Airborne Measurements of Turbulent Heat Flux

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Abstract

The results of the turbulent heat flux obtained by airborne measurements over flat terrain and coastal areas are reported, and the structure of the atmospheric boundary layer under the condition of clear daytime is investigated.

The main results are summarized as follows:

(1) In clear daytime, turbulent heat flux q over land decreases linearly with height in the lower layer and q becomes zero at some height, z_q . It then has a negative peak at about a height of $1.3 z_q$ and its absolute value is about 20% of the surface value of q . The height z_q coincides with the height of the inversion base of the potential temperature.

(2) By the analyses of the balance of turbulent energy in the atmospheric boundary layer under clear daytime conditions, we have reached the conclusion that the buoyant energy production balances the dissipation rate of the turbulent energy.

(3) Under sea-breeze conditions, the temperature variation with time can be explained by the heating due to q and the cooling due to the advection of the cool sea-breeze.

(4) The height of the internal boundary layer formed by the thermal convection is investigated and the relation between z_q and x/U (where x is the distance from the coastline along the wind direction and U is mean wind velocity) is obtained.

1. Introduction

The atmospheric boundary layer has important roles not only in the heat and momentum transfer between the surface and the upper atmosphere but also in the dispersion of the air pollutants. However, the structure of the atmospheric boundary layer, particularly its turbulence characteristics, has not yet been completely clarified owing to the lack of appropriate observational means. Remedying such conditions, we developed airborne observation devices and have carried out observations of the turbulence structure in the atmospheric boundary layer over several sites.

The observed data were analysed first to obtain the quantities such as mean temperature and the dissipation rate of kinetic energy to heat which are not greatly affected by the slow fluctuations of the airplane itself. The results have already been reported in Gamo *et al.* (1973, 1976a, 1976b), Yamamoto *et al.* (1974) and Yokoyama *et al.* (1977).

In the next step, we directly analysed the vertical heat flux by using a correction technique developed to compensate for the fluctuations of the airplane using an inertial platform. In this

paper, we investigated the relations among the heat flux, the dissipation rate of kinetic energy to heat and temperature profiles by using the observed data collected in the convective atmospheric boundary layer over relatively flat terrain.

2. Method of observation and data processing

2.1 Instrumentation

The present measurements were carried out by using a light airplane (Cessna-207). The measuring system was composed of sensor for the wind, temperature and airplane motion and a data recorder. Figure 1 shows the positions of the sensors. In Table 1, the sensors for the wind and temperature and measuring items are cited. The airplane motions, *i.e.*, pitching, rolling and yawing angles and vertical, lateral and longitudinal accelerations are measured by the inertial platform system. The performance of this system is cited in Table 2.

2.2 The flight course and the method of data processing

The flight path was chosen to be approximately parallel to the mean wind direction and the flight speed (air-speed) was 120 mph (53.6 m/s). The

Table 1. Sensors for the wind and temperature and measured quantities.

Instrument	Measured quantity	Response
Hot-wire anemometer	longitudinal velocity fluctuation	DC-5 kHz
Sonic anemometer	vertical velocity fluctuation	DC-70 Hz
Sonic thermometer	temperature fluctuation	DC-70 Hz
Thermister thermometer	mean temperature	DC-0.3 Hz
Infrared radiative thermometer	ground and sea surface temperature	DC-10 Hz

Fig. 1 Location of the various sensors; HW (Hot-wire anemometer), SAT (Sonic anemometer and thermometer), T (Thermister thermometer) and IPS (Inertial platform system).

flight time of one run which included 4-5 measurements at different heights was approximately one hour and the length of the flight course at each height was about 30 km.

Data were recorded by magnetic tape recorder. The signals were passed through a low-pass filter (cut-off frequency was 2.5 Hz) and digitized by an A-D converter. The sampling interval was 0.25 sec and sampling duration was about 10 minutes. For cutting the drift of the recording level during measurement, fluctuations longer than 100 sec were cut off by a numerical filter.

2.3 The method of correction for the airplane motion in the turbulence measurements

The method of correction for the fluctuations of the vertical velocity and temperature measured by the sonic anemometer and thermometer is described briefly.

