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Observation Research of the Turbulent Fluxes of Momentum, Sensible Heat and Latent Heat over the West Pacific Tropical Ocean Area

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ABSTRACT

This paper describes results of the fluxes of momentum, sensible heat and latent heat for the West Pacific Tropical Ocean Area (127° E – 150° E, 5° N –3° S). The data were collected by the small tethered balloon sounding system over this ocean area including 6 continuous stations (140° E, 0°; 145° E, 0°; 150° E, 0°; 140° E, 5° N; 145° E, 5° N and 150° E, 5° N) from 11 October to 15 December, 1986. These fluxes were calculated by the semiempirical flux–profile relationships of Monin–Obukhov similarity theory using these observed data. The results show that for this tropical ocean area the drag coefficient C_D is equal to $(1.53 \pm 0.25) \times 10^{-1}$ and the daily mean latent flux H_l is greater than its daily mean sensible flux H_s by a factor of about 9.

1. INTRODUCTION

In the recent ten odd years the scientists of many countries over the world have been interested in the climatic anomaly that is characterized by El Nino and Southern Oscillation phenomena. The research of the air–sea exchange, the air–sea interaction and teleconnection of some physical factors is the key to this problem. The Chinese meteorologists think that the West Pacific Tropical Ocean Area is a stronger heat source on the earth. Some severe weather processes are produced by it. This ocean area continuously absorbs solar radiation. There are three ways, i.e. the vertical exchange of the small–scale turbulent motion, the meso–scale weather processes and the large–scale severe weather processes, to transfer the stored heat energy in this ocean area to the affected region. Many observation researches have been made of the turbulent fluxes of momentum, sensible heat and latent heat over the land. But it is rare to research vertical exchange of turbulent fluxes of momentum, sensible heat and latent heat over the water, especially over the ocean. Among them AMTEX, BOMEX, MASEX, JASINEX, FASINEX and GATE are some famous air–sea exchange experiments. From 11 October to 15 December, 1986, the science observation ship "Shi Yan 3" of Academia Sinica completed the extensive comprehensive investigation of meteorology, oceanology, and air–sea interactions over the West Pacific Tropical Ocean Area (127° E–150° E,5° N–3° S,see Fig.1). The data were collected by the tethered balloon sounding system (See Zhou et al., 1984; Qu et al., 1984) over this ocean area including 6 continuous stations (i.e. 140° E, 0°;

[1] An Leiming and Sui Dong took part in observational work.

145° E, 0°; 150° E, 0°; 140° E, 5° N; 145° E, 5° N and 150° E, 5° N).The profile data of wind, air temperature, humidity, and pressure amount to 30 runs. The results of the fluxes of momentum (that is the drag coefficient), sensible heat,and latent heat were mainly based on these profile data by the semiempirical flux–profile relationships of Monin–Obukhov similarity theory (see Businger et al; 1971). Table 1 summarizes these flux results. Besides, the stability parameter z/L, air temperature, humidity, mean wind speed, Bowen ratio β etc. are given in Table 1. These observation results were compared with those obtained during AMTEX (Japan) and BOMEX (U.S.A.).

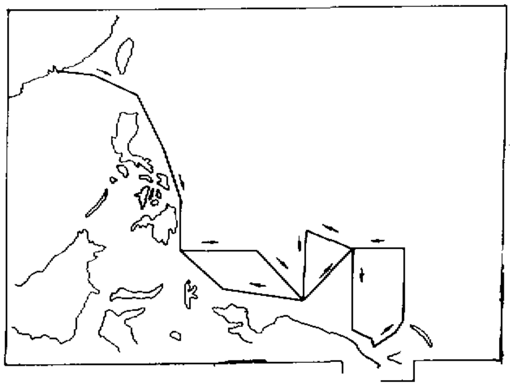

Fig 1. Surveying line of 'Shi Yan 3' ship.

II. OBSERVATION INSTRUMENTS AND METHODS OF DETERMINING FLUXES

A small tethered balloon sounding system consisting of a blimp–shaped balloon, airborne telemetry package, ground receiver,data processing equipment and releaser, was used for sampling air temperature, humidity, pressure and wind every 6 m in the atmospheric boundary layer. Every observation took about fifty or sixty minutes. The mean profile data were obtained in this period. There is one observation every three hours. Therefore we obtained variation data of the turbulent fluxes.

