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SIMULATION OF KATABATIC WINDS AT MIZUHO STATION, ANTARCTICA

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Abstract A new simulation method for solving dynamic equations for stationary katabatic wind is suggested by the assumption that the turbulent exchange coefficient is a function of height and the effect of the cold sloping surface is a multinomial function of height. Calculated wind profiles agree with observational data at Mizuho Station, Antarctica.

Key word simulation, katabatic wind, Mizuho Station.

Introduction

Katabatic wind is a typical phenomenon over Antarctic plateau. It has been investigated by many researchers (Adachi, 1983; Adachi, 1984; Mahrt and Schwerdtfeger, 1970). Mahrt and Schwerdtfeger (1970) assumed the force caused by cold sloping surface is an exponential function of the height. Adachi (1984) assume that pressure gradient force is a linear function of the height. But they took the turbulent exchange coefficient K as a constant and hence the wind profile does not depend on the surface roughness. From measurements and numerical simulations of boundary layer, however, we know it is not the case. One of shortcomings is that it may cause disagreement of wind profiles with observational data of surface layer (Adachi, 1984).

This paper presents a new analytical simulation of katabatic wind profiles in consideration of non—constant of turbulent exchange coefficient. Solutions show that the wind profile also depend on the surface roughness. The simulation result is agreeable with observational data at Mizuho Station (70°41′53″S, 44°19′54″E, 2230m a.s.l.), Antarctica.

The study on katabatic winds may be helpful in both forecasting katabatic winds and understanding baroclinicity of boundary layer.

General equations and their solutions

1. *General equations*

Horizontally dynamic equations can be written as

$$\frac{d(Kdu/dz)}{dz} - f(v - vg) = g\,\frac{\theta u - \theta}{\bar{\theta}}\frac{\partial h}{\partial x} \tag{1a}$$

$$\frac{d(Kdu/dz)}{dz} - f(u - ug) = g\,\frac{\theta u - \theta}{\bar{\theta}}\frac{\partial h}{\partial y} \tag{1b}$$

where h is height of terrain above sea level, f is absolute value of Coriolis parameter, θu is undisturbed potential temperature, θ is potential temperature profile, and $\bar{\theta}$ is mean potential temperature.

We introduce the following assumptions:

a) Turbulent exchange coefficient $K = \beta\eta(1 - \alpha\eta)u^* H$; $0 < \alpha < 1$, β and α are constants. u^* is friction velocity, H is boundary layer height, and η is dimensionless height which is equal to z/H where z is the height:

b) The force caused by cold sloping surface $\vec{F}c = g\,\frac{\theta u - \theta s}{\bar{\theta}}(1 - \eta)^n(\frac{\partial h}{\partial x}\frac{\partial h}{\partial y})$

where θs is potential temperature at surface, n is taken as 3 in this paper.

This form of K has been first introduced by Yikoyama (1977) basing on observations of neutral boundary layer and adopted by Nieuwstadt (1983) for non—adiabatic boundary layer. In their expression α is equal to 1. However, we do not think it is true because numerical simulation (Adachi, 1983) shows K is larger than 0 at the top of boundary layer. Besides it does not satisfy the boundary conditions for the equations(1a, 1b) as $K=0$ (Nieuwstadt, 1983). For the two reasons we assume α is less than 1. The adoption of α will be discussed in section 2. 3. Here we should also point out that althrough such an adoption of K is more liable than that of constant K, a flaw exists when it is used in stable boundary condition as it was deduced from neutral condition.

We also use a linear variation of pressure gradient force with the height (Adachi, 1984).

$$ug = ug_0 + C_{ug}\eta \tag{2a}$$

$$vg = vg_0 + C_{vg}\eta \tag{2b}$$

where ug_0 and vg_0 are components of pressure gradient force at the surface, C_{ug} and C_{vg} are constants.