Correction of vertical velocity fluctuations

Vertical velocity fluctuations measured by the sonic anemometer are affected by the airplane motion. The each component of the vertical motion measured by the sonic anemometer and the inertial platform is related as follows;

$$W_C' = W_O' + W_A' + W_P' + W_R', \quad (1)$$

where W_C' : the corrected vertical fluctuation,

W_O' : the vertical fluctuation measured by sonic anemometer,

$W_A' = \int A_z dt$: the vertical motion of the

airplane, where A_z is the acceleration detected by the inertial platform,

$W_P' = -U_A \sin \theta_P$: the vertical speed caused by the pitching of the airplane, where θ_P is the pitching angle and U_A is airspeed measured by the hot-wire anemometer,

$W_R' = -r \dot{\theta}_R$: the vertical speed caused by the rolling of the airplane, where $\dot{\theta}_R$ is the rolling angular velocity and r is the distance between sonic anemometer and gravity center of the airplane (see Figure 1).

The value of W_O' does not contain long period fluctuations. In other words, the airplane can be regarded as a high pass filter for the fluctuations and its cut-off frequency has been found to be approximately 0.2 Hz by analysing the power-spectra of W_O' and W_C' (Yamamoto *et al.*, 1975). On the other hand, in the terms W_A' and W_P' , the long period fluctuations are dominant, so that corrected vertical velocity, W_C' , contains the long period fluctuations. The value of W_R' be-

Table 2. Properties of the inertial platform system.

Measuring device	Measured quantity	Range	Accuracy	Response
Vertical gyro	pitching and rolling angles (θ_P, θ_R)	$\pm 40^\circ$	0.2°	$60^\circ/\text{sec}$
Rate gyro	yawing rate	$\pm 40^\circ/\text{sec}$	$0.1^\circ/\text{sec}$	DC-20 Hz
Accelometers	three-dimensional accelerations (A_x, A_y, A_z)	$\pm 5g$ ($g=9.8 \text{ m/s}^2$)	$10^{-6}g$	DC-250 Hz

comes 0.1 m/s with taking the typical values of $\dot{\theta}_R$ and r , i.e., $\dot{\theta}_R = 0.05$ rad/sec and $r = 2$ m. This value is relatively small to the other terms.

Correction of temperature fluctuations

The temperature observed by the sonic thermometer gives lower value than the true value due to the bending of sound-path at high airspeed. The difference between the observed and the true values is approximately proportional to U_A^2 as shown by Yamamoto *et al.* (1975). The proportional coefficient is determined empirically by using the output-differences of the sonic thermometer for several airspeeds between 40 m/s and 60 m/s at constant height (nearly constant temperature). This proportional coefficient is not affected by the air temperature in the experimental range between 15°C and 25°C.

Computation of turbulent heat flux

The correction of errors due to the airplane motion is important for the calculation of turbulent heat flux. Using the corrected values W_C' and T_C' , the turbulent vertical heat flux q is calculated by

$$q = C_p \rho \overline{W_C' T_C'}, \quad (2)$$

where C_p is the specific heat at constant pressure and ρ is the density of air. The over-bar in equation (2) indicates that q is the mean value over the flight path (about 30 km). When we calculate the horizontal distribution of q , the flight path is divided into several sections and q is computed for each section.

The values of q obtained by the airborne measurements are affected by the both variations in time and space. Nevertheless, the frozen field of turbulence can be assumed because of the airplane speed is about ten-times faster than the wind speed. Therefore, the averaging time can be approximated by U_A/U times measuring interval, where U is the mean wind speed.

In the present analysis, the fluctuations of longer period than 100 sec (corresponds to about 5400 m in space) were cut off by using a numerical filter. Thus, the obtained q in the analysis does not contain contribution of longer than 100 sec.

3. Observation

3.1 Sites of observations

Measurements were carried out over three sites in Japan; the southeast part of the Kanto Plain, the Kashima coastal area in Ibaraki Prefecture

Fig. 2 Map of the Kanto Plain and flight courses.

Fig. 3(a) Map of the Kashima coastal area and flight courses.

Fig. 3(b) Map of the Shibushi coastal area and flight courses.

and the Shibushi coastal area sited in the southern part of Kyushu. The maps and flight courses of these sites are shown Figures 2 and 3(a), (b).

The southern part of the Kanto Plain is nearly flat terrain and the southern part of the flight course is over an urban area as shown in Figure 2. The Kashima and Shibushi coastal areas have a comparatively simple coastline; therefore, these areas are suitable for research on the sea-breeze and internal boundary layer formed by the differences of the surface temperature and surface roughness between sea and land.

