The structure of katabatic flows down a simple slope

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SUMMARY

An analysis of the katabatic flow structure is presented, based on data collected along a simple slope over a two-month period. Four meteorological stations were deployed along the nearly two-dimensional east slope of Mt. Hymettos in Greece. The study focuses on the momentum and energy budgets of the flows during carefully selected stationary periods with negligible ambient winds. In order to achieve a proper selection, the mean features of the observed flows are presented along with their response to external winds. The observed profiles are successfully compared with the theoretical Prandtl profiles. Bulk quantities of the katabatic layer are compared with proposed experimental values in order to check the representativity of the observed flows. These comparisons offer a further test of theoretical considerations and assist in discussions of processes affecting mean-flow features.

The flows were ~ 20 m deep, with velocities 1–2 m s⁻¹. During steady-state periods the katabatic acceleration was mainly balanced by the surface friction, the contribution of advective terms being small. The estimated interfacial drag was less important than the surface drag, which proved large compared with usually suggested values. Within the experimental uncertainties, values of the drag coefficient of the order of 10⁻² seem appropriate. The steady-state energy balance showed that the downslope advection balanced the turbulent heat flux to the ground surface.

KEYWORDS: Down-slope winds Mesoscale field-study Orography

1. INTRODUCTION

Thermally driven flows observed over tilted surfaces have been the subject of many studies, particularly in the context of the major ASCOT programme (Dickerson and Gudiksen 1984; Clements et al. 1989). The complexity of real topography has dictated a categorization of such flows (Atkinson 1981). Most of the flows observed and numerically simulated were down-valley drainage flows and downslope flows along valley side-walls. The former are deeper and stronger than katabatic flows along a simple slope. However, even though the local flows developed at the interior of valleys are relatively decoupled from the external meteorology, their dynamics are complicated by the interaction between slope and valley flows. In this case slope flows are difficult to isolate, embedded as they usually are in the valley flow. On the other hand, the simple approach of analysing slope flows along isolated slopes has to deal with their sensitivity to external disturbances due to limited topographic sheltering. This was evident, for example, for the data presented by Doran and Horst (1983) and Horst and Doran (1986). The latter attempted a detailed study of slope flows along a simple slope, but the ambient wind was not negligible and, as the authors noted, ‘the wind direction was generally not downslope within the katabatic layer’. Therefore, the evaluation of their results with regard to the dominant terms of the momentum budget was contaminated by ambient winds.

This motivated an experimental campaign at the nearly two-dimensional east slope of Mt. Hymettos in Greece. Moreover, downslope flows are an unstudied feature of the nocturnal circulation regime in the Athens Metropolitan Area (AMA, Fig. 1), since past studies concentrated on the daytime phase of local flows, mainly sea-breezes.

The analysis focuses on the steady-state momentum and energy budgets of katabatic flows, aiming to provide useful data on the principal mechanisms involved. The data were

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carefully selected to avoid incorrect interpretation of the budget’s terms. The methodology for this choice involves the following steps:

(i) determination of the frequency of occurrence of downslope flows at the east slope of Mt. Hymettos,

(ii) identification of the background conditions promoting katabatic flow development,

(iii) presentation of the mean features of the flow (wind and temperature profiles, jet height, bulk parameters), and

(iv) interpretation of the external wind’s effect on the flow structure, in order to be able to identify and isolate periods of pure katabatic motion.

Steps (iii) and (iv) include a comparison of measured parameters with those of previous experimental or theoretical efforts, and guarantee that the budget analysis is as reliable as possible within the unavoidable experimental uncertainties. A significant part of the discussion concerns the two dominant retarding mechanisms of katabatic flows, i.e. the surface and the interfacial drag.

In addition to the steady-state periods examined here, the interesting evolution of the katabatic flows at the same site of Mt. Hymettos has been studied by Helmis and Papadopoulos (1996); the morning and evening transition phases of the flow, including micro-scale frontal characteristics of its initial phase, are the subject of subsequent work.
TABLE 1. Instrumentation of meteorological stations A–D with measuring heights above the ground surface

<table>
<thead>
<tr>
<th>Station</th>
<th>Cup anemometer (m)</th>
<th>Wind vane (m)</th>
<th>UVW propeller (m)</th>
<th>Thermometer (m)</th>
<th>Hygrometer (m)</th>
<th>Turbulence probes (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>10.0</td>
<td>10.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>2.5, 9.0</td>
<td>9.0</td>
<td></td>
<td>2.5, 5.4, 9.0</td>
<td>5.4</td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>2.5, 5.6, 8.5, 14.8, 19.7, 5.6, 5.6, 12.1, 19.7, 29.1</td>
<td>5.6, 19.7, 12.1, 29.1</td>
<td>0.2, 0, (soil)</td>
<td>2.5, 5.6, 11.5, 12.1, 19.7, 29.1</td>
<td>6.5, 8.5</td>
<td>or</td>
</tr>
<tr>
<td>D</td>
<td>3.0</td>
<td>3.0</td>
<td></td>
<td></td>
<td>4.0</td>
<td>8.5</td>
</tr>
</tbody>
</table>

For station locations see Fig. 2.

2. Experimental layout

The east slope of Mt. Hymettos (Fig. 1) was selected for the measuring campaign of the autumn 1992 experiment (10 September to 31 October) because it is quite homogeneous around the experimental site. There are sparse olive trees on the lowest part of the slope, and some scattered pine trees and juniper bushes at mid-slope. The east slope can be considered as two dimensional (for a 3 km cross-slope distance) at least regarding the small extent of the observed katabatic flows. The mountain ridge-line is 15 km from the Messogia Plain, runs parallel to the east coast, and peaks at 1024 m, lowering to the north and south of the experimental site.

