4.5 THE DYNAMICS OF IDEALISED KATABATIC FLOW OVER A MODERATE SLOPE AND ICE SHELF

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1. Introduction

Katabatic winds are downslope buoyancy-driven flows. Over Antarctica and Greenland, the high latitude location leads to a net radiative loss to space during winter which cools the snow surface and overlying atmosphere, and thus, due to the domed topography, generates widespread katabatic flows. This ubiquity means katabatic winds play an important part in the atmospheric general circulation at high latitudes and so a detailed understanding of their dynamics is important for a complete understanding of the high latitude climate system.

In this study, the focus is on 'ordinary' katabatic flows, that is, those that occur over topography that is more typical of coastal Antarctica: relatively modest in slope (maximum ~5%) and uniform across the slope. To be specific, the focus is that of Coats Land and the adjoining Brunt Ice Shelf, as this was the subject of a recent climatological study that raised a number of interesting questions. Renfrew and Anderson (2002) showed that archetypal katabatics flow from around 10° to the east of the fall line and with near-surface wind speeds of 7.5 m s^{-1} at the steepest part of the slope, and 5.1 m s^{-1} higher up (see Figure 1).

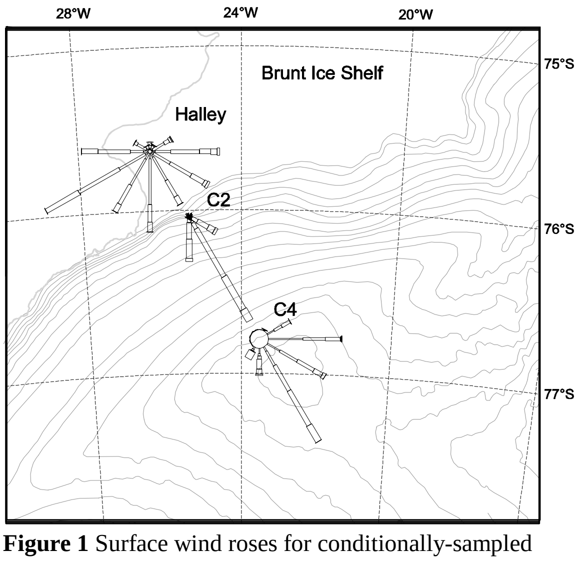

Figure 1 Surface wind roses for conditionally-sampled katabatic conditions, over a topographic map of Coats Land, Antarctica (the contour interval is 100 m).

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Interestingly, they found no coherent surface katabatic flow signature over the Brunt Ice Shelf; an aspect of the local climate also discussed by King (1993). Renfrew and Anderson (2002) found that during katabatic conditions potential temperatures over the ice shelf were also, on average, some 10 K lower than over the continent. They therefore hypothesised that damming due to this pool of cold air in the ice shelf boundary layer was causing the katabatic winds flowing from the continental slopes to ride over the boundary layer and/or be dissipated in some way as the flow decelerates. This study investigates the dynamics of katabatic winds over Coats Land, and as a consequence this hypothesis. The results are pertinent to all 'ordinary' katabatic flows , i.e., those over moderate and uniform slopes; in other words, much of coastal Antarctica and Greenland. The simulations reveal several new facets of katabatic flow dynamics that, due to their generic nature, have relevance to the entire Antarctic climate system.

2. Numerical Model

To investigate the dynamics of Coats Land katabatic winds a state-of-the-art numerical weather prediction (NWP) model, the Regional Atmospheric Modeling System (RAMS), is utilised. A series of idealised model experiments are carried out using a topographic slice of Coats Land. The simulations are initialised with climatological infra-red brightness temperatures at the snow surface and a homogeneous isothermal atmosphere at rest. At the bottom boundary the atmospheric model interfaces with a multi-layer soil model. The soil model is essentially two diffusion equations, for temperature and moisture, with the thermal diffusivity a function of soil-type and moisture content. This has been adapted for use over compacted snow.