3.2 Observational results

The observations over the Kanto Plain were carried out on 28, 29 September, 1974 and 13, 14, 15 August, 1975 as cited in Table 3. The wind direction was nearly *S* and its speed was about 2 m/s–4 m/s, and fair-weather cumulus clouds were observed in almost all cases. Figure 4 shows vertical profiles of q in KAW-1–7. The profiles of q obtained in clear daytime (KAW-3, 5, 6) are positive (upward heat flux) in the lower layer and decrease linearly with height becoming negative at some height, z_q . The values of z_q and the base heights of the potential temperature inversion layer, z_i , are shown in Figure 4. The q profiles still decrease with height above

Table 3. Observations over the Kanto Plain; Wind direction and speed indicate the representative values in the layer up to 500–700 m (data measured by the pilot-balloons except *-estimated from the difference between the ground-speed and the air-speed of airplane).

Run	Date	Time	Wind Direction	Wind Speed	Cloud Species	Cloud Amount
KAW-1	9/28, 1974	15:49–16:45	SE	2–3 m/s	Cu	6/10
2	9/29, 1974	9:53–11:01	—	1–2	As, Cu	8
3		12:31–13:26	SE	2–3	As, Cu	4
4	8/13, 1975	7:05–7:58	—	weak*	Ci, Cu, St	8
5	8/14, 1975	9:48–10:52	SSE	2	—	0
6		15:50–16:55	SE	3–4	Cu	1
7	8/15, 1975	7:08–8:11	~SSW*	~2*	Cu, Sc	4

Table 4. Same as Table 3 except over the Kashima and the Shibushi coastal areas.

Run	Date	Time	Wind Direction	Wind Speed	Cloud Species	Cloud Amount
KAS-1	8/24, 1974	9:29–10:47	SSW	5–8 m/s	—	0/10
2	8/17, 1975	10:01–11:00	SSE	5–15	Sc, Ac	4
3		14:28–15:29	SSE	5–10	Sc, Ac, Cu, Ci	8
4	8/18, 1975	15:28–16:26	SSE	5–10	Sc, Cu, Ci	5
5	8/19, 1975	9:29–10:23	~SSW*	~5–~10*	Sc, Cu, Ci	4
SHI-1	9/9, 1975	9:41–10:52	ESE	3–4 m/s	Cu	2/10
2		12:29–13:36	ESE	3–5	Cu	2
3		15:26–16:12	ESE	4–7	Sc, St	9
4	9/10, 1975	9:40–10:51	E–SE	3–6	—	0
5		12:31–13:41	ESE	2–5	—	0
6		15:30–16:41	ESE	2–4	—	0
7	9/11, 1975	9:48–11:00	WSW	3–7	—	0
8		12:32–13:41	SW	5–7	—	0

Fig. 4 The vertical profiles of q obtained over the Kanto Plain.

Fig. 5 The vertical profiles of q obtained over land of the Kashima and Shibushi coastal areas.

Fig. 6 The vertical profiles of q obtained over sea of the Kashima and Shibushi coastal areas.

Fig. 7 The vertical profiles of q measured by other researchers shown in Table 5.

z_q and have a negative peak at some height.

In the Kashima area, the measurements were made on 24 August, 1974 and 17, 18, 19 August, 1975. Thermal convection was observed only few runs because of the cloudy and strong wind conditions with the exception of run KAS-1.

The measurements over the Shibushi area were made on 9, 10, 11 September, 1975. The weather at the time of the measurements was fair, except the afternoon of the 9th as cited in Table 4. The sea-breeze developed during the daytime on the 9th and 10th, and the wind directions were S-E (3–5 m/s) at heights between the surface and 500 m and W (4–5 m/s) at heights above 500 m. Figure 5 shows the vertical profiles of q obtained by the measurements over the land portions of the Kashima and Shibushi coastal areas. As in Figure 4, the values of q are posi-

tive in the lower layer and they have a negative peak in the upper layer. In Figure 6, the vertical profiles of q measured by the same runs in Figure 5, but over sea, are shown. Generally, the values of q over the sea are small and negative. This result indicates that the stable surface layer is formed over the sea because of the lower sea-surface temperature (refer to Appendices 2 and 3).

For reference, we reviewed the results of the airborne measurements of the turbulent heat flux in clear daytime by other researchers shown in Table 5, and vertical profiles of q are graphed in Figure 7, where each number of profile curves indicates the observation number in Table 5.

Table 5. Observations of turbulent heat flux by other reseachers.