Meteorological stations (depicted in Fig. 2) were deployed on the ridge (station A, height 900 m), on the slope (station B, height 410 m), near the foot of the slope (station C, height 240 m) and over the plain (station D, height 120 m). The instrumentation details for each station are given in Table 1. The downslope direction is from 270°, and the slope of the surface is 34° and 9° around B and C respectively. The surface roughness length, $z_0$, was computed from near-neutral logarithmic wind speed profiles at the foot of the slope, yielding median values of 15 cm for north-west wind directions, 5 cm for northerly and 10 cm for southerly winds. Neutral wind speed profiles for westerly directions (from where the katabatic flows blow) deviated significantly from the logarithmic law, therefore no $z_0$ value could be computed. However, the homogeneity of the ground surface, along with the $z_0$ values estimated above, lead to an approximate value of 10 cm for the katabatic flow direction.

The data recorded by stations A, B and D were stored as mean and fluctuating quantities at 10 min intervals, while sampling and storing was set at 1 s$^{-1}$ for station C. An additional 10 m pole, installed next to the 29.1 m mast at location C, was instrumented with two turbulence probes, developed at the University of Athens (Papadopoulos 1995). The two systems, each based on a triple hot-wire sensor attached to a wind vane, were operated during selected periods at sampling rates of 1 or 20 s$^{-1}$, providing data for the turbulent flow structure.

All instruments were intercalibrated at the beginning and end of the experiment.

3. Data analysis
(a) General features of the wind flow over the experimental site during the campaign

According to Amanatidis et al. (1992), the katabatic flows on the west slope of Mt. Hymettos for the period January to June 1990 were most frequently observed in April and May. Considering that spring and autumn have a similar weather pattern in Athens, and that
Figure 2. (a) Two- and (b) three-dimensional views of the east slope of Mt. Hymettos with the locations of meteorological stations A–D.
TABLE 2. THE WIND DIRECTION DISTRIBUTION (PER CENT) AT ALL EXPERIMENTAL SITES FOR DAYTIME (0630-1800 H (GMT + 2 HOURS)) AND NIGHT-TIME (1800-0630 H, IN PARENTHESES) CONDITIONS FOR THE EXPERIMENTAL PERIOD

<table>
<thead>
<tr>
<th>Station</th>
<th>N</th>
<th>NE</th>
<th>E</th>
<th>SE</th>
<th>S</th>
<th>SW</th>
<th>W</th>
<th>NW</th>
</tr>
</thead>
<tbody>
<tr>
<td>HMS*</td>
<td>16</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>25</td>
<td>25</td>
<td>21</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>(19)</td>
<td>(0)</td>
<td>(0)</td>
<td>(0)</td>
<td>(23)</td>
<td>(19)</td>
<td>(33)</td>
<td>(9)</td>
</tr>
<tr>
<td>A</td>
<td>5</td>
<td>31</td>
<td>9</td>
<td>4</td>
<td>12</td>
<td>6</td>
<td>25</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>(12)</td>
<td>(19)</td>
<td>(6)</td>
<td>(1)</td>
<td>(13)</td>
<td>(13)</td>
<td>(30)</td>
<td>(4)</td>
</tr>
<tr>
<td>B</td>
<td>12</td>
<td>27</td>
<td>15</td>
<td>23</td>
<td>11</td>
<td>1</td>
<td>5</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>(8)</td>
<td>(3)</td>
<td>(1)</td>
<td>(4)</td>
<td>(16)</td>
<td>(11)</td>
<td>(21)</td>
<td>(37)</td>
</tr>
<tr>
<td>C (5.6 m)</td>
<td>13</td>
<td>31</td>
<td>9</td>
<td>9</td>
<td>22</td>
<td>4</td>
<td>4</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td>(18)</td>
<td>(2)</td>
<td>(0)</td>
<td>(1)</td>
<td>(10)</td>
<td>(13)</td>
<td>(20)</td>
<td>(36)</td>
</tr>
<tr>
<td>C (29.1 m)</td>
<td>11</td>
<td>31</td>
<td>13</td>
<td>10</td>
<td>21</td>
<td>3</td>
<td>2</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td>(31)</td>
<td>(5)</td>
<td>(1)</td>
<td>(3)</td>
<td>(15)</td>
<td>(11)</td>
<td>(12)</td>
<td>(23)</td>
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<tr>
<td>D</td>
<td>10</td>
<td>29</td>
<td>15</td>
<td>6</td>
<td>16</td>
<td>14</td>
<td>4</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>(30)</td>
<td>(14)</td>
<td>(4)</td>
<td>(2)</td>
<td>(4)</td>
<td>(11)</td>
<td>(13)</td>
<td>(22)</td>
</tr>
</tbody>
</table>

For station locations see Figs. 1 and 2.
* The radiosonde data have been averaged for the layer 1000-2000 m, roughly corresponding to the 850 hPa level, i.e. close to the mountain crest height. The radiosondes were released at 1400 h and 0200 h.

autumn is usually drier than spring. September and October were selected for observing the katabatic flows at the east slope. This choice satisfies the dual purpose of avoiding the dominant summer and winter weather patterns which suppress local circulations, and of observing katabatic flows under the varying wind conditions of a transient season.

The identification of the background conditions permitting the development of local flows over the AMA was based on radiosonde wind data from the upper-air station of the Hellenic Meteorological Service (HMS, Fig. 1) at the 850 hPa level, and data from the ridge-top station A. Throughout the paper, the continuous measurements at the ridge-top station are considered representative of the background flow over the experimental area, because of the similarity of the corresponding wind direction distributions at stations A and HMS (see Table 2). The basic difference of the ridge-top wind direction distribution from that at HMS (850 hPa level) is due to the development of upslope motion (easterly winds) over the mountain slope during days with weak synoptic forcing.

The wind direction distributions of stations B–D (Table 2) imply a dominant daily cycle of a local thermal circulation with easterly winds during the daytime, and weak westerly flows (< 2 m s⁻¹) starting after sunset. The former are linked to upslope winds over the slope (B, C) and to the development of a plain-to-mountain circulation (D), while the latter are linked to katabatic and mountain-to-plain flows. A progressive shift from westerly (of katabatic origin) to northerly winds with increasing measuring height is apparent at mast C (Table 2), indicating the shallowness of the katabatic layer.