First we take $W = u + v$i and (1a) + (1b)i. Considering the two assumptions and the formulas (2a, 2b), we tranform the equations (1a, 1b) into dimensionless equations:

$$\frac{d}{d\eta}[\eta(1 - \alpha\eta)\frac{dW}{d\eta}] + C\mathrm{i}W = \sum_{k=0}^{n} a_k\eta^k \tag{3a}$$

where $C = fH/\beta u^*$, which is a stability parameter (Nieuwstadt, 1983).

$$a_0 = C\mathrm{i}(ug_0 + vg_0\mathrm{i}) + Fc_0C_0^n$$

$$a_1 = C\mathrm{i}(C_{ug} + C_{vg}\mathrm{i}) - Fc_0C_1^n$$

$$a_k = Fc_0C_n^k(-1)^k;\ k = 2, 3, \ldots\ldots n \tag{3b}$$

$$Fc_0 = (g/f)C[(\theta u - \theta s)/\bar{\theta}](\partial h/\partial x + \mathrm{i}\partial h/\partial y)$$

2. *Solutions*

Let $\eta' = a\eta, W' = W - \sum_{k=0}^{n} b_k \eta'^k$, then we transform equation (3a) to ahypergeometric differential equation.

$$\frac{d}{d\eta'}\left[\eta'(1-\eta')\frac{dW'}{d\eta'}\right]+i\frac{C}{a}W'=0 \tag{4a}$$

where b_k (k = 0, 1, 2, ..., n) can be deduced from

$$b_n=\frac{a_n a_{-n}}{-a_1 n(n+1)+Ci} \tag{4b}$$

$$b_k=\frac{a_k a^{-k}-a(k+1)^2 b_{k+1}}{-ak(k+1)+Ci}, \qquad k=n-1,\cdots,2,1 \tag{4c}$$

The equation (4a) can be solved (see Wang, 1979). Then

$$W=C_1F(a,b,1,\eta')+C_2\{F(a,b,1,\eta')ln\eta'+\sum_{j=0}\frac{(a)j(b)j}{(j!)^2}[\psi(j+a)-\psi(a)+\psi(j$$
$$+b)-\psi(b)-2\psi(j+1)+2\psi(1)]\}+\sum_{j=0}^{n}b_k\eta'^k \tag{5}$$

where $a=0.5+0.5(1+\frac{4Ci}{a})^{\frac{1}{2}}$; $b=0.5-0.5(1+\frac{4Ci}{a})^{\frac{1}{2}}$, F is hypergeometric function and ψ is psi—function (Wang, 1979).

Substituting the two boundary conditions for η', we can find C_1 and C_2, then get the solutions.

3. *Boundary layer conditions*

As the first solution of the hypergeometric differential equationdiverges at η' = 1 and the second diverges at both η' = 0 and 1, we cannot set boundary conditions at these two points. They can be set from the following conditions:

$$z=z, \quad u=0; \quad v=0 \tag{6a}$$

$$z=H, \quad u=ug_0+C_{ug}; \quad V=vg_0+C_{vg} \tag{6b}$$

into

$$\eta'=\frac{z_0}{H}; \quad W'=-\sum_{k=0}^{n}b_k(\frac{az_0}{H})^k \tag{6c}$$

$$\eta'=a; \quad W'=(ug_0+C_{ug})+(vgo+C_{vg})i-\sum_{k=0}^{n}b_k a^k \tag{6d}$$

4. *Adoption of parameter a*

Here we use a method which Adachi (1983) introduces to calculate the K at the top of boundary layer (K_{Top})

$$K_{Top}=\frac{\kappa z\left|\frac{\tau}{\rho}\right|^{\frac{1}{2}}}{\varphi} \tag{7a}$$

$$\varphi=(1-12\frac{Z}{L})^{\frac{1}{2}} \tag{7b}$$

$$\frac{\tau}{\rho}=K_{Top}\frac{d\,vg}{dz} \tag{7c}$$

$$\frac{q}{C_p\rho}=-K_{Top}\frac{d\theta}{dz} \tag{7d}$$

$$\frac{Z}{L}=\frac{\kappa\frac{g}{\theta}\frac{q}{C_p\rho}z}{|\frac{\tau}{\rho}|^{\frac{3}{2}}} \tag{7e}$$

$$a=1-\frac{K_{Top}C}{fH^2} \tag{7f}$$

where τ is shear stress, q heat flux, C_p specific heat, L local value of Monin—Obukhov length, φ non—dimensional wind shear function and κ Von Karman constant ($\kappa=0.41$).