Number	Reseachers	Date	Additional information
1	Koprov and Tsvang (1965)	—	Over the steppe in summer day, developed convection (mean of 8 Runs)
2	Telford and Warner (1964)	12, July, 1962	Developed convection
3		23, July, 1962	
4	Warner (1971)	26, June–13, July, 1966	Over the sea (16.51°S, 145.43°E), Wind 5–9 m/s (mean of 19 Runs)
5	Lenschow and Johnson (1968)	31, March, 1964	Over the forest and slightly rolling terrain
6		8, June, 1964	
7		10, June, 1964	
8	Lenschow (1970)	25, April, 1968	Flat treeless area, slope of the terrain = 0.2%
9	Holmes (1972)	19, July, 1970	Over dry-land forms
10	Myrup (1969)	25, June, 1966	Over dry-lake in the desert, typical convective condition
11	Cattle and Weston (1973)	24, March, 1972	Over the meadow land, gently undulating region
12		16, June, 1972	
13	Milford (1973)	13, July, 1973	Over the land
14	Gamo, Yamamoto and Yokoyama	23, 24, Feb., 1975	Over the sea, convective condition (mean of 5 Runs)
15	Yamamoto, Gamo and Yokoyama	refer to Tables 3 and 4	Over the land, convective condition (mean of 15 Runs)

4. Discussion of the results

4.1 The daily variation of the vertical profiles of turbulent heat flux

The daily variation of q is investigated in clear daytime. The vertical profile of q and its time-variation are directly related to the tem-

perature vertical profile and its time-variation. Figure 8 shows the vertical profiles of q which are the averages of the data over land in clear daytime for each time section of 6–9 h, 9–12 h, 12–15 h and 15–18 h. In the morning time section 1, q was small and negative in the whole layer. This result indicates that the thermal convection did not reach up to the flight height because the heating of the ground surface by the insolation was weak and strong convection did not appear yet. And q in the lower layer increases rapidly with time, and the change from profile 1 to profile 2 seem to be discontinuous. The height, z_q , moves upward with time and becomes 1000 m at the time section 3 and the negative peak of q diminishes gradually. At the time section 4, q is small because the strength of the convection becomes weak with the decrease of the temperature difference between ground surface and air above.

Fig. 8 Time variation of the vertical profile of q over land in clear daytime.

4.2 The vertical profile of q and turbulent energy balance in convective conditions

In the atmospheric surface layer, the heat energy is transported mainly by mechanical turbulence under moderately windy and unstable conditions. This surface layer is then thought to be in the forced convective condition. The

height of this layer is 200 m at most. Above this layer, the mechanical turbulence induced by the roughness of the surface diminishes and the potential temperature gradient is small. In this layer, the heat transport by forced convection becomes negligible, and the thermal convection has a very important role in the vertical heat transport.

The depth of this upper layer may coincide with the average height at which thermal convections diminish, or the height z_q . In Figure 9, the relation between z_q and the heights of the inversion base of potential temperature, z_i , is shown. From this figure, we may regard z_q as approximately coinciding with z_i in spite of the scattered data and the thermal convection are held down by the inversion layer.

Figure 10 is the plot of q normalized by the extrapolated value at ground level, q_0 , against z/z_q . The value q/q_0 decreases linearly with z/z_q between $z/z_q=0$ and 1. The profile of q/q_0 has a negative peak (-0.2) at $z/z_q=1.3$ which is caused by the penetrative thermal convection into the upper inversion layer. If we assume that all of the kinetic energy at $z=z_q$ is converted into potential energy at $z=1.3 z_q$, the ascent velocity of the thermal convection at $z=z_q$ becomes 2.9 m/s for the values of $z_q=700$ m and potential temperature gradient, $d\theta/dz=0.3^\circ\text{C}/100$ m. The values of q/q_0 are small and negative above 1.3 z_q because of the downward heat transport by the kinetic turbulence in the upper stable layer.

The above, analysis was made under the implicit assumption that the atmospheric boundary layer is in a free convective condition in clear daytime. We will investigate this assump-

Fig. 9 Relation between z_q and z_i under the convective condition.

Fig. 10 The vertical profile of q/q_0 against z/z_q under the convective condition.

tion concerning the balance of the turbulent energy. The equation of the turbulent energy for the horizontally uniform and stationary atmospheric boundary layer is

$$\varepsilon = \frac{g}{T} \overline{W'T'} - \frac{\partial}{\partial z} \left(\overline{W'e'} + \frac{\overline{W'P'}}{\rho} \right) + \frac{\tau}{\rho} \frac{\partial V}{\partial z} \quad (3)$$

where ε is the dissipation rate of turbulent energy, e' is the turbulent total energy, P' is the fluctuation of pressure, τ is the shear stress, and V is the wind vector. The pressure fluctuation term was not measured but it may be negligible except at the top of the atmospheric boundary layer.