(b) Frequency of occurrence of katabatic flows and conditions promoting their development

A weak synoptic forcing and clear skies are necessary conditions for the formation of the nocturnal surface inversion that drives the katabatic flow. Katabatic flows quickly respond to ambient conditions, since well developed flows are easily eroded through the processes of warm advection or vertical turbulent mixing.
Near the foot of the slope (C) wind direction profiles show the katabatic forcing to decline with height. The 29.1 m level usually overlies the katabatic layer: for weak ambient winds the 29.1 m level is characterized by negligible velocities, for non-negligible ambient wind it features wind directions that significantly deviate from downslope. In the second case, approaching the top of the katabatic layer the wind usually shifts to the dominant nocturnal N–NW direction of the Messogia Plain, due to the outflow from the Penteli and Hymettos Mts. (Fig. 1). This is verified by occasional tethersonde flights.

Katabatic flows were more frequently observed at the foot-slope station than at mid-slope, because the latter was influenced by the ridge-top flow. They were observed at the foot even during high ridge-top wind speed conditions (mostly from the west), while lower limits for the ambient winds were quoted in previous studies (3 m s\(^{-1}\), Horst and Doran (1986) and 6 m s\(^{-1}\), Dickerson and Gudiksen (1984)). This feature (for westerly winds) is probably linked to the steepness of the upper slope of Mt. Hymettos, leading to a flow separation in the lee of the ridge. Westerly winds often produce a southerly flow in the lee of Mt. Hymettos (Varvayanni et al. 1993). This flow favours the katabatic wind, probably prohibiting the southwards propagation of the nocturnal N–NW wind regime (see above) over the Messogia Plain. It might also explain the differing degree of interaction between the katabatic flow and the larger scale wind; i.e. the observation that northerly winds are more effective than westerly winds in eroding the katabatic flow. Indeed, NE background winds below 6 m s\(^{-1}\) did not affect the katabatic flows, while stronger NE winds inhibited them in 80% of the cases. The flows that developed under such conditions ultimately merged with the nocturnal NW wind regime over the Messogia Plain.

The nocturnal potential temperatures at B and C were similar to each other, but higher than over the plain. The main part of the night-time cooling over the slope was completed by 1900–2000 h (GMT+2 hours), followed by an additional temperature drop of 1–2 K by sunrise. Sometimes the surface air actually warmed, although the ground surface cooling continued; this implies replacement of the cold surface air by warmer air masses from aloft, maintaining a thermal belt (Kobayashi et al. 1994) over the slope, with temperatures exceeding those over the plain. The total nocturnal cooling at mid-slope was greater than at the foot, and never exceeded 13 K.

Eleven nights of sustained katabatic motion were found during the experimental period. The median values of total cooling at B and C were 7.5 K and 6.3 K, respectively, for non-draining nights and 8.1 K and 7.8 K for the draining nights. The small difference between the total cooling for draining and non-draining nights demonstrates the importance of ambient conditions in the development of katabatic flows. Indeed, the basic difference between draining and non-draining nights was the intensity and direction of the regional winds. This point needs some explanation. The non-draining nights were mostly characterized by NW winds originating as a cold outflow from Mt. Penteli (see above), therefore causing cooling by advection, probably comparable with the radiation cooling observed on draining nights. On the other hand, the radiation cooling of draining nights was limited by the usual intrusion of a moist southerly flow late in the afternoon related to the sea-breeze from the Saronikos Gulf (see Fig. 1). As a result, the total cooling difference between draining and non-draining nights was small. Based on our observations it was estimated that the effect of draining on the near-surface cooling rate was about 1 K h\(^{-1}\).

\(c\) Wind speed and temperature profiles—bulk quantities of the katabatic layer

The main features of the katabatic flow over a simple slope are presented and compared with theoretical predictions and experimental studies (usually for valley flows) with the following objectives:
TABLE 3. The jet height of the katabatic flow at station C (see Fig. 2) as a function of the magnitude and sign of the external flow, measured at the highest measuring level (29.1 m) on the mast

<table>
<thead>
<tr>
<th>Jet height (m)</th>
<th>Occurrences</th>
<th>Downslope wind component at 29.1 m (m s⁻¹)</th>
<th>Frequency distribution (per cent)</th>
<th>((T_{19.7} - T_{2.5})^*) (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.5</td>
<td>6</td>
<td>&lt; 0.5</td>
<td>3</td>
<td>2.0</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.5 to 1.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total 7</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5.6</td>
<td>1</td>
<td>2.5 to 1.5</td>
<td>75</td>
<td>2.3</td>
</tr>
<tr>
<td>1</td>
<td>-2.5 to -1.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>151</td>
<td>-1.5 to -0.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>23</td>
<td>0.5 to 1.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total 176</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8.5</td>
<td>11</td>
<td>-0.5 to 0.5</td>
<td>13</td>
<td>2.5</td>
</tr>
<tr>
<td>20</td>
<td>0.5 to 1.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total 31</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>12.1</td>
<td>11</td>
<td>-0.5 to 0.5</td>
<td>9</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>0.5 to 1.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>1.5 to 2.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total 21</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Mean temperature values for the case when |μ| < 0.5 m s⁻¹, |v| < 0.5 m s⁻¹ at the top of mast C, located near the foot of the slope.

(i) to provide katabatic-flow measurements over a simple slope under weak ambient winds;

(ii) to show the effect of ambient wind on the wind profiles, and discuss its implications with respect to the mean-flow dynamics; indeed, the knowledge of the characteristic depth, velocity and buoyancy scales of the katabatic layer, as well as their quantitative inter-relations, is of importance when attempting simulations of the katabatic air volume flow rates, at least in the usual case where limited experimental resources are available; and

(iii) to provide guidelines for selecting periods of pure katabatic motion for a reliable budget analysis (section 3(d)).

Relatively steady periods of katabatic flow at the foot of the mountain were used to identify the features of the pure flow. A typical flow is shown in Fig. 3; a weak 12 m deep flow developed by 1900 h, then gradually strengthened and attained maximum intensity during 0300–0700 h, responding to the temperature profile evolution. Between 0700 and 0800 h the dilution of the surface inversion caused the elimination of the flow. The characteristic maximum in the wind speed profile (the ‘jet’) was evident.