From equations (7a—7e) we can obtain K_{Top} at the top of boundary layer by iteration method, that is, setting initial K_{Top} into equations (7c, d), we can find $\frac{\tau}{\rho}$ and $\frac{q}{C_p\rho}$. Substitute them into equations (7b, e) and 7(a), we obtain a new K_{Top}. Then a can be obtainedfrom equation (7f).

5. *Observational data*

We adopted aerological sounding data, which were sampled at Mizuho Station indifferent months, 1980 (Kawaguchi *et al.*, 1985). The terrain slope is 3×10^{-3}. $\frac{\partial h}{\partial x}$ and $\frac{\partial h}{\partial y}$ are taken as 2.67×10^{-3} and -1.67×10^{-3}, respectively, f is 1.4×10^{-4} and g is 9.8m/s^2.

Results and Discussion

Table 1 shows samples of observed vertical temperatures T, two horizontal wind components (U_0, V_0), necessary parameters used in this method, and calculated wind components (U, V) with this method. The calculated wind components agree with observed ones. Here z_0 is equal to 0.1mm (Adachi, 1983). The x—axis, is to east and y to north.

In order to understand influence of roughness z_0 on wind profiles, we present calculated two wind components with different z_0 in Figure 1 from data in Table1(a). The influence of roughness on wind profiles is mainly in surface layer.

The disagreement between the calculated and observed data may be mainly caused by the short accuracy of terrain slope $\partial h/\partial x$, $\partial h/\partial y$ and observation of katabatic wind.

Table 1. Comparison of calculated and observed components of katabatic wind.

(a)

April 24, LT 13:24, Po=741.0hPa

$C=7.0$; $ug_0=-3.0$, $C_{ug}=11.0$, $vg_0=10.0$, $C_{vg}=3.0$(m/s), $K_{Top}=0.082 m^2/s$

Z(m)	T(°C)	Z(m)	U(m/s)	U0(m/s)	V(m/s)	V0(m/s)
0	−46.1	10	−12.57	−11.50	0.90	0.00
36	−31.5	102	−11.40	−9.51	1.68	3.09
165	−26.3	310	−5.61	−3.30	4.01	5.95
330	−22.8	516	−0.88	2.20	6.24	7.69
495	−22.4	727	2.78	3.55	8.22	10.94
1040	−25.8	938	5.41	6.34	9.81	11.92
1440	−27.2	1142	7.16	7.70	11.02	10.22
		1340	8.00	8.06	13.00	12.89

(b)

April 28, LT 17:03, Po=724.6hPa

$C=5.5$; $ug_0=6.0$, $C_{ug}=9.5$, $vg_0=8.3$, $C_{vg}=0.0$(m/s), $K_{Top}=0.014 m^2/s$

Z(m)	T(°C)	Z(m)	U(m/s)	U0(m/s)	V(m/s)	V0(m/s)
0	−44.7	10	−3.44	−6.50	0.26	0.00
50	−34.3	130	−0.68	−0.29	1.08	1.16
155	−31.4	398	5.41	8.31	2.94	5.00
210	−31.2	667	9.89	12.01	4.55	4.14
320	−29.1	937	12.97	14.32	5.82	3.31
480	−28.1	1152	14.61	15.33	6.59	8.85
905	−29.4	1368	15.50	15.63	8.30	8.31
1260	−31.7					

(c)

April 14, LT 10:26, Po=737.4hPa

$C=30.0$; $ug_0=-8.0$, $C_{ug}=2.5$, $vg_0=0.0$, $C_{vg}=-1.5$(m/s), $K_{Top}=0.027 m^2/s$

Z(m)	T(°C)	Z(m)	U(m/s)	U0(m/s)	V(m/s)	V0(m/s)
0	−30.9	10	−12.51	−12.31	−6.00	−2.17
150	−23.0	187	−10.56	−10.29	−5.25	−6.18
225	−20.7	570	−7.48	−4.95	−2.26	−5.90
375	−20.5	948	−6.16	−3.24	−1.36	−1.31
765	−23.4	1302	−5.50	−5.51	−1.50	−1.48
910	−23.8					
1125	−24.8					