Figure 11 shows the relation between ε and $g/T \overline{W'T'}$ under clear daytime conditions and in Figure 12, the variation of $g/T \overline{W'T'}/\varepsilon$ with height z/z_q is shown. From these figures, it can be concluded that the equation, $\varepsilon = g/T \overline{W'T'}$, holds in the atmospheric boundary layer except in the surface layer and near the height of z_q and, in this layer, free convection is predominant and the production of the turbulent energy due to the wind shear and the vertical diffusion of turbulent energy is relatively small.

4.3 The horizontal distribution of q under sea-breeze conditions

In the former sections, only the vertical profiles of q are dealt with, but the horizontal distributions of q and temperature are important for the internal boundary layer formed by the differences in the roughness and temperature of the surface between the sea and land. Figure 13

Fig. 11 Relation between ϵ and $g/T W'T'$ under the convective condition.

Fig. 12 The vertical profile of $g/T W'T'/\epsilon$ against z/z_q under the convective condition.

shows examples of the horizontal distributions of q and T of the 10th of September, 1975 at the Shibushi coastal area. The wind direction was SE in the lower layer and changed to SW above 1000 m as shown in the left graph of Figure 13. In the morning (SHI-4), the values of q are positive at the heights of 300 and 500 m over land and negative over the sea. The layer of positive q extends with time and q becomes positive at $z=700$ m in the noon (SHI-5). The values of q at some height and time increase with the axis extending into the inner land. Corresponding to this horizontal distribution of q , air temperature over land is 0.3–0.7°C higher than it over sea.

Fig. 13 An example of the horizontal distributions of heat flux and temperature on 10th September, 1975 over the Shibushi coastal area.

In these cases, the time variation of the air temperature depends on the heating by q and cooling due to the cool sea-breeze. This can be described by the following relation,

$$\frac{\partial T}{\partial t} = -U \frac{\partial T}{\partial x} - \frac{\partial}{\partial z} \left(\frac{q}{C_p \rho} \right), \quad (4)$$

where t is time, U is wind and x is the horizontal distance from the coastline along the wind direction. From equation (4), we can estimate the value of q by using the data of the observed temperature and wind and integrating with respect to z . The result is shown in Figure 14. The

Fig. 14 An example of vertical profiles of q estimated by equation (4).

Fig. 15 Mean horizontal distributions of q under sea-breeze conditions over the Shibushi coastal area.

estimated q at a middle time between the two measurements approximately coincide with the observed value of q .

Now, we will investigate the relation between the depth of the internal boundary layer and the distance from the coastline along the wind direction. Figure 15 shows the mean horizontal distributions of q obtained from the data at the heights of 300 and 500 m under sea-breeze conditions on the 9th and 10th of September, 1975 at the Shibushi coastal area. The heights at which q becomes zero, z_q , are determined by linear interpolation from the horizontal distribution of q at 300 and 500 m for every two kilometers along x . The height z_q can be considered to be the depth of the internal boundary layer formed by thermal convection.

Assuming the invariance of q_0 and potential temperature gradient in the upper layer with time and neglecting the temperature gap of the upper part of the internal boundary layer, the height z_q is proportional to the square root of the time, t (e.g., Panofsky, 1973). Substituting x/U to t ,

$$z_q \propto \left(\frac{x}{U} \right)^{0.5} \quad (5)$$

is obtained, where U is the wind speed at $z =$

Fig. 16 Relation between z_q and x/U under sea-breeze conditions over the Shibushi coastal area.

100 m.

Figure 16 is a graph of the relation between z_q and x/U obtained by the measurements. The dotted line in Figure 16 indicates the relation $z_q \propto (x/U)^{0.5}$. Near the coastline, z_q increases suddenly, but the starting point of the internal boundary layer can not be clearly determined because of insufficient data.

5. Conclusions

The results of the turbulent heat flux obtained by the airborne measurements and the structure of the atmospheric boundary layer are investigated.

In clear daytime, q over the land decreases approximately linearly with height in the lower layer and q becomes zero at some height z_q . It has a negative peak at about height of $1.3 z_q$ and its absolute value is about 20% of q_0 . The height z_q coincides with the height of the inversion base of the potential temperature.

