Vol. 18 No. 1 Advances in Atmospheric Sciences January 2001

The Effects of Climate on Development of Ecosystem in Oasis①

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(Received April 21, 2000; revised June 12, 2000)

ABSTRACT Q14 A

When vegetation and bare soil coexist, in consideration of some ecological conditions of plant, the total evapotranspiration rate of the oasis and the temperature of vegetation and soil in different climatic and ecological conditions are calculated by using the thermal energy balance equations of vegetation and soil. The evapotranspiration rate depends on climatic and ecological conditions. In some conditions, quasi–bifurcation and multi–equilibrium state appear in the solutions of evapotranspiration rate in the areas covered by small part of vegetation.

Key words: Energy balance equations, Bifurcation, Bi–equilibrium state

1. Introduction

The development of ecological system in oasis, to a great extent, depends on the response of ecological system to environmental conditions, such as climate and water conditions, besides the physiological and biochemical conditions of vegetation. The theory concerning the development of ecological system is generally called ecological dynamics. It is substantively the interaction between the processes of ecosystem and climate.

In recent years, many numerical models have been developed on the interaction between biosphere and atmosphere, such as BAT of Dickinson et al. (1986) and SiB of Seller et al. (1986). But after Charney's deserts theory (Charney, 1975), there are few analytical theories about ecology–climate interaction.

This paper attempts in the way of theoretical analysis to study the effects of climate on the development of ecological system in oasis, in which the dynamic process of general atmospheric circulation is temporarily neglected, but the important process of the energy balance in a simple system that includes vegetation, soil and atmosphere, is considered. Basically the response of the development of different kind of vegetation in oasis to climatic conditions is just considered. The whole oasis is assumed to be a close system in the horizontal direction, and the energy budget is balanced among vegetation, soil and atmosphere.

2. Energy budget

For the vegetation in oasis, the group energy budget is calculated by averaging all of the single plant's energy budget, for example, taking

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①Sponsored by the National Key Project of Fundamental Research "The ecological environment evolution and control in western China arid area" (G1999043500).

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$$\overline{A}=\sum_{i}^{N} A_i / N \ , \tag{1}$$

where N is the number of the plants. The another way is to average the connective areas of vegetation. The oasis coverage is assumed to be S, in which a is the coverage of vegetation and $b=S-a$ is the coverage of bare soil. The average value is defined by the connective area, that is

$$\overline{A}^{a,b}=(a,b)^{-1}\sum_{i} A_{l,S}^{(i)}(da,db) \ , \tag{2}$$

where d represents one small cell.

So the average energy budget equation of vegetation in oasis is

$$\overline{F_l^{rn}}^{a}=\overline{F^h}^{a}+L_v\overline{E_l}^{a}+\overline{F_{l,s}^{a}} \ , \tag{3}$$

where l,s represent vegetation and bare soil respectively, L_v is the evaporating latent heat, E_l is the evaporating rate of vegetation. The net radiation flux in Celsius scale ($T=T^*+T'$, $T^*=273$ K) is approximately

$$\overline{F_l^{rn}}^{a}=(1-\overline{\alpha_l}^{a})Q_a-4\varepsilon\sigma T^{*3}(\overline{T'_l}^{a}-T'_a) \ , \tag{4}$$

where α is albedo, σ is the Stefan–Botzmann constant, ε is the grey body coefficient.

The sensible heat flux is

$$\overline{F^h}^{a}=\rho_a C_p C_d V(\overline{T'_l}^{a}-T'_a) \ , \tag{5}$$

where T_a is air temperature (because the scale of climate variation is larger than the scale of oasis, the contant air temperature is taken over oasis), ρ_a is air density, C_p is air specific heat on constant pressure, V is wind speed, C_d is the aerodynamics drag coefficient for heat.

According to Dickinson (1984), the latent heat flux is taken as

$$L_v\overline{E}_l=\rho_a L_v L_{AE}(r_E+\overline{r}_s^{a})^{-1}[B_e^{-1}L_v^{-1}C_p(\overline{T'_l}^{a}-T'_a)+(1-r)q^{sat}(T_a)] \ , \tag{6}$$

where B_e is the Bowen ratio, q^{sat} is the air saturation specific humidity, r is the air relative humidity, $r_E^{-1}=C_d V$, L_{AE} is the leaf surface size coefficient, r_s is the resistance coefficient of stomata.

The last term on the right–hand side of Eq.(3) indicates that there are some vertical mixed processes over the land, so the temperature is different in the air just above vegetation and bare soil. There is also some horizontal heat exchange in the air. Now we assume that the temperature in the air near the ground is proportional to the temperature of vegetation and bare soil. The distribution of vegetation and bare soil is assumed to be uneven, so the horizontal heat exchange in the air near the ground can also be assumed to take the type of turbulence. This process can be parameterized as

$$\overline{F_{l,s}}^{a} \propto \rho_a C_p C_h V(\overline{T'_l}^{a} - \overline{T'_S}^{b}) , \tag{7}$$

and

$$C_h \propto \overline{u'l} / VL , \tag{8}$$

where u' is the fluctuating value of horizontal velocity in the turbulent process, l is the horizontal mixing length, the denominator is the characteristic value of larger flow motion, V is the characteristic velocity, L is the characteristic scale, i.e. it shows the intensity ratio of turbulent and laminar flow, normally its order of magnitude is 10^{-4}–10^{-3}.