3. A wintertime simulation

A coherent downslope flow develops over the first few hours of model simulation, becoming quasi steady-state by ~12 h. Figures 2-4 illustrate the evolution of the downslope flow via model soundings taken at the location of Halley and the AWS sites at C1 and C2. Examining the ice shelf sites, the θ profiles show the development of a strongly stable surface layer, on top of a general background cooling throughout the 500 m depth shown. Examining the ice shelf sites, Halley and C1, the θ profiles show the development of a strongly stable surface layer, on top of a general background cooling throughout the 500 m depth shown. The surface layer is formed as a result of downward sensible heat flux into the snow surface - partially balancing the upward net radiative heat flux. The background cooling is mainly a result of net radiative heat loss to space.

Turning to the velocities, it is evident that there is a transient model response to the initial

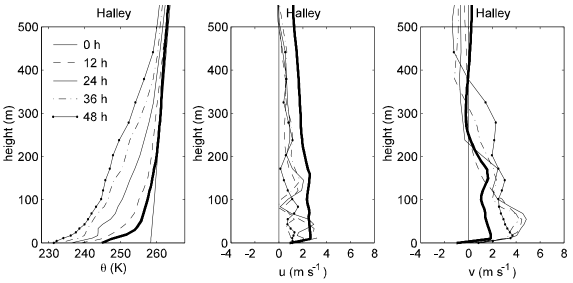

Figure 2 Model soundings of θ, u, and v at Halley for t = 0, 12, 24, 36 and 48 h of simulation time as indicated. Note the dots in the 48-h sounding indicate the model σ_z levels. Overlaid (in bold) are mean radiosonde soundings for when conditions are katabatic at C2 - see text for details.

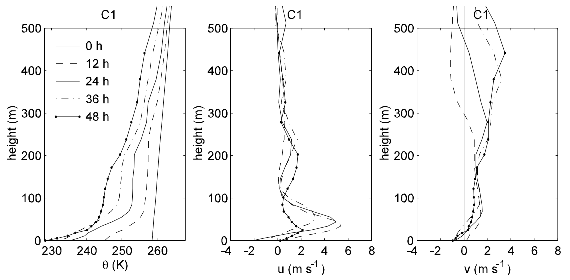

Figure 3 Model soundings of θ, u, and v at C1 for t = 0, 12, 24, 36 and 48 h of simulation time as indicated.

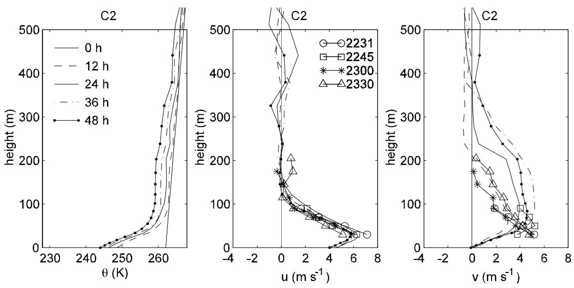

Figure 4 Model soundings of θ, u, and v at C2 for t = 0, 12, 24, 36 and 48 h of simulation time as indicated. Overlaid are four typical velocity profiles for apparently 'pure' katabatic flow on the 24 February 2002. The velocity profiles are preliminary results from an autonomous Doppler sodar wind profiling system.

conditions and forcing. There is a well-defined jet in u at C1 from 12-24 h, but it is fading by 36 h, and is not really discernable by 48 h. The katabatic flow retreats up the slope (past the C1 site) over the 48-h period. At Halley, one can also see transient evidence of this flow in both u and v. Note that one would expect a backing of the flow on the ice shelf, relative to that on the slope, due to a change in the balance of forces, i.e., there is no downslope buoyancy forcing, so a simple inertial balance (with frictional effects) should apply. By the end of the simulation the ice shelf sites have settled to low wind speeds, with two weak maxima in the lowest 500 m. At Halley, one can also see transient evidence of katabatic flow in both u and v.

In Figure 4a well-defined downslope flow develops after less than 12 h and is in approximately a steady-state over the rest of the 48-h period. At C2, approximately the steepest part of the slope, the surface layer is around 100 m deep, and is less stratified than over the ice shelf or further up the slope. The downslope flow peaks in a 6-7 m s^{-1} jet at about 30 m above the surface. Interestingly, the height of the v-component 'jet' tends to increase with time. This is a geostrophic response to the continuous cooling over the ice shelf.