It is well known that the methods of determining turbulent fluxes cover (a) eddy correlation method, (b) energy dissipation method, (c) bulk aerodynamic method, and (d) mean profile method. It is clear that the turbulent fluxes of momentum, sensible heat, moisture, and latent heat can be written as

Table 1. The Observational Pesults of Turbulent Fluxes for the West Pacific Tropical Ocean Area

No.	L.(°E)	Latd.(°N)	Time	T_a(°C)	T_d(°C)	h(%)	U(ms^{-1})	Ri	ζ	u_*^2(cm · s^{-1})	u_*(cm^2 · s^{-2})	θ_*(°C)	10^3C_D	H(w · m^{-2})	β
1	140	0	14 : 55–15 : 46, 31 / 10	31.4	30.5	66	4.7	−0.10	−0.10	20	400	−0.07	1.81	16	
2	140	0	20 : 07–21 : 07, 31 / 10	30.2	28.8	74	4.9	−0.20	−0.25	20	400	−0.09	1.67	21	
3	140	0	23 : 50–00 : 30, 31 / 10	29.9	28.5	76	5.4	−0.18	−0.22	22	484	−0.09	1.66	22	
4	140	0	03 : 56–04 : 38, 1 / 11	29.8	28.3	73	5.3	0.16	0.20	18	324	−0.07	1.15	19	
5	140	0	08 : 07–08 : 52, 1 / 11	29.8	29.6	62	8.2	0.10	0.10	33	1089	0.05	1.62	19	
6	145	0	12 : 14–12 : 54, 5 / 11	30.6	29.3	62	5.8	−0.10	−0.10	17	289	0.06	0.86	12	
7	145	0	17 : 25–18 : 05, 5 / 11	30.1	27.9	67	7.4	−0.16	−0.20	28	784	−0.05	1.43	15	
8	145	0	23 : 25–24 : 00, 5 / 11	29.8	27.6	68	9.1	−0.14	−0.18	35	1225	−0.06	1.48	20	
9	145	0	02 : 13–02 : 48, 6 / 11	29.8	27.8	65	8.2	−0.10	−0.10	28	784	−0.04	1.17	17	
10	145	0	06 : 15–06 : 30, 6 / 11	30.0	25.5	74	7.5	−0.10	−0.10	28	784	−0.05	1.39	18	
11	150	0	22 : 51–23 : 28, 17 / 11	29.2	28.4	66	8.1	0.20	0.25	34	1156	0.07	1.76	27	
12	150	0	02 : 56–03 : 38, 18 / 11	28.9	28.1	66	7.5	−0.18	−0.22	34	1156	−0.06	2.06	21	
13	150	0	08 : 05–08 : 37, 18 / 11	29.4	29.3	63	9.6	−0.16	−0.20	35	1225	−0.05	1.34	20	
14	150	0	11 : 42–12 : 14, 18 / 11	29.3	28.9	76	8.2	−0.14	−0.18	31	961	−0.05	1.43	17	
15	150	0	21 : 45–22 : 21, 18 / 11	29.1	26.8	70	7.0	−0.18	−0.22	24	576	−0.08	1.18	22	
16	150	0	05 : 35–06 : 00, 19 / 11	29.1	26.1	98	9.6	0.16	−0.20	40	1600	−0.07	1.74	24	
17	150	0	10 : 20–10 : 50, 19 / 11	29.6	28.2	82	10.5	0.18	0.22	39	1521	0.05	1.38	20	
18	150	5	12 : 53–13 : 42, 21 / 11	29.4	28.3	65	5.4	−0.16	−0.20	22	484	−0.06	1.66	16	0.07
19	150	5	17 : 04–17 : 48, 21 / 11	29.4	27.5	83	5.6	−0.18	−0.22	23	529	−0.07	1.69	18	
20	150	5	21 : 07–21 : 41, 21 / 11	29.4	27.3	69	5.8	−0.23	−0.30	24	576	−0.08	1.71	21	
21	150	5	01 : 03–01 : 43, 22 / 11	29.4	28.0	67	4.4	−0.18	−0.22	16	256	−0.09	1.32	15	
22	150	5	05 : 37–06 : 07, 22 / 11	29.4	28.3	79	5.6	−0.20	−0.25	22	484	−0.06	1.54	17	
23	145	5	13 : 17–14 : 28, 23 / 11	30.1	29.7	72	7.4	0.14	0.18	30	900	0.05	1.64	18	
24	145	5	17 : 01–17 : 44, 23 / 11	29.9	28.5	78	8.8	−0.14	−0.18	33	1089	−0.08	1.41	19	
25	145	5	20 : 57–21 : 37, 23 / 11	29.8	28.2	79	6.7	−0.18	−0.22	27	729	−0.07	1.62	24	
26	145	5	00 : 54–01 : 40, 24 / 11	29.7	28.0	68	6.3	−0.16	−0.20	24	576	−0.06	1.45	19	
27	145	5	05 : 31–06 : 18, 24 / 11	29.6	28.5	78	4.0	−0.10	−0.10	18	324	−0.08	2.02	14	
28	140	5	12 : 11–12 : 53, 26 / 11	29.7	28.8	65	6.7	−0.20	−0.25	25	625	−0.06	1.39	16	
29	140	5	16 : 23–17 : 03, 26 / 11	29.4	28.6	76	6.1	0.20	0.25	25	625	0.05	1.68	15	
30	140	5	20 : 24–21 : 04, 26 / 11	29.3	28.1	68	6.0	0.23	0.30	24	576	−0.09	1.60	20	