For the bare soil, the energy budget equation is

$$\overline{F_s^{rn}}^{b} = \overline{F_s^h}^{b} + L_v \overline{E_s}^{b} + \overline{F_{s,l}}^{b} , \tag{9}$$

where

$$\overline{F_s^{rn}}^{b} = (1 - \overline{\alpha_s}^{b}) Q_a - 4\varepsilon\sigma T^{*3} (\overline{T'_s}^{b} - T'_a) , \tag{10}$$

$$\overline{F_s^h}^{b} = \rho_a C_p C_d V(\overline{T_s}^{b} - T'_a) , \tag{11}$$

$$L_v \overline{E_s}^{b} = \rho_a L_v C_d V(\overline{q_s}^{b} - q_a) \approx \rho_a C_p C_d V B_a^{-1} (\overline{T'_s}^{b} - T'_a) , \tag{12}$$

$$\overline{F_{s,l}}^{b} = \rho_a C_p C_h V(\overline{T'_s}^{b} - \overline{T'_l}^{a}) . \tag{13}$$

3. Temperature of vegetation and bare soil

Substituting Eq.(4)–Eq.(7) into Eq.(3), we have

$$A_l \overline{T'_l}^{a} - B\overline{T'_l}^{a} = C_l , \tag{14}$$

where

$$A_l = \rho_a C_p V(C_d + C_h) + 4\varepsilon\sigma T^{*3} + \rho_a C_p L_{AE} B_e^{-1} (r_E + \overline{r_s}^{a})^{-1} , \tag{15}$$

$$B = \rho_a C_p C_h V , \tag{16}$$

$$\begin{aligned} C_l = {} & \{(1 - \overline{\alpha_l}^{a}) Q_a + [\rho_a C_p (r_E^{-1} + L_{AE} B_e^{-1} (r_E + \overline{r_s}^{a})^{-1})] T'_a\} \\ & - \rho_a L_v L_{AE} (r_E + \overline{r_s}^{a})^{-1} (1 - r) q^{sat} (T_\mathrm{a}) . \end{aligned} \tag{17}$$

Similarly, substituting Eq.(10)–Eq.(13) into Eq.(9), we have

$$A_s \overline{T'_s}^{b} - B\overline{T'_l}^{a} = C_s , \tag{18}$$

where

$$A_s = \rho_a C_p V(C_d + C_h + B_e^{-1} C) + 4\varepsilon\sigma T^{*3} , \tag{19}$$

$$B = \rho_a C_p C_h V , \tag{20}$$

$$C_s = (1 - \overline{a_s}^{b}) Q_a - \rho_a C_p C_d V (1 + B_e^{-1}) T'_a . \tag{21}$$

Then the average temperatures of vegetation and bare soil are respectively represented by

$$\overline{T'_l}^a = \frac{(A_s C_l + BC_s)}{(A_l A_s - B^2)} \quad , \tag{22}$$

$$\overline{T'_s}^b = \frac{(A_l C_s + BC_l)}{(A_l A_s - B^2)} \quad . \tag{23}$$

The temperatures of vegetation and bare soil are calculated. The parameter values used here are:

$Q_a = 500$ W/m^2, $V = 2.0$ m/s, $C_d = C_h = 2.75 \times 10^{-3}$, $\sigma = 5.673 \times 10^{-8}$ J/(s · m^2 · K^4), $\rho_a = 1.293$ kg/m^3, $C_p = 1004$ J/(kg · K), $L_v = 2.5 \times 10^{16}$ J/kg, other parameters will be given in figures.

Figures 1a, b show the relationship between the vegetation temperature and the two parameters $\overline{r_s}^a$, $\overline{\alpha_l}^a$ ($\overline{\alpha_s}^b = 0.12$, $r = 0.50$). In Fig. 1a, the vegetation temperature is generally lower than the air temperature 20°C, except when the resistance of stomata is very large and the albedo of vegetation is very small. In Fig.1b, the vegetation temperature is generally higher than the air temperature 5°C, except when the resistance of stomata is very small and the albedo of vegetation is very large. If there exist some vertical mixing processes between the underlying surface and the above air, vegetation can modify air temperature and reduce the variable range of air temperature.

In the above analysis B_e takes 1.0. For the case of $B_e = 0.5$ and 1.5, the results are shown in Figs. 2a–b and Figs. 2c–d, and the other parameters are the same as those used before. We can see that there is no great change in the results, so in the following calculations B_e takes 1.0.