We analysed the balance of the turbulent energy in the atmospheric boundary layer under clear daytime conditions and reached the conclusion that $g/T \overline{w'T'}$ was balanced with ϵ in the atmospheric boundary layer except in the surface layer and near the height z_q .

By the analysis of the horizontal distributions of q and T , the temperature variation with time can be explained by the heating due to q and cooling due to the advection of the cool sea-breeze. We determined the height of the internal boundary layer formed by the thermal convections using the horizontal distributions of q and obtained the result that z_q is approximately proportional to $(x/U)^{0.5}$.

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Appendix 1. Data obtained by the KAW-1-7 Runs; Wind: pilot-balloon measurement, T : air temperature, q : turbulent heat flux, ε : energy dissipation rate, T_s : land (L) and sea (S) surface temperature, Z_i : the base height of the potential temperature inversion layer, Z_q : the height of $q=0$.

Run	Height (m)	Wind (m/s)	T (°C)	q (cal/sm ²)	ε (m ² /s ³ × 10 ⁻⁴)	T_s (°C)	Z_i (m)	Z_q (m)
KAW-1	1,500	2 WNW	16.4	-3.1	0.14	$T_s = -$		
	1,000	1 E	19.9	0.3	0.85	$Z_i = -$		
	500	3 SE	21.2	-2.5	0.51	$Z_q = 200$		
	300	2 SE	23.1	-0.9	0.57			
KAW-2	1,500	1.5S	11.1	-1.6	0.31	$T_s = 31.2$		
	1,000	2 SE	13.2	-2.5	0.46	$Z_i = 500$		
	700	1.5ESE	15.4	—	4.2	$Z_q = 500$		
	500	1.5E	16.7	0.0	12.9			
	300	2 NNE	19.0	5.6	16.5			
KAW-3	1,500	3.5SW	12.3	-2.5	0.16	$T_s = 29.5$		
	1,000	3 SE	15.2	-3.7	0.33	$Z_i = 1,000$		
	500	3 ESE	20.7	5.9	5.2	$Z_q = 750$		
	300	2.5ESE	23.2	10.6	8.5			
KAW-4	1,500	—	18.9	-0.6	0.08	$T_s = 24.6$		
	1,000	—	20.4	-0.3	0.01	$Z_i = -$		
	700	—	21.5	0.0	0.04	$Z_q = -$		
	500	—	22.6	-1.6	0.33			
KAW-5	1,500	2 SSE	19.2	-0.6	0.05	$T_s = 39.0$		
	700	2 SSE	23.2	11.2	13.2	$Z_i = 700$		
	500	2 SSE	25.4	15.2	13.7	$Z_q = 1,250$		
	300	1 SSE	25.8	—	—			
KAW-6	1,500	1 E	19.7	-0.6	0.05	$T_s = 34.5$		
	1,000	2.5SE	22.4	-3.4	1.6	$Z_i = -$		
	500	4 SE	26.2	3.4	3.7	$Z_q = -$		
KAW-7	1,500	—	18.1	-1.6	0.42	$T_s = 29.0$		
	1,000	—	20.6	-0.9	0.05	$Z_i = 500$		
	700	—	22.9	-0.6	0.05	$Z_q = 400$		
	500	—	23.5	-1.2	0.90			
	300	—	24.1	1.6	0.48			

Appendix 2. Data obtained by the KAS-1-5 Runs.