$$M=\rho\overline{w'u'}=-\rho u_*^2=-\rho K_M\frac{\partial\bar{u}}{\partial z},$$
$$H_s=c_p\rho\overline{w'\theta'}=-c_p\rho u_*\theta_*=-c_p\rho K_H\frac{\partial\bar{\theta}}{\partial z}, \tag{1}$$
$$E=\rho\overline{w'q'}=-\rho u_*q_*=-\rho K_q\frac{\partial\bar{q}}{\partial z},$$

where the bar and the prime over a quantity indicate averaging and fluctuation values of the quantity respectively. ρ is the air density, c_p the specific heat at constant pressure (for the dry air $C=1.005\text{J}\cdot\text{g}^{-1}\cdot\text{K}^{-1}$), u_* the friction velocity ($u_*^2=-\overline{u'w'}=\tau\cdot\rho^{-1}$, τ the surface stress), θ_* the temperature scale ($\theta_*\equiv-\overline{\theta'w'}\cdot u_*^{-1}$), q_* the characteristic specific humidity ($q_*\equiv-\overline{q'w'}\cdot u_*^{-1}$). u,θ and q are the instantaneous values of the wind speed, the potential temperature and the specific humidity respectively. K_M,K_H and K_q are the eddy diffusivities of momentum, heat and moisture respectively.

For the latent heat flux of vaporization it is written in form:

$$H_l=-L\rho K_q\frac{\partial\bar{q}}{\partial z}, \tag{2}$$

where L is the latent heat of vaporization and it is the function of air temperature, and takes 2510 $\text{J}\cdot\text{g}^{-1}$ at 0 °C . The ratio of sensible heat flux to latent heat flux may be written in the form (i. e. Bowen ratio) approximately:

$$\beta=\frac{H_s}{H_L}\approx\frac{c_p}{L}\frac{\Delta\bar{\theta}}{\Delta\bar{q}}. \tag{3}$$

Here $K_H=K_q$ is assumed.

According to the semiempirical flux–profile relationships of Monin–Obukhov theory, the wind shear, the temperature gradient and specific humidity gradient as functions of dimensionless height are written in the forms

$$\varphi_M(\zeta)=\frac{\kappa z}{u_*}\frac{\partial\bar{u}}{\partial z},$$
$$\varphi_H(\zeta)=\frac{\kappa z}{\theta_*}\frac{\partial\bar{\theta}}{\partial z}, \tag{4}$$
$$\varphi_E(\zeta)=\frac{\kappa z}{q_*}\frac{\partial\bar{q}}{\partial z},$$

where $\zeta=z/L, L=-Tu_*^3/g\overline{w'T'}\kappa$ is the Minin–Obukhov length, for $\zeta>0$ it represents stable atmosphere and for $\zeta<0$ unstable atmosphere. κ is a von Karman constant, here we take $\kappa=0.35$.