The jet height, \(n_{\text{max}}\), a measure of the katabatic flow depth, is usually found around the middle of the layer and is affected by the ambient wind. As noted in section 3(b), the wind measurements at the 29.1 m level of mast C may be taken as a rough measure of the ambient wind immediately above the katabatic layer, since the pure flow is characterized by a depth of ~ 20 m. Thirty-minute mean profiles of the downslope component (\(u\), along 270°) during katabatic motion were selected, rejecting cases with \(u_{\text{max}} < 0.5\) m s⁻¹. This limit is dictated by the instrument’s measuring threshold. The 29.1 m wind was < 0.5 m s⁻¹ and deviated from the downslope direction in 76% of the cases (Table 3), suggesting that the katabatic flow was shallower than 29.1 m and that the immediate external forcing was minimal. The most frequent case featured a jet at 5.6 m (keeping in mind the height
resolution of wind measurements), a total depth less than 30 m and negligible wind speed at 29.1 m, indicating the typical pure flow for the particular experimental site. Deviations from this pattern could be attributed mainly to external influences. Accordingly, the appearance of the jet at other measuring levels of mast C is attributed to the presence of an ambient downslope wind component above the main katabatic layer, as will be proved below with the aid of Fig. 4.

When periods of weak external winds were considered in order to exclude disturbances on the vertical flow structure, \( n_{\text{max}} \) was found to be proportional to the intensity of surface cooling (last column of Table 3).

These conclusions are applied to the discussion of Fig. 4, where the experimental profiles are plotted against those proposed by Prandtl (1952) (see appendix B for details).
A satisfactory agreement between the experimental and theoretical profiles was found when $n_{\text{max}}$ 'equalled' 5.6 or 8.5 m. The inclusion of cases with non-negligible external wind introduced deviations in the upper part of the profiles which is most vulnerable to external forcing (compare Fig. 4(f) with 4(g) and 4(h) with 4(i)). The existence of the jet, with the associated small wind shear and the stronger stability near the ground, assists in the decoupling of the lowest part of the katabatic layer from the flow aloft.

The upper part of the mean experimental wind speed profile with a jet at 2.5 m (Fig. 4(e)) was similar to the experimental and theoretical profiles with a jet at 5.6 m (Fig. 4(f)), while the 12.1 m cases (Fig. 4(j)) were characterized by increased scatter. Best agreement was found for the 5.6 m case. When $n_{\text{max}}$ exceeded 5.6 m the measured profiles showed a faster decline of the temperature deficit of the katabatic layer, $d$, than in the Prandtl profiles. The opposite trend was found for the 2.5 m cases. This could be
Figure 4. Comparison of the mean normalized profiles of $u$ and $d$ (see text) at station C (see Fig. 2) (squares) with the Prandtl profiles (solid lines, Eqs. (B.3), (B.4)) for different jet heights. Averaging time is 30 minutes. Velocity normalization is performed with respect to the maximum velocity. Temperature deficit is computed with respect to the temperature at 19.7 m (where $d$ is assumed zero) and normalized to its value at 2.5 m. Horizontal bars are proportional to the scatter of experimental values. The jet height, the number of available experimental profiles and the upper limit of the external wind speed (as measured) at 29.1 m are indicated in each plot.
explained by assuming that the 2.5 m and 12.1 m cases involved, essentially, disturbed appearances of the typical katabatic flow with a jet at 5.6 m. Indeed, the measured profiles of $d$ did not differ appreciably between these jet-height classes, supporting the conclusions drawn from Table 3.

The bulk quantities of the katabatic layer to be used in section 3(d) were introduced by Manins and Sawford (1979a):

$$
\begin{align*}
U h &= \int u \, dn \\
U^2 h &= \int u^2 \, dn \\
U \Delta h &= \int u g d / \Theta_0 \, dn
\end{align*}
$$

(1)

where $U$, $h$, $\Delta$ are the velocity, height and buoyancy scales of the katabatic layer, respectively. All symbols used are defined in appendix A. The same authors proposed the use
TABLE 4. CHARACTERISTIC BULK PARAMETERS OF THE KATABATIC LAYER AT STATION C (SEE FIG. 2) AND COMPARISON WITH THEORETICAL PREDICTIONS

<table>
<thead>
<tr>
<th></th>
<th>h (m)</th>
<th>h' (m)</th>
<th>z_{inv} (m)</th>
<th>U/u_{max}</th>
<th>n_{max}/h</th>
<th>S_1</th>
<th>S_2</th>
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<tbody>
<tr>
<td>Median</td>
<td>19</td>
<td>4</td>
<td>0.19</td>
<td>16</td>
<td>0.8</td>
<td>0.3</td>
<td>0.49</td>
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<tr>
<td>Lower quartile</td>
<td>16</td>
<td>3</td>
<td>0.16</td>
<td>13</td>
<td>0.7</td>
<td>0.2</td>
<td>0.4</td>
</tr>
<tr>
<td>Upper quartile</td>
<td>22</td>
<td>5</td>
<td>0.24</td>
<td>20</td>
<td>0.9</td>
<td>0.4</td>
<td>0.7</td>
</tr>
<tr>
<td>Manins and Sawford (1979a)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.5</td>
<td>0.9</td>
</tr>
<tr>
<td>Ellison and Turner (1959)</td>
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<td></td>
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<td></td>
<td></td>
<td>0.2-0.3</td>
<td>0.6-0.9</td>
</tr>
<tr>
<td>Nappo and Rao (1987)</td>
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<td></td>
<td></td>
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<td>0.64</td>
<td>0.14</td>
</tr>
<tr>
<td>Prandtl profiles</td>
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<td></td>
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<td></td>
<td></td>
<td>0.78</td>
<td>0.39</td>
</tr>
</tbody>
</table>

See text for explanation of headings and β.