Run	Height (m)	Wind (m/s)	T (°C)		q (cal/sm ²)		ϵ (m ² /s ³ × 10 ⁻⁴)		T_s (°C)	Z_i (m)	Z_q (m)
			L	S	L	S	L	S			
KAS-1	1,500	8 SW	19.9	19.9	-2.5	0.9	0.01	0.01	$T_s(L)=34.3$		
	1,000	9 SW	21.8	22.1	-2.5	-2.2	0.05	0.19	$T_s(S)=24.6$		
	700	7 SW	22.8	22.9	-1.2	-3.1	0.45	2.1	$Z_i=300$		
	500	5 SSW	24.3	24.2	-5.0	-0.3	4.8	0.68	$Z_q=350$		
	300	6 S	24.5	24.0	3.4	-2.2	7.6	1.2			
	200	5 SSW	—	—	32.6	-1.6	11.4	0.70			
KAS-2	1,500	—	18.6	18.3	—	-0.6	—	0.02	$T_s(L)=34.0$		
	1,000	6 S	21.2	21.0	-0.3	—	0.05	—	$T_s(S)=23.3$		
	700	17 SSE	23.1	23.2	1.2	-0.9	0.09	0.03	$Z_i=500$		
	500	15 SSE	24.4	24.5	-1.6	-1.6	0.02	0.15	$Z_q=400$		
	300	10 SSE	26.4	26.5	4.3	—	4.3	0.02			
	200	7 SSE	—	27.3	—	-2.2	—	—			
	100	6 SSE	—	27.8	—	-0.9	—	0.01			
KAS-3	1,000	9 SSE	21.6	21.9	0.0	-3.1	0.02	0.07	$T_s(L)=35.3$		
	700	9 SSE	23.6	23.9	-0.6	-0.9	0.04	0.02	$T_s(S)=23.9$		
	500	8 SSE	24.9	24.9	0.9	-0.6	0.10	0.03	$Z_i=—$		
	300	6 S	26.2	26.4	—	0.0	—	0.03	$Z_q=—$		
	200	6 SSE	—	27.4	—	-0.3	—	0.03			
	100	5.5SSE	—	27.8	—	-0.6	—	0.01			
KAS-4	1,500	—	17.8	18.0	1.2	-0.6	0.07	0.02	$T_s(L)=31.5$		
	1,000	9 SSE	22.2	21.5	2.5	-0.9	0.01	0.13	$T_s(S)=23.2$		
	700	10 S	24.1	23.8	-0.3	0.0	0.03	0.09	$Z_i=—$		
	500	7 SSE	24.9	24.9	-0.6	—	0.01	0.05	$Z_q=—$		
	300	6 SSE	26.9	27.0	-1.2	-1.2	0.50	0.01			
	200	8 SSE	—	27.8	—	0.6	—	0.04			
	100	10 SSE	—	28.3	—	0.0	—	0.03			
	50	—	—	28.3	—	-0.3	—	0.04			
KAS-5	1,500	—	20.9	20.1	—	—	—	0.05	$T_s(L)=35.4$		
	900	—	22.9	22.2	-4.3	—	0.65	1.1	$T_s(S)=22.3$		
	500	—	24.4	24.4	8.1	-5.3	7.3	2.1	$Z_i=300$		
	300	—	26.1	26.6	14.9	-6.8	11.7	0.48	$Z_q=650$		
	200	—	27.4	27.2	44.1	-6.2	8.6	2.3			
	150	—	—	27.9	—	-5.9	—	0.6			
	50	—	—	29.1	—	-2.5	—	1.0			

Appendix 3. Data obtained by the SHI-1-8 Runs.