Using the observed data of Air Force Cambridge Laboratory,Businger (1971) gives the real forms of φ_M, φ_H and φ_E

$$\varphi_M=\begin{cases}1+\beta_m\frac{z}{L}, & (\frac{z}{L}\geqslant 0)\\ (1-\gamma_m\frac{z}{L})^{-\frac{1}{4}}, & (\frac{z}{L}\leqslant 0)\end{cases} \tag{5}$$

$$\varphi_H=\varphi_E=\begin{cases}1+\beta_n\frac{z}{L}, & (\frac{z}{L}\geqslant 0)\\ (1-\gamma_n\frac{z}{L})^{-\frac{1}{2}}, & (\frac{z}{L}\leqslant 0)\end{cases}. \tag{6}$$

where the constants $\gamma_m=15$, $\beta_m=4.7$, $\gamma_n=9$, and $\beta_n=6.4$.

The integration of Eqs. (5) and (6) leads to the explicit expressions of wind profile and temperature profile (Paulson , 1970).

for $z/L<0$.

$$\frac{\bar{u}}{u_*}=\frac{1}{\kappa}(\ln\frac{z}{z_0}-\psi_1), \tag{7}$$

where

$$\psi_1=2\ln(\frac{1+x}{2})+\ln(\frac{1+x^2}{2})-2\mathrm{tg}^{-1}x+\frac{\pi}{2},$$

$$x=(1-15\zeta)^{\frac{1}{4}}=\varphi_M^{-1},$$

for $z/L>0$

$$\frac{\bar{u}}{u_*}=\frac{1}{\kappa}(\ln\frac{z}{z_0}+4.7\zeta), \tag{8}$$

and for temperature profile:

for $z/L<0$

$$\frac{\bar{\theta}-\theta_0}{\theta_*}=0.74(\ln\frac{z}{z_0}-\psi_2), \tag{9}$$

where

$$\psi_2=2\ln(\frac{1+y}{2}),$$

$$y=(1-9\zeta)^{\frac{1}{2}}=\varphi_H^{-1},$$

for $z/L>0$

$$\frac{\bar{\theta}-\theta_0}{\theta_*}=0.74\ln\frac{z}{z_0}+4.7\zeta. \tag{10}$$

Here θ_0 is the extrapolation temperature at $z=0$ (It may not be the true surface temperature).

Therefore based on these data obtained by the small tethered balloon sounding system over the West Pacific Tropical Ocean Area for a period of time and using the least square method we have acquired mean wind profile and mean temperature profile. We use these mean profiles to fit formulas(7)–(10). On the basis of mean profilies we calculate Ri value $[\equiv(g/\bar{\theta})(\partial\bar{\theta}/\partial z)/(\partial\bar{u}/\partial z)^2]$ at a fixed level, for example, 10 m. From Ri–ζ curve (Businger et al., 1971) the corresponding ζ values are found. Based on the formulas (7)–(10) the u_* and θ_* values for different conditions are calculated. Finally according to the formulas $H_s=-c_p\rho u_*\theta_*$ and $M=-\rho u_*^2$ we easily obtain the sensible heat flux

H_s , the momentum flux M (or the drag coefficent C_D) and the latent heat flux H_L.

III. SOME RESULTS

Table I gives the daily variations of the turbulent vertical fluxes obtained by the above mentioned computational method, corresponding to atmospheric stratification values and routine meteorological observational data for six continuous stations (140 ° E, 0 ° ; 145 ° E, 0 ° ; 150 ° E, 0 ° ; 140 ° E, 5 ° N,145 ° E, 5 ° N; 150 ° E, 5 ° N) in the investigated tropical ocean area. Based on these results we will discuss the transfer properties of the turbulent fluxes of momentum, sensible heat and latent heat for this tropical ocean area.