The integrations cover the whole layer. They incorporate significant scatter for the wind speed profiles due to external influences and experimental uncertainties, particularly for a quite weak flow. The integration of temperature profiles in the katabatic layer is more accurate, since they possess less scatter (Fig. 4) and d slowly approaches zero at the top of the katabatic layer (Figs. 3 and 4). Based on this, Doran and Horst (1983) proposed alternative height and buoyancy scales:

\[
S_1 \Delta h^2 = 2 \int gd/\Theta_0 dn \tag{2}
\]

\[
S_2 \Delta h = \int gd/\Theta_0 dn. \tag{3}
\]

The integrations cover the whole layer. They incorporate significant scatter for the wind speed profiles due to external influences and experimental uncertainties, particularly for a quite weak flow. The integration of temperature profiles in the katabatic layer is more accurate, since they possess less scatter (Fig. 4) and d slowly approaches zero at the top of the katabatic layer (Figs. 3 and 4). Based on this, Doran and Horst (1983) proposed alternative height and buoyancy scales:

\[
\begin{align*}
\Delta' h' &= \int gd/\Theta_0 dn \\
\Delta' h'^2 &= \int gd/\Theta_0 dn. \tag{4}
\end{align*}
\]

According to Horst and Doran (1986), h ≈ z_{inv}, the surface inversion depth. Combining (2), (3) and (4) and using the Manins and Sawford’s values it is found that h'/h = S_1/2S_2 = 0.28.

It is worth estimating the characteristic parameters describing the integral framework of katabatic flows, since adequate data from simple slopes are not very common. Thirty-minute averages during periods of calm external winds have been used. The comparison with theoretical values is included in Table 4. It is verified that h almost equals the inversion depth, and that h'/h equals 0.18 or S_1/2S_2 according to the computed values of the shape factors. Significant deviations from the simulated values of Nappo and Rao (1987) were observed for the ratios U/u_{max} and n_{max}/h, the measured values of which were large. The overestimation with regard to theoretical values can be explained if the measured profiles
have a less sharp jet at a greater height, as expected over a rough surface supporting satisfactory vertical mixing in stable conditions (Kondo and Sato 1988). Briggs (1981) also noted that the existence of scattered trees (see section 2) favours such a change in the profile shape. On the other hand, the computed values of $U/u_{\text{max}}$ and $n_{\text{max}}/h$ based on the Prandtl profiles are very close to the experimental values, in accordance with the results shown in Fig. 4, and also the measurements of Clements et al. (1989) for a down-valley flow which yielded $n_{\text{max}}/h = 0.20 \pm 0.04$. The same authors quoted a value of 0.25 for simple slope flows.

Fitzjarrald (1986) suggested that the similarity between the Prandtl and measured profiles depends on the entrainment rate of ambient air into the katabatic layer as it deepens down the slope. When entrainment is vigorous, the upper part of the profiles deviates from the Prandtl profile shape, which is based on the assumption of constant diffusivities, $K$, in the katabatic layer. Indeed, Rao and Snodgrass (1981) used height-dependent $K$s to enable greater upward mixing of the downslope momentum than for the constant $K$ case. Based on the agreement with the Prandtl profiles, it is argued that the deviations observed in the upper part of the profiles can be attributed to the external flow’s influence (compare Figs. 4(e), (f) and (g)) rather than produced from increased entrainment (also see the next subsection). Thus, the assumption of constant $K$ is applicable for the description of shallow katabatic flows, which are usual over a simple isolated slope.

The last two columns of Table 4 refer to the shape factors. The laboratory experiments of Ellison and Turner (1959), which should be considered representative of atmospheric drainage rather than slope flows, yielded higher values than the present ones. However, Manins and Sawford’s (1979a) estimates for a drainage flow in the presence of lateral flow convergence are rather close. Nappo and Rao (1987) computed the shape factors using the numerically simulated profiles of a katabatic flow along a simple slope and found them to be dependent on the slope angle $\beta$, particularly for $\beta < 20^\circ$. Their values for $\beta = 9^\circ$ (Table 4) are remarkably close to the present results.

(d) Dynamics of stationary katabatic flows

Steady-state periods of the katabatic flow at stations B and C were analysed to establish the dominant terms in the momentum and energy balance at the foot of the slope. The analysis was based on thirty-six 10 min periods satisfying the following requirements: (i) the downslope component of the external flow, measured at the top of mast C, was lower than 0.5 m s$^{-1}$ to avoid contributions from external downslope momentum; (ii) the integrated cross-slope momentum through mast C was lower than 50% of the integrated downslope momentum; and (iii) the flow was steady and the wind speeds at B and C were correlated.

(i) Momentum balance. The integral form of the momentum balance in the downslope direction is given by Manins and Sawford (1979a) as:

$$\frac{\partial}{\partial t} \int u\,dn + \frac{\partial}{\partial s} \int u^2\,dn = -\frac{\partial}{\partial s} \int \frac{g}{\Theta_0} n \cos \beta \,dn + g \int \frac{d}{\Theta_0} \sin \beta \,dn + (u'w')_0$$

where $d > 0$. The limits of integration are from the surface (actually at $n = z_0$) to a height incorporating the katabatic layer depth. In the present analysis the top of the mast (C) has been taken as the upper integration limit.

Term II is the along-slope advection term, IV is the katabatic acceleration and V is the surface drag. Term III is a retarding factor due to the downslope increase of stability. The time-tendency (storage) term I is zero for stationary periods. For the periods chosen, its
maximum value was 0.01 m²s⁻², with a typical value of 0.005 m²s⁻². As a first approximation it was assumed that \( u'w' = 0 \) at the upper limit of integration. An assessment of this is given later in this section. Measured values at 29.1 m were of the order of 0.001 m²s⁻². Using (1), term II becomes,

\[
\frac{\partial(U^2h)}{\partial s} = U \frac{\partial(Uh)}{\partial s} + Uh \frac{\partial U}{\partial s} = EU^2 + Uh \frac{\partial U}{\partial s} \tag{6}
\]

where the entrainment coefficient, \( E \), is defined by:

\[
E = U^{-1} \frac{\partial(Uh)}{\partial s}. \tag{7}
\]

The form of the first term on the right-hand side of (6) suggests that \( E \) is an interfacial drag coefficient.