During the simulation, there are times when the atmosphere becomes temporarily statically unstable as a result of the advection of cold air over a relatively warm surface; for example, at t = 48 h at the C3 site. This occurs because the rate of surface cooling is not uniform throughout the model domain, rather the surface temperature is a prognostic variable calculated from the model's surface energy balance. This means spatial differences in (for example) wind speed will lead to different surface sensible heat fluxes and thus different surface and near-surface temperatures. Advection can then lead to a statically unstable surface layer, the buoyant generation of TKE, and a mixing out of the instability to create a neutral layer. The mixing warms the near-surface atmosphere and so restores stability; which means the periods of static instability tend to be short-lived and the neutral layers vertically confined. An example of such a neutral layer is evident around 20-30 m above the surface in the 24 h sounding at Halley (Fig. 2). Its elevated position is due to further atmospheric cooling after the period of instability.

4. An analysis of the katabatic flow dynamics

The forcing terms in the downslope momentum budget have been calculated following the framework set out in Mahrt (1982). A two-layer model is assumed, that is, an active lower layer and a quiescent upper layer. The downslope momentum equation for the lower (katabatic) layer, is

$$\frac{Du_{kls}}{Dt} = \frac{g\Delta\theta}{\theta_0}\sin\alpha - \cos\alpha\frac{g}{\theta_0}\frac{\partial(\Delta\theta\, h_{inv})}{\partial x} + fv_{kls} - \frac{\partial\overline{(u'w')}}{\partial z},$$

where u_{kls}, v_{kls}, are katabatic-layer average velocities, θ_0 is a reference potential temperature (here taken as the upper-layer θ), $\Delta\theta$ is the potential temperature deficit between the lower and upper layers (i.e. $\Delta\theta = \theta_0 - \theta_{kl}$), g is the gravitational acceleration, α is the positive slope angle, f is the Coriolis parameter, and h_{inv} is the height of the katabatic layer (defined below). Note in this section the coordinate system (x, y, z) indicates directions downslope, cross-slope and perpendicular to the slope with the corresponding velocity components $(u_{kls}, v_{kls}, w_{kls})$. For the moderate slopes of the topography considered here $\sin\alpha \approx \alpha$ and $\cos\alpha \approx 1$. Integrating over the katabatic layer the terms can be broken into the following: $F_{adv}$is the total advection (i.e. the inertial force); F_b is the buoyancy (or katabatic) force; $F_{\Delta\theta}$ and F_h are the so called thermal wind forces (although F_b is also related to thermal wind effects - Mahrt 1982), specifically $F_{\Delta\theta}$ is due to the gradient in potential temperature deficit, and F_h is due to the gradient in katabatic-layer height; F_{Cor} is the Coriolis force; and F_{div} is the momentum flux divergence force. In addition, there is a forcing due to the downslope component of the model's horizontal diffusion; however this turns out to be negligible in these simulations and so is not discussed any further.

Figure 5 illustrates the forcing terms of the downslope momentum budget. The terms are calculated using 3-h mean data. The illustration is

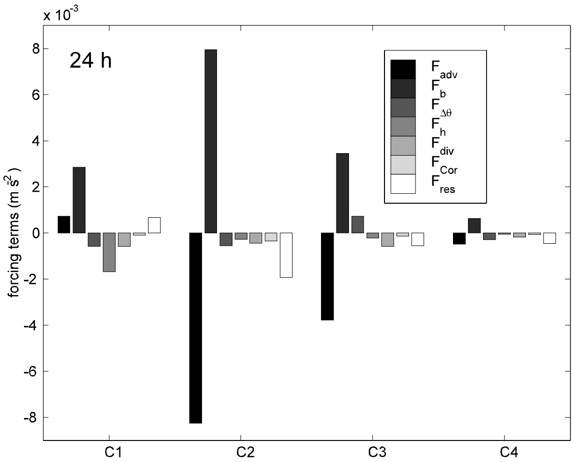

Figure 5 Forcing terms in the downslope momentum equation after 24 h simulation time: F_{adv} is the advection term, F_b is the buoyancy force, $F_{\Delta\theta}$ is due to the gradient in potential temperature deficit, F_h is due to the gradient in katabatic-layer height, F_{div} is the momentum flux divergence force, F_{Cor} is the Coriolis force, and F_{res} is the residual.