1. *Momentum Flux*

The process through which momentum flux was transferred to over and in ocean, is an important part of air–sea interactions. The sea wave, ocean current, and ocean swell, break-down of sea wave, and wind storm tide etc. are the direct response to the above process. Dynamic models of air–sea interactions usually have available only bulk weather data for initial input parameters. Consequently the bulk aerodynamic formula $\tau = \rho C_D U^2$ was introduced, where C_D is the drag coefficient. It is important for us to study air–sea interactions and air–sea interchanges. Besides, it is necessary for remote sensing of ocean wave and wind field over ocean area by satellite. The dependence of the drag coefficent on the atmospheric stratification and wind speed was studied in some detail over land. The dependence of the drag coefficient on wave state has also received considerable attention in recent years. Geerñaert (1987) summarized some regression equations of the dependence of the drag coefficient on wind speed for the neutral atmosphere up to 1986. On the basis of some observational results Businger (1973) gave the drag coefficient formulas for the stable and unstable atmospheres. In the recent years both theory and observation have some works for the drag coefficient over ocean area.

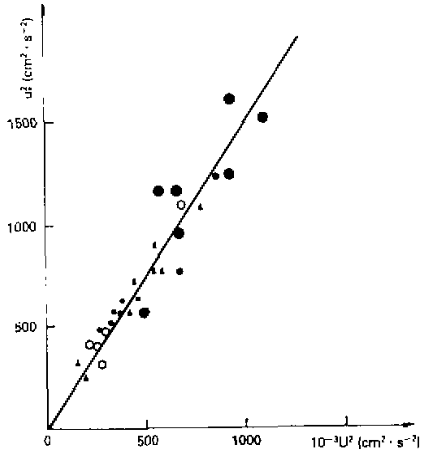

Fig.2. u_*^2 vs. $10^{-3}U^2$. The straight line corresponds to $C = 1.53 \times 10^{-3}$.

Fig.2 shows u_*^2 determined by mean profile method as a fuction of $10^{-3}U^2$ for the West Pacific Tropical Ocean Area. The straight line in Fig.2 corresponds to $C_D = 1.53 \times 10^{-3}$. This result [$C_D = (1.53 \pm 0.25) \times 10^{-3}$ by mean profile method] seems a bit large as compared with Hasse's (1970) result [$C_D = (1.21 \pm 0.24) \times 10^{-3}$] and Miyake's (1970) result [$C_D = (1.10 \pm 0.24) \times 10^{-3}$]. As a matter of fact it may be so. Hasse's and Miyake' sresults are obtained for slightly unstable conditions. It is well known that the drag coefficient C_D has a tendency to increase with increase of unstability. The unstability for our observational period is stronger than that for their observational period. But our result is consistent with Pond's (1971) results [$C_D = (1.52 \times 0.40) \times 10^{-3}$by eddy correlation method and $C_D = (1.55 \pm 0.40) \times 10^{-3}$] by dissipation method obtained near Barbados island. The above results are obviously believable because their observational conditions (over ocean, unstable atmosphere i.e. $\zeta = z / L < 0$, and variation range of the stability parameters Ri or ζ and mean speed) are basically similar. From our observational results it is seen that over the West Pacific Tropical Ocean Area in the same condition of the stability parameter Ri or ζ, the correlation dependence of the drag coefficient C_D on the mean wind speeds U is better. In the same U conditions the correlation dependence of the drag coefficient C_D on stability parameters Ri or ζ is also better (figure omitted). If wind speeds are different (or for different ζ values in unstable conditions), the correlation dependence of the drag coefficient C_D on ζ (or on the mean wind speed U) is not obvious (figure omitted).

2. *Sensible Heat Flux and Latent Heat Flux*

On the other hand, in air–sea interactions, as in day and night all the year around the sea surface temperature is almost larger than the air temperature for the observed West Pacific Tropical Ocean Area (Qu,to be published) , the solar radiation stored in the ocean is transferred to the atmosphere in the form of sensible heat and latent heat. Therefore this ocean area is a main heat source of the earth.

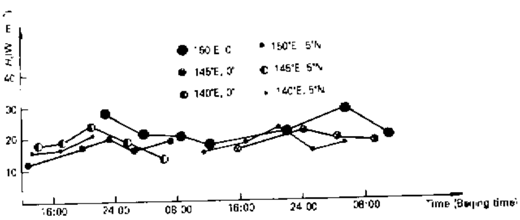

Fig.3. The diurnal variations of the sensible heat flux over the West Pacific Tropical Ocean Area.