(d)

July 7, LT 13:11, Po=740.0hPa

$C=2.0$; $ug_0=-12.0$, $C_{ug}=9.0$, $vg_0=-2.0$, $C_{vg}=-2.0$(m/s), $K_{Top}=0.017 m^2/s$

Z(m)	T(°C)	Z(m)	U(m/s)	U0(m/s)	V(m/s)	V0(m/s)
0	−36.3	10	−13.23	−13.80	−2.44	0.00
50	−34.3	125	−14.53	−14.23	−3.61	−2.77
75	−25.9	387	−12.61	−7.99	−4.82	−7.99
305	−22.0	678	−10.30	−6.30	−5.67	−4.93
415	−22.5	991	−8.16	−7.19	−6.26	−4.15
585	−24.1	1317	−6.29	−5.43	−6.54	−4.72
1000	−27.0	1629	−3.00	−3.08	−4.00	−3.94
1600	−32.0					

Conclusions

By the assumptions about turbulent exchange coefficient and force caused by cold sloping surface, this analytical solution agree with observed data at Mizuho Station, Antarctica. Adoption of turbulent exchange coefficient K and stability parameters C remain to be studied further.

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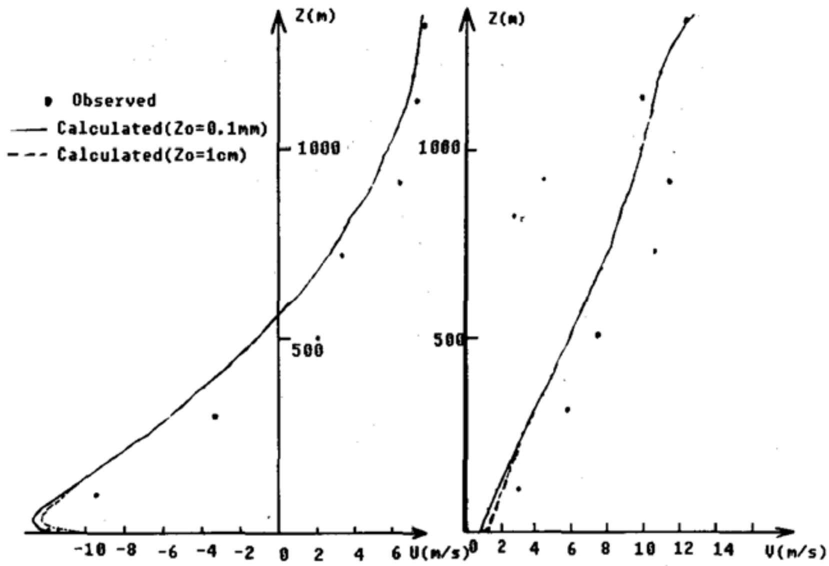

Fig. 1 Comparison of the observed wind components with calculated ones under two conditions of z_0

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References

Adachi, T., (1984): Analytical solutions of katabatic wind at Mizuho and Syowa stations, Antarctica. *Memoirs of national institute of polar research, Special issue*, No. 34, Proceedings of the sixth symposium on polar meteorology and glaciology.

Adachi, T., (1983): Numerical simulation of strong katabatic winds at Syowa and Mizuho stations, Antarctica. Proceedings of the fifth symposium on polar meteorology and glaciology, National institute of polar research, Japan

Kawaguchi, A., *et al.*, (1985): *JARE data reports*, No. 104 (Meteorology 17).

Mahrt, L. J. and Schwerdtfeger, W., (1970): Ekman spirals for exponential thermal wind, *Boundary — Layer Meteorology*, 1, 137—145

Nieuwstadt, F., (1983): On the solution of the stationary, baroclinic Ekman—layer equations with a finite boundary—layer height, *Boundary—layer Meteorology*, 26, 377—390

Wang Zouqi, Guo Dunren, (1979): Introduce to special functions(in Chinese), Science Press, p152.

Yokoyama, O. M. *et al.*, (1977): On the turbulence quantities in neutral atmospheric boundary layer, *J. Meteorol. Soc. Japan*, 55, 312—318