after 24 h simulation time, but the results are representative of any time after ~12 h. It is clear that higher on the continental slope, at C2, C3 and C4, there is a balance between the advection term, F_{adv}, and the buoyancy forcing F_b. The thermal wind terms and the stress divergence generally retard the flow, but are much smaller. The final term shown, F_{res}, is the residual and is due to approximations made in using this two layer model to analyse a continuously stratified flow over a varying slope (the model is strictly only applicable for constant slopes - Mahrt 1982). At C1, the foot of the slope, the balance of forces is quite different with F_b balanced by F_h, $F_{\Delta\theta}$ and F_{div}. The buoyancy force is balanced by a combination of up-slope forcing primarily caused by the pool of potentially cold air that forms over the ice shelf. This acts to dam the downslope flow, leading to the deceleration seen at C1. Given the archetypal nature of the simulation, this up-slope thermal-wind related forcing is thought to be the reason for the observed difference in surface wind climatology on the slopes of Coats Land and the Brunt Ice Shelf discussed in the Introduction (see Fig. 1).

The two-layer analysis framework employed above allows the calculation of a local Froude number (Fr):

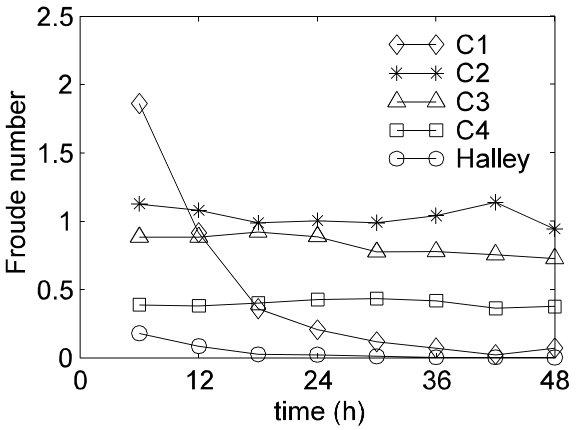

Figure 6 Model time series of Froude number at Halley and C1 to C4 using 3-h mean data.

$$Fr = \frac{u_{kls}^{\;2}}{(g\Delta\theta \;/\; \theta_0)h_{inv}}.$$

In terms of hydraulic theory (e.g. Ball 1956; Mahrt 1982) the flow is shooting if $Fr > 1$ and is tranquil if $Fr < 1$. Figure 6 shows that, in this case, the katabatic flow moves from tranquil (C4) to shooting (C2) to tranquil (C1 / Halley) down the slope. At C2, Fr is around 1 to 1.2 so only just in the shooting regime. At C1 the initially shooting flow becomes tranquil over time. In a strictly two-layer fluid flow, these transitions of flow type would be through flow discontinuities or hydraulic jumps; indeed, such jumps have been observed in katabatic flows (e.g. Lied 1964; Pettré and André 1991). In this case, the Fr variations are relatively small and, as the model is multi-level, the flow remains continuous.

In the wintertime Coats Land simulations, the katabatic flow cessation is caused solely by cold-air damming, leading to up-slope $F_{\Delta\theta}$ and F_h forces, there are no diurnal effects. Furthermore as the simulations are non-hydrostatic, the transition from shooting to tranquil flow is modelled more realistically (Cassano and Parish 2000). This is demonstrated in Fig. 7 which shows a cross-section of vertical velocity, w, at 24h.

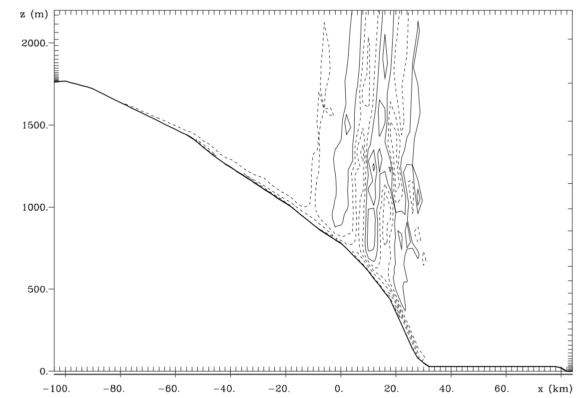

Figure 7 Cross-section of vertical velocity *(w)* at 24 h. The contours are at ...-0.075, -0.025, 0.025, 0.075,... m s^{-1} with negative contours dashed.