The derivatives in (5) for any parameter \( X \), say, were computed as \( (X_C - X_B)/s \). A zero velocity at the surface was assumed in the integration of wind profiles. The temperature deficits were computed with respect to the top of the masts, a reasonable assumption since the temperature gradient approached zero above 5 m at mid-slope and 15 m at the foot. Thus, the relevant terms were probably slightly underestimated. The lower limit of integration of the temperature profile was taken at the ground surface. As the ground surface temperature was only measured at the foot, a linear extrapolation of the profile to the ground was performed at the mid-slope station.

Due to the considerable scatter, the median value of each term is given in Table 5. The scatter was substantially less for the terms involving integration of temperature profiles, as noted above. Term V has been computed in two ways: (i) as a residual of the temperature balance (last column in Table 5), and (ii) using the measured \( \bar{u}'w' \) value from the turbulence probe at a height of 4 m above the ground. The extrapolation of this quantity to its surface value will be analysed later in this section.

The advective terms (II + III) could balance only 10–20% of the katabatic acceleration. Specifically, term III alone balanced only 2% of IV, in agreement with Horst and Doran (1986). The same holds true for the experimental data reviewed by Mahrt (1982), with the exception of the flow studied by Mahrt and Larsen (1982), where the slope angle was only 2°, which resulted in a significant contribution from the thermal-wind term, as defined by Mahrt (1982).

The possible contribution from the flow non-stationarity could be of either sign, and in the best case could balance an additional 10% of the katabatic acceleration. Therefore, the surface drag should balance 70–80% of the katabatic acceleration at C (last column of Table 5).

Fortunately, the availability of turbulent momentum flux measurements allows this conclusion to be verified. The \( \bar{u}'w' \) value in Table 5 (fourth column) was measured by the eddy-correlation method at 4 m above ground. The 10 min averaging time is usually short.
for flux estimations (the problem is less serious in stable conditions), but longer averaging times did not significantly alter the results. It should be noted that the accuracy of turbulent flux estimations is around 20% according to Doran et al. (1989).

The measured value of $\bar{u}'w'$ was of the correct order of magnitude for ‘closing’ the momentum balance, but four times smaller than the calculated residual (term V in (5)) that referred to the surface value. Therefore, a formula is needed to link $(\bar{u}'w')_{z=4 \text{ m}}$ with its surface value $(\bar{u}'w')_0$.

The height of measurement of $(\bar{u}'w')_{z=4 \text{ m}}$ is close to the jet height, where the reduction of shear weakens the turbulent momentum flux. This statement was supported by the observation of an increasing trend in $-\bar{u}'w'$ values at 4 m, as the jet height increased, but available data were not sufficient to prove it definitely.

The following formula can be applied for stable conditions over a horizontal surface (Stull 1988):

$$\frac{-\bar{u}'w'}{u_s^2} = 1 - \left( \frac{z}{h_1} \right)^{0.7}$$  \hspace{1cm} (8)

where $h_1$ is the depth of the stable boundary layer at the top of which the turbulent fluxes are zero. The approximation $h_1 \approx n_{\text{max}}$ was used, because measurements showed a relative decoupling of the layer below $n_{\text{max}}$ from the layer above. Substituting $z = 4 \text{ m}$ and $h_1 = 5-6 \text{ m}$ in (8), a value of 0.21 to 0.25 was found for the left-hand side, or $u_s^2 = (-\bar{u}'w')_0 = 4.24 - 3.60 \times 10^{-2} \text{ m}^2\text{s}^{-2}$, i.e. very close to the value of the last column of Table 5.

Term V may be also expressed as:

$$(\bar{u}'w')_0 = -C_D U^2$$  \hspace{1cm} (9)

where $C_D$ is the surface drag coefficient and can be estimated from (9). In a flux-interface equation like (9), a measure of velocity at, say, 10 m above the surface is used. This selection is not straightforward for a slope flow exhibiting a maximum in the near-surface wind speed profile. Although the velocity-scale used in (9) is of the same order of magnitude as the true velocities of the katabatic layer, the exact $C_D$ value is sensitive to the velocity parameter used (e.g. $U$ or $u_{\text{max}}$). Substituting the estimated value of $U$ (alternatively $u_{\text{max}}$) in (9), it was found that median values of $C_D \approx 0.07$ (0.05 if $u_{\text{max}}$ is used) when term V is estimated as a residual, or 0.06 to 0.07 when estimated through the extrapolated surface value of $\bar{u}'w'$.

The exact value of $C_D$ is difficult to calculate, but its order of magnitude determines the dominant retardation mechanism (surface $C_D$ or interfacial $E$ drag) and some features of its time evolution (McNider 1982). Manins and Sawford (1979a) suggested the dominance of interfacial drag, but recent studies (Manins 1992) have indicated the equal importance of surface drag. Typical $C_D$ values for stable conditions over horizontal surfaces lie between $10^{-4}$ and $10^{-3}$ (Manins and Sawford 1979a; Mahrt 1982), but higher values are calculated from experimental data. Horst and Doran (1986), for example, calculated values between 0.03 and 0.04. Some comments are necessary with regard to possible explanations for the discrepancy between the values proposed in the literature, and those found here and by Horst and Doran (1986). The latter attributed the large values to the increased near-surface shear in the presence of the maximum in the velocity profile.

(a) The effect of the small height of the mid-slope mast can be ruled out, because it could only lead to underestimations of the velocity and temperature deficit integrals at B, therefore decreasing the residual term V.

(b) The katabatic acceleration term IV could be overestimated in cases of an ambient downslope wind increasing the residual term V. However, the periods chosen for the budget analysis were carefully selected to avoid external influences.
(c) The significant change of slope downstream from the mid-slope station could trigger a portion of the katabatic layer at mid-slope to leave the slope surface and reduce the value of the advective term II. Therefore, term III was re-estimated by splitting the integral into two parts to take into account the effect of the slope change $\partial \beta / \partial s$; yet, only a 1% difference from the initial value of term III was found.