Run	Height (m)	Wind (m/s)	T (°C)		q (cal/cm ²)		ϵ (m ² /s ³ × 10 ⁻⁴)		T_s (°C)	Z_i (m)	Z_q (m)
			L	S	L	S	L	S			
SHI-1	1,500	4 WNW	17.8	17.5	-1.9	—	0.12	0.05	$T_s(L)=—$		
	1,000	3.5SW	20.5	20.6	0.3	—	0.09	—	$T_s(S)=—$		
	700	1 SW	22.7	22.9	-0.9	-1.2	1.1	0.13	$Z_i=500$		
	500	3 ESE	24.2	24.1	5.0	-6.2	7.5	0.33	$Z_q=600$		
	300	3 E	25.7	25.5	3.7	-3.7	3.3	0.78			
	200	2.5ESE	—	26.7	—	0.0	—	1.0			
SHI-2	1,500	4.5NW	18.5	18.8	2.8	—	0.16	0.40	$T_s(L)=32.1$		
	1,000	2 WNW	21.5	21.2	2.5	1.9	0.28	—	$T_s(S)=27.3$		
	500	3 ESE	24.9	24.9	8.7	-2.2	9.6	1.1	$Z_i=500$		
	300	5 ESE	26.6	26.4	1.9	0.0	5.5	1.8	$Z_q=—$		
	200	5.5ESE	—	27.2	—	—	—	1.5			
SHI-3	1,000	2 SW	21.0	20.8	-4.0	—	0.22	—	$T_s(L)=28.0$		
	700	1 SSE	23.4	23.1	-1.6	0.6	0.03	0.46	$T_s(S)=27.3$		
	500	6.5SE	24.6	24.5	1.9	-1.9	1.7	0.27	$Z_i=500$		
	300	7 ESE	26.1	26.1	4.7	-1.9	2.4	1.4	$Z_q=600$		
	200	6 ESE	—	27.1	—	-2.5	—	0.98			
SHI-4	1,500	3 SW	18.1	18.3	—	-1.6	0.12	0.12	$T_s(L)=31.3$		
	1,000	2 SSW	21.1	21.3	0.0	0.0	0.10	0.03	$T_s(S)=27.1$		
	700	3 SE	23.0	23.2	-3.1	0.0	0.25	0.12	$Z_i=300$		
	500	6 ESE	24.7	24.7	10.9	-2.8	4.2	0.02	$Z_q=600$		
	300	5 ESE	25.7	25.5	16.8	-0.6	2.8	0.01			
	200	4 E	—	26.4	—	-0.6	—	0.10			
SHI-5	1,500	4 WSW	17.7	17.2	-1.6	-1.9	0.05	0.12	$T_s(L)=32.3$		
	1,000	1 SSW	21.3	21.4	-0.9	—	0.07	—	$T_s(S)=27.2$		
	700	2 SE	23.5	23.4	-2.5	-0.3	0.41	0.73	$Z_i=300$		
	500	1.5SE	25.0	24.6	4.6	2.2	4.8	0.71	$Z_q=—$		
	300	6 ESE	26.4	25.9	7.5	0.9	8.4	0.77			
	100	4 ESE	—	27.8	—	—	—	—			
SHI-6	1,500	3.5WNW	18.2	18.3	-3.4	-0.3	0.10	0.07	$T_s(L)=28.5$		
	1,000	2 SW	21.2	21.4	-2.8	2.2	0.14	0.07	$T_s(S)=27.3$		
	700	2 SSE	24.2	24.3	-0.3	-0.6	0.39	0.62	$Z_i=—$		
	500	2 ESE	25.7	26.1	0.3	-3.4	1.5	0.21	$Z_q=—$		
	300	3.5ESE	27.0	26.7	—	-0.6	—	0.39			
	200	4 SE	—	27.8	—	0.0	—	0.35			
SHI-7	1,500	—	19.5	19.4	-0.9	0.9	0.09	0.36	$T_s(L)=33.1$		
	1,000	8 SW	21.3	21.0	0.0	—	1.3	—	$T_s(S)=27.3$		
	700	6 WSW	22.8	23.1	-4.0	0.0	2.9	1.8	$Z_i=700$		
	500	6 WSW	24.8	24.8	7.8	—	6.4	—	$Z_q=650$		
	300	4 WSW	26.7	26.2	15.5	-2.8	5.8	2.3			
	200	2.5SSW	—	27.6	—	-0.6	—	10.8			
SHI-8	1,000	2 SSW	—	28.4	—	-0.6	—	4.8			
	1,500	6 WSW	21.1	20.9	-2.2	1.6	0.08	0.40	$T_s(L)=36.5$		
	700	6.5SW	24.2	24.4	0.9	-1.9	4.7	0.87	$T_s(S)=27.3$		
	500	6 SW	26.3	26.2	3.4	—	5.5	—	$Z_i=700$		
	300	4.5SW	28.3	27.8	5.3	0.0	5.9	1.5	$Z_q=750$		
	200	6 WSW	—	28.6	—	1.6	—	4.8			
	100	4.5WSW	—	29.2	—	-0.3	—	2.7			

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飛行機による乱流熱輸送量の観測について

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平坦地および海岸地帯上空での飛行機観測による乱流熱輸送量 q の測定結果について述べ、それにもとづいて主に晴天時日中の大気境界層の状態について考察した。

ここで得られた主要な結論は

(1) 晴天時日中の q は下層で高度 z とともに直線的に減少し、ある高度 z_q で $q=0$ となりさらに上層の $z=1.3z_q$ 近傍で負のピークをもつ。そのピーク値の絶対値は地上での値の 20% 程度である。また z_q と温位の逆転層底部高度 z_i はほぼ一致する。

(2) 晴天時日中では接地気層と大気境界層上端部をのぞく高度において乱流運動エネルギーのバランスは粘性による運動エネルギー消散項と浮力からの運動エネルギー生成項の間に成立している。

(3) 海風発生時の q と気温の水平分布および時間変化について考察し、気温の時間変化は q による加熱と海風による冷気の流入による冷却により表わされることがわかった。また熱対流により形成される内部境界層の厚さを示す z_q と吹流時間 (x/U) の間の関係を調べ、 z_q は $(x/U)^{0.5}$ に比例するという結果を得た。