Figure 7 shows there is descent into the shallow katabatic layer, as one would expect from continuity constraints, but also that there is a train of internal gravity waves over the continental slope. A comparison with contemporaneous θ cross-sections shows a $\pi/2$ phase lag - as one would expect from internal gravity waves. Furthermore, a rough calculation (e.g. Holton 1992) shows that with a horizontal wavelength ~10 km and a Brunt-Vaisala frequency ~4x10^{-4} s^{-1}, the horizontal phase speed should be ~1 m s^{-1}, consistent with the phase speed of the waves in the simulation. The gravity waves appear to be triggered near the foot of the slope - at the transition from shooting to tranquil flow. The horizontal phase speed of the waves is in the $-x$ direction (i.e. up-slope), consistent with an upward group velocity. This means the gravity waves are propagating energy upwards - away from the katabatic flow jump. All of the katabatic flow simulations carried out exhibit internal gravity waves, which suggests they are a common feature of katabatic flows over Coats Land. As the 2-dimensional topography used here is representative of much of coastal Antarctica, this suggests they are ubiquitous for areas prone to katabatic winds. A number of previous studies have also demonstrated a transition from shooting to tranquil flow near the coast of Antarctica (e.g. Gallée and Schayes 1992; Gallée and Pettré 1998; Gallée *et al.* 1996), but as these were hydrostatic studies, the generation of internal gravity waves was not possible. To generalise the non-hydrostatic implications of this study suggests an active source of vertically propagating internal gravity waves around the Antarctic continent. If this is the case, there may be important implications for gravity-wave parameterisation in large-scale numerical models.

5. Conclusions

A state-of-the-art NWP model has been used to simulate idealised katabatic flows over a moderate slope and the adjoining ice shelf. The simulations are for wintertime clear-sky climatological conditions and are initialised with the atmosphere at rest. A shallow katabatic flow develops over the first 12 hours or so to a quasi steady-state. The modest downslope jet has a peak of ~7 m s^{-1} about 30 m above the surface and a total depth of 50 to 100 m. There is a band of high *TKE* at the top of the katabatic layer. Over time, the katabatic flow retreats from the ice shelf and some way up the slope, while the ice shelf boundary layer cools. The pool of cold air causes an up-slope thermal wind-related forcing which counteracts the downslope buoyancy forcing and decelerates the flow. At the transition, the downslope flow changes from shooting to tranquil. This triggers a train of internal gravity waves which propagate energy upwards, away from the katabatic flow jump. To generalise these findings would suggest a source of internal gravity waves, as a by-product of katabatic flow deceleration, around much of coastal Antarctica and Greenland. This has implications for modelling the general atmospheric circulation and climate of these areas which clearly merits further study.

The model simulations have been validated against near-surface and sounding observations of archetypal and typical katabatic conditions in Coats Land and the adjoining Brunt Ice Shelf, Antarctica. The archetypal katabatic conditions are means of conditionally-sampled subsets of data that have small perturbation pressure gradients and large stabilities (see Renfrew and Anderson 2002). The typical katabatic soundings are from a novel autonomous Doppler Sodar wind profiling system in Coats Land. In general, the model corresponds well to the observational data, capturing the scale and shape of the katabatic flow, especially after some tuning of the turbulence parameterisation. In the horizontal, the scalar diffusion coefficient (K_H) is set as a constant proportion of the momentum diffusion coefficient (K_M), and it was found that $K_H/K_M \approx 0.5$ was optimal. It is suggested that the model's surface-layer parameterisation (based on Louis 1979) may not be entirely adequate, as towards the top of the slope, the surface-layer stabilities are too large and model turbulence is suppressed. Occasionally *TKE* is generated at the model's snow surface by cold-air advection, a process that results in persistent elevated neutral layers, such as those that have been observed at Halley on the Brunt Ice Shelf.

References

Please contact the author for a copy of the full-length paper which contains the full reference list. The observational paper discussed above is:

Renfrew, I. A. and P. S. Anderson, 2002: The surface climatology of an ordinary katabatic wind regime in Coats Land, Antarctica, *Tellus* ,**54A,** 463-484.