Fig.3 is the diurnal variation curves of the sensible heat flux H_s for the 6 continuously observed stations. From Fig.3 it may be obviously seen that all day and all night for these stations the sensible heat fluxes H_s always possess positive values (i.e. the sensible heat is transferred from the ocean to the atmosphere), which has no obviously diurnal variation.In addition we may be found that the nighttime sensible heat flux is larger than that in the daytime. Fig.4 indicates the diurnal variations in the sensible heat flux H_s for the Beijing area in spring, summer, and winter using the above mentioned method on the basis of

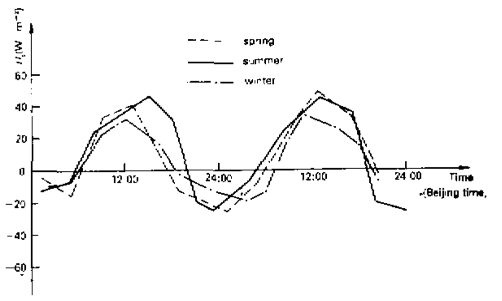

Fig.4. The diurnal variations of the sensible heat flux over the Beijing area in spring, summer, and winter.

Table 2. The Comparison of the Mean Sensible Heat Flux H_s and the Mean Latent Heat Flux H_L among the West Pacific Tropical Ocean Area, AMTEX, BOMEX, and the Beijing Area.

Region	L.	Latd.	Mean sensible heat flux H_S $(w \cdot m^{-2})$	Mean latent heat flux H_L $(w \cdot m^{-2})$
WPTOA	140° E	0	19.4	176.3
	145° E	0	16.4	158.0
	150° E	0	21.6	196.0
	150° E	5° N	17.4	102.4
	145° E	5° N	18.8	110.5
	140° E	5° N	17.0	154.5
AMTEX	125° E	25° N	17.0	88.0
			129.0	289.0
BOMEX	60° W	13° N	21.0	77.0
			11.0	135
			12.0	126.0
			13.0	158.0
			13.0	129.0
BEIJING	116° E	40° N	11.0	25.0
			10.0	48.0

Madian 325 m meteorological tower data. From Fig.4 it may be seen that sensible heat flux over the Beijing in spring, summer, and winter Beijing area all have obviously diurnal variations in spring, summer, and winter.The daytime sensible heat flux is positive but the nighttime sensible heat flux is negative. Its values, above all its positive value and its negative value. correlate obviously with sunset and sunrise. In addition to different latitudes,the cause of the difference in transferred heat for the two areas is that their surface thermal characteristics are obviously different. In addition,it may be seen (see Table 2) that the daily mean sensible heat flux values for the West Pacific Tropical Ocean Area (its daily mean values for 140 ° E, 0 ° ; 145 ° E, 0 ° ; 150 ° E, 0 ° ; 140 ° E, 5 ° N; 145 ° E, 5 ° N; 150 ° E, 5 ° N are about 19.4 W / m^2, 16.4 W / m^2, 21.6 W / m^2, 17.0 W / m^2, 18.8 W / m^2, 17.4 W / m^2 respectively) are larger than that for the Beijing area (they are about 11.0 W / m^2 and 10.0 W / m^2 in spring and summer respectively). The daily mean sensible heat flux value for the West Pacific Tropical Ocean Area is greater than that for the Beijing area by a factor about 2. It is noteworthy that this result that the sensible heat flux for the West Pacific Tropical Ocean Aera is greater than that for the Beijing area by a factor about 2 is the daily mean value. For some time of daytime the sensible heat flux over the Beijing area is greater than that of the West Pacific Tropical Ocean Area. Besides,from Fig.3 it may be seen, that generally speaking, the sensible heat flux values for the equatorial ocean area are greater than those for 5 ° N ocean area at the same longitude. The sensible heat flux for 150 ° E equatorial ocean area is a maximum. It is because the surface temperature of this ocean area reaches maximum.

Let us compare the latent heat flux for the West Pacific Tropical Ocean Area with that over the Beijing area.To estimate reasonably the latent heat fluxes for the two areas, based on the humidity profile and wind profile observed by the small tethered balloon sounding system at the West Pacific Tropical Ocean Area using formulas (2) and (3), we obtain the sensible heat flux and the latent heat flux for No.18 (November 21, see Table 1), and the Bowen ratio 0.07. Combining this result with foreign [1] observational data of the Bowen ratio β and surface temperature T (sea surface temperature or land surface temperature). we yield the dependence of the Bowen ratio β on the surface temperature T_0 which is plotted in Fig.5.