(d) The simplification of the turbulent flux divergence term may contribute; i.e. term V was taken as $(\overline{u'w'})_{0}$ rather than $(\overline{u'w'})_{0} - (\overline{u'w'})_{H}$. However, considering that the measured values of $\overline{u'w'}$ at the top of the foot-slope mast were of the order of 0.001 m²s⁻², the $C_{D}$ estimate would be reduced by 10%. This interfacial stress was found very important in the analysis of Manins and Sawford (1979b) who proposed a formula for its calculation indicating that it is proportional to the fourth order of the velocity difference between the katabatic layer and the ambient wind. Using the suggested values it is found that for a velocity difference $\delta U = 0.5$ m s⁻¹, $(\overline{u'w'})_{H} = 0.001$ m²s⁻². For the periods selected, the maximum $\delta U (= 1$ m s⁻¹) occurred for ambient upslope winds, therefore $(\overline{u'w'})_{H}$ could contribute no more than 0.016 m²s⁻² of the residual term V, and the minimum value of $C_{D}$ could be lowered by 30%, bringing it down to 0.05.

(e) The exact choice of the characteristic velocity to be used in (9) may be important. According to the derivation of (5), $U$ should be used. On the other hand, it could be anticipated that the existence of the near-surface jet determines the surface stress, allowing for the use of $u_{max}$.

(f) The approximation of derivatives by finite differences introduces some errors.

(g) Equation (9) is strictly valid for a continuously turbulent process, whilst the momentum flux is known to be often intermittent in stable conditions, with major contributions from turbulent bursts.

The entrainment coefficient $E$ may be calculated either from (7) or from the equation proposed by Briggs (1981) for slope angles larger than 12°:

$$E = 0.05 (\sin \beta)^{2/3}.$$  

Equation (10) seems applicable at the foot of the mountain, yielding a value of 0.015 ($\beta = 9^\circ$), while the estimate based on our data and (7) is 0.015 ± 0.007.

Thus, the ratio $E / C_{D}$ is estimated to be 20 to 30%. Horst and Doran (1986) estimated a value of 50% for the drag coefficient ratio, stating that it could be expected to change along the slope, an aspect not included in analytic models.

The above calculations, which referred to the katabatic flow as a bulk layer, showed that the steady katabatic flows at the foot-slope station are primarily retarded by surface friction, classifying them as equilibrium shooting flows (Mahrt 1982). This also explains the similarity between the experimental and Prandtl profiles, which assume a negligible contribution from along-slope advection to the momentum balance. It could be argued that the present results strictly apply to the lowest part of the layer while, above the jet height, the katabatic acceleration and surface drag are reduced and advective terms become important (see Fig. 3 for the effect of increasing external flow on the profiles; also Fitzjarrald (1986)). It should be anticipated that advective terms become important, at least in the first stage of development of the katabatic flows at the foot of the slope, especially when frontal characteristics are evident.

(ii) Energy balance. The integral form of the energy balance for stationary periods is (Manins and Sawford 1979a):

$$U h N^{2} (\sin \beta - E \cos \beta) + \frac{\partial}{\partial s} (U \Delta h) = - \frac{g}{\Theta} (\overline{w' \Theta'})_{0}.$$  

$$1 \quad II \quad III$$
The turbulent heat flux (term III) is expressed normal to the underlying surface, and \( N \) is the Brunt–Väisälä frequency \((N = g \Theta^{-1} d\Theta/dz)^{1/2}\), a typical value of the temperature gradient is 0.005 K m\(^{-1}\)). Terms I and II are associated with the advection of the ambient flow’s thermal characteristics, and the downslope advection of the katabatic layer’s buoyancy, respectively. Term III will be estimated as a residual. The radiation flux divergence term has been neglected because the katabatic layer was shallow and the change of air temperature within it was usually about 2 K. Calculations by Nieuwstadt and Driedonks (1979) showed this term to be of the order of \(10^{-7}\) for a 20 m deep layer in stable conditions. Derivatives were approximated as for the momentum balance.

<table>
<thead>
<tr>
<th>Term I</th>
<th>Term II</th>
<th>Storage or time-tendency term (omitted)</th>
<th>Term III (residual)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Median</td>
<td>((Fr &gt; 1))</td>
<td>4.1</td>
<td>-0.3</td>
</tr>
<tr>
<td></td>
<td>((Fr &lt; 1))</td>
<td>3.2</td>
<td>0.5</td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>((Fr &gt; 1))</td>
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<td>2.0</td>
</tr>
<tr>
<td></td>
<td>((Fr &lt; 1))</td>
<td>1.0</td>
<td>1.0</td>
</tr>
</tbody>
</table>

All terms are expressed in m\(^2\)s\(^{-3}\) \(\times\) \(10^{-4}\).

* The Froude number \(Fr\) is given by \(Fr = U^2/g'h\).

The change of \(U \Delta h\) along the slope is small (Table 6), but has a preference for positive (negative) values when the Froude number \(Fr < 1\)(> 1). An increase of the downslope wind speed \((Fr > 1)\) produces a larger value of term I. At the same time term II acts in the opposite direction, since the total flux of buoyancy is reduced along the slope. The theoretical results of Nappo and Rao (1987) show that \(U \Delta h \propto s\) for a near-neutral ambient stability and a constant slope angle. A likely explanation for the discrepancy is the decrease of the slope angle from B to C, which causes a reduction of the katabatic acceleration even for constant \(d\).

The sum of terms I and II in (11) is rather independent of the \(Fr\) value and equals \(3.8 \times 10^{-4}\) m\(^2\)s\(^{-3}\). The required heat flux normal to the ground surface is \(13 \pm 6\) W m\(^{-2}\). Unfortunately, the heat flux was not measured near the surface and was effectively zero above \(n_{max}\). Typical values of heat fluxes during katabatic flows are found in the literature to be 10 to 30 W m\(^{-2}\).

