinite wake downstream is formed. The exact locations of the separation points were not known from observations and thus they were chosen so that the flow separated tangentially from the island along the eastern and western sides. Because this flow does not close behind the island, there is less increase in wind speed at the ends, since the airflow does not have to go around the island. Had it been assumed that the flow separation occurred at some point on the lee side of the island, the bounding streamlines of the wake would have closed behind the island forming a finite wake, and the wind speed would have increased as the air rounded the island.
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Any small topographic protrusions near the ends of the island might markedly increase the wind speed locally, and these smaller topographic effects were neglected in favour of a simple expression that accurately represented the island shape in the bight. It is concluded that the dynamical model of airflow interaction with this simple island shape explains the stagnation zone in Windless Bight, but the increased winds at Capes Royds and Crozier might better be explained with a model topography where the flow rounds the ends of the island before separating on the lee side.

Figure 7. As Fig. 6, but for climatological case \( (U_r = 5 \text{ m s}^{-1}) \). Maximum pressure increase at the origin is 0.175 mb.

ACKNOWLEDGMENTS

The authors wish to thank Dr Jean-Marc Vanden-Broeck for suggesting the computational fluid dynamics approach. Two anonymous reviewers provided suggestions that measurably improved the article. This research was supported by National Science Foundation grant DPP 8314613, and is contribution 574 of the Byrd Polar Research Center, The Ohio State University.

REFERENCES

Arakawa, S. 1973 Numerical experiments on the local strong winds, Bora and
Fohn. Climatological Notes, 14, 1–20

Baines, P. G. 1979 Observations of stratified flow past three-dimensional barriers.
J. Geophys. Res., 84, 7834–7838


Dunn, L. 1987 Cold air damming by the Front Range of the Colorado Rockies and its relationship to locally heavy snows. Weather Forecasting, 2, 177–189

Gjevik, B. 1980 ‘Orographic effects revealed by satellite pictures: Mesoscale flow phenomena’. Ch. 9, pp. 301–316, in Orogenic effects in planetary flows. WMO GARP Pbl. No. 23

Heine, A. J. 1963 Ice breakout around the southern end of Ross Island, Antarctica. New Zealand J. Geol. Geophys., 6, 395–401


Kozo, T. L. and Robe, R. Q. 1981 On the stagnant zone of an obstacle within a stably stratified flow in the atmosphere. ibid., 59, 373–385


AIRFLOW AROUND WINDLESS BIGHT 937


Savage, M. L. and Stearns, C. R. 1985 Climate in the vicinity of Ross Island, Antarctica. ibid., 20(1), 1–9


1979b Meteorological aspects of the drift of ice from the Weddell Sea toward the mid-latitude westertlies. J. Geophys. Res., 84, 6321–6328


Shackleton, E. H. 1909 Cloud investigation by satellite. Ellis Horwood, Chichester


1923 British Antarctic Expedition 1910–1913, Meteorology, Vol. 1, Discussion. Thacker and Spink, Calcutta