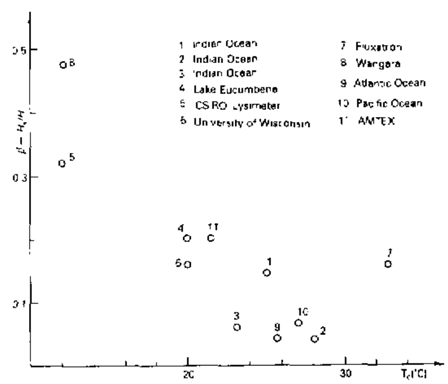

Fig.5. The observational results of the dependence of the bowen ratio β on the surface temperature T_0

[1] Priestley and Taylor (1972) discussed specially the dependence of the Bowen ration on the surface temperature T .

According to Fig.5 and observational results of the sensible heat fluxes H_s , sea surface temperature T_{ws} and land surface temperature T_{al} for the two areas, we obtain the latent heat flux H_{Ls} for the West Pacific Tropical Ocean Area and the heat fluxes H_{LB} for the Beijing area in summer and spring. Table 2 summarizes the daily mean latent heat fluxes $\overline{H}_L$ and daily mean sensible heat fluxes $\overline{H}_s$ for the West Pacific Tropical Ocean Area,the Beijing area, AMTEX, and BOMEX. From Table 2 it may be seen that the daily mean heat fluxes ($\overline{H}_L$ and H_s) over the West Pacific Tropical Ocean Area are the largest.But there is an exception to the daily mean heat flux for AMTEX during 23–28 February. It is caused by a cold air process. The daily latent heat flux H_L for the West Pacific Tropical Ocean Area is greater than that of the Beijing area by a factor about 3 in summer and by a factor about 6 in spring. This difference is caused by different thermal characteristics of the land surface and sea surface. In addition, from Table 2 it may be found that the daily mean latent flux $\overline{H}_l$ for the West Pacific Tropical Ocean Area is greater than the daily mean sensible flux $\overline{H}_s$ by a factor of about 9. Therefore the vapour latent heat transport is the most important in the all transferred heat from this ocean area to the atmosphere.

IV. SOME CONCLUSIONS

According to the above observational results of meteorology and oceanology in the West Pacific Tropical Ocean Area (127 ° E – 150 ° E, 5 ° N – 3 ° S) the following conclusions are easily drawn.

(1) On the basis of the profile data of wind, air temperature and humidity about 30 runs and by means of mean profile method it is obtained that for this tropical ocean area the drag coefficient C_D is equal to $(1.53 \pm 0.25) \times 10^{-3}$. This result is close to the results obtained by eddy correlation method and energy dissipation method under similar conditions (i.e. maritime boundary layer, the variation range of stability parameter ζ and mean wind speed U are similar). Therefore for the sea–air exchange model of this ocean area it is reliable to use the drag coefficient $C=(1.53 \pm 0.25) \times 10^{-3}$. In addition, for the remote sensing of the sea wave and wind field of this ocean area by satellite it is also dependable to use this drag coefficient value.

(2) The West Pacific Tropical Ocean Area is a heat source on the earth which has no obviously diurnal and annual variations.So far the heat transferred from this tropical ocean area to the atmosphere is concerned, the sensible heat flux only amounts to about 10%, the rest is mainly the latent heat flux.

The sensible heat fluxes for the West Pacific Tropical Ocean Area and for the Beijing area are obviously different. For the latter the sensible heatflux has obviously diurnal varation and it is larger in daytime than in nighttime.But for the former, the sensible heat flux has no obviously diurnal variation and it is usually larger in nighttime than in daytime. For different longitudes and different latitudes of this ocean area the sensible heat flux is different, and reaches maximum in the 150 ° E, equtorial ocean area. The daily mean sensible heat flux for this tropical ocean area is about 18.0 W / m^{2}and is greater than that for the Beijing area by a factor about 2.

The daily mean latent heat flux for this tropical ocean area is about 150 W / m^2,which is greater than that in spring by a factor about 6 and that in summer by a factor about 3. From the above results it is clear that for the heat transferred from this tropical ocean area to the atmosphere the latent heat flux is important.

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