Thus, the foot-slope energy balance in steady conditions is accomplished between the advection of warmer air down the slope and the transfer of turbulent heat flux to the ground surface. Prandtl’s simplified approach is then applicable (instead of (11)) for the thermal structure of the katabatic layer:

\[
UhN^2 \sin \beta = -\frac{g}{\Theta} \langle w'\Theta' \rangle_0. \tag{12}
\]

It should be mentioned that Manins and Sawford (1979a) argued that, as the night progresses and stability is increased, the layer-cooling process is dominated by the radiation flux divergence while the turbulent heat flux at the surface decreases. However, the magnitude of the surface heat flux is also determined by the katabatic flow’s intensity which also grows as the surface inversion grows.
4. CONCLUSIONS

The east slope of Mt. Hymettos offers a favourable site for observing katabatic flows over a simple slope. In the past, slope flows have been mainly studied in complex topography involving a multitude of valleys, tributaries etc. and were usually embedded in deep down-valley or drainage flows. On the other hand, although an ideally simple flow may be expected to be found over a relatively isolated slope, in practice the limited topographic sheltering favours the interactions with (or the erosion by) phenomena of larger scale. Such processes of local interest have not been discussed in the preceding analysis.

The katabatic flows were more organized and deeper at the foot of the slope than at mid-slope. Their depth was about 20 m and a jet (1–2 m s\(^{-1}\)) was found below the middle of the layer. The jet height was sensitive to the wind speed and direction immediately above the katabatic layer, but the profiles below the jet were well protected from external influences.

Notwithstanding the fact that the weakness of the katabatic flows over isolated slopes may reduce the validity of delicate computations, the following interesting conclusions were reached.

Observed vertical profiles of wind and air temperature conformed with the Prandtl profiles, the agreement being explained by the justification of the simplifying assumptions of Prandtl during the experimental conditions. The analysis of steady-state periods of katabatic flow showed that the katabatic acceleration was balanced at 70–80% by the surface friction, while advective terms were of lesser importance. The small contribution of advective terms may be associated with the weak and shallow nature of the observed flows, rather than with increased surface roughness. A quite large value of surface drag was computed, and was found to be 3–5 times the interfacial drag. Although \(C_D\) was verified by two independent methods, possible mechanisms for spuriously increasing its value were discussed. It seems that the estimations of Horst and Doran (1986) are realistic, and values of the order of \(10^{-2}\) should be considered for katabatic flows, at least for equilibrium shooting flows. The correct \(C_D\) values may be of considerable importance in numerical simulations of the nocturnal boundary layer over inclined surfaces, and further research is necessary.

The energy balance of the steady-state katabatic flow was accomplished between the advection of warmer air down the slope and the turbulent heat flux to the ground surface.

Finally, bulk quantities of the katabatic layer, computed by vertically integrating wind and temperature profiles, compared well with theoretical values. In particular, the shape factors were found to agree with the results of the numerical model of Nappo and Rao (1987).

APPENDIX A

List of symbols

\(C_D\) surface drag coefficient
\(d\) temperature deficit of katabatic layer (K)
\(E\) entrainment coefficient
\(Fr\) Froude number of katabatic layer (\(U^2/g'h\))
\(g\) gravitational acceleration (m s\(^{-2}\))
\(g'\) reduced gravity (\(gd/\Theta_0\)) (m s\(^{-2}\))
\(h, h'\) depth scales of katabatic layer (m)
APPENDIX B

Derivation of Prandtl profiles

Prandtl dealt with the problem of a laminar flow over an incline. If the turbulent diffusivities replace the molecular ones in the equations used by Prandtl then, for steady conditions over a homogeneous surface ($\partial d / \partial s = 0$), the momentum and energy balances become ($\Theta = \Theta_0 + d$):

\[
0 = -g \frac{d}{\Theta_0} \sin \beta - \frac{\partial u'w'}{\partial n} \Leftrightarrow 0 = -g \frac{d}{\Theta_0} \sin \beta + K_m \frac{\partial^2 u}{\partial n^2} \tag{B.1}
\]

\[
0 = -u \frac{\partial \Theta}{\partial s} - \frac{\partial \Theta'w'}{\partial n} \Leftrightarrow 0 = -u \frac{\partial \Theta_0}{\partial s} - \frac{\partial \Theta'w'}{\partial n} \Leftrightarrow
\]

\[
0 = -u \frac{\partial \Theta_0}{\partial z} + K_h \frac{\partial^2 d}{\partial n^2} \Leftrightarrow 0 = uy \sin \beta + K_h \frac{\partial^2 d}{\partial n^2} \tag{B.2}
\]

with the assumption of height-independent diffusivities. It has been further assumed that $\partial \Theta_0 / \partial n \ll \partial d / \partial n$, and that the slope is infinitely long so that the advective terms of the momentum equation may be neglected. The solutions of the two differential equations are

\[
U(\eta/1) = U_c e^{-(\eta/1)} \sin(\eta/1) \tag{B.3}
\]

\[
U_c = \frac{d(0)N}{\gamma} \left( \frac{K_h}{K_m} \right)^{1/2}
\]
\[ d(n/1) = d(0)e^{-(n/1)\cos(n/1)} \]
\[ 1 = \left( \frac{2K_m}{N \sin \beta} \right)^{1/2} \left( \frac{K_h}{K_m} \right)^{1/2} \]
\[ N = \left( \frac{g}{\Theta_0 v} \right)^{1/2} \]
\[ (B.4) \]

The ratio \( K_h/K_m = 1.35 \), \( K_m = 0.06 \text{ m}^2 \text{s}^{-1} \) as in Rao and Snodgrass (1981). Using data from katabatic periods, \( K_m \) near the ground surface was estimated by

\[ K_m = -\frac{\mu'\nu'/\Delta z}{\Delta u} \]
\[ (B.5) \]

where the turbulent momentum flux was measured at 4 m by the turbulence probe and the wind speed difference \( \Delta u \) between the 2.5 and 5.6 m levels of mast C. The median value was 0.07 \text{ m}^2 \text{s}^{-1} with upper and lower quartiles at 0.04 and 0.14 \text{ m}^2 \text{s}^{-1}.

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