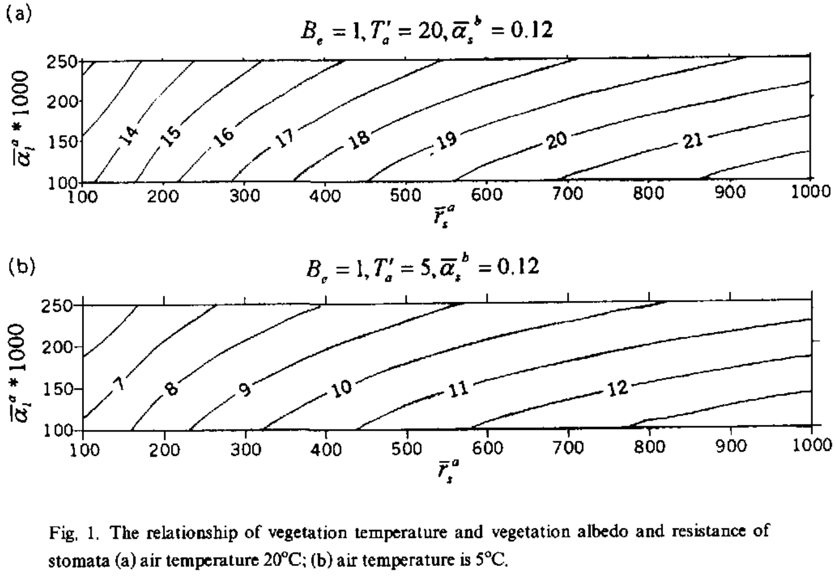

Fig. 1. The relationship of vegetation temperature and vegetation albedo and resistance of stomata (a) air temperature 20°C; (b) air temperature is 5°C.

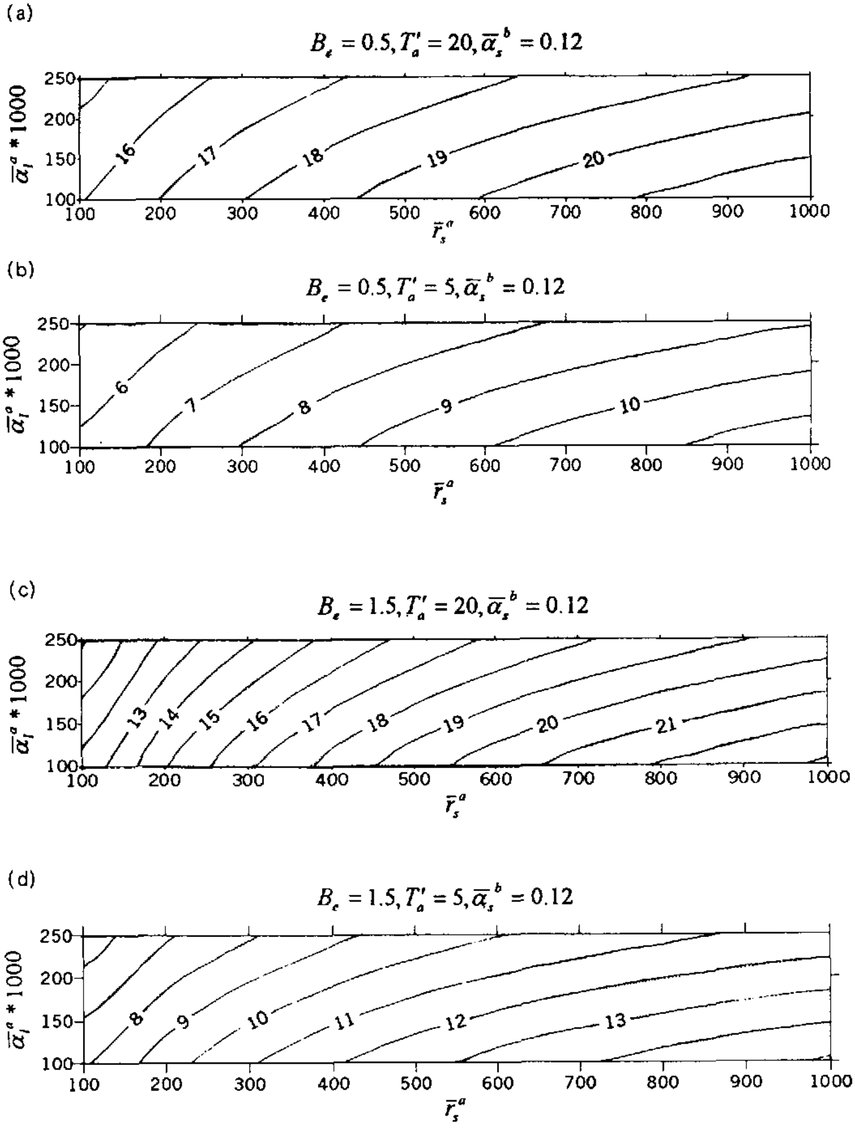

Fig. 2. With different B_e, the relationship of vegetion temperature and vegetation albedo and resistance of stomata. (a) $B_e = 0.5, T'_a = 20$; (b) $B_e = 0.5, T'_a = 5$; (c) $B_e = 1.5, T'_a = 20$; (d) $B_e = 1.5, T'_a = 5$

When the albedo of vegetation and that of soil are 0.15 respectively, and resistance of stomata is 400 s / m., the result shows that the effect of air temperature on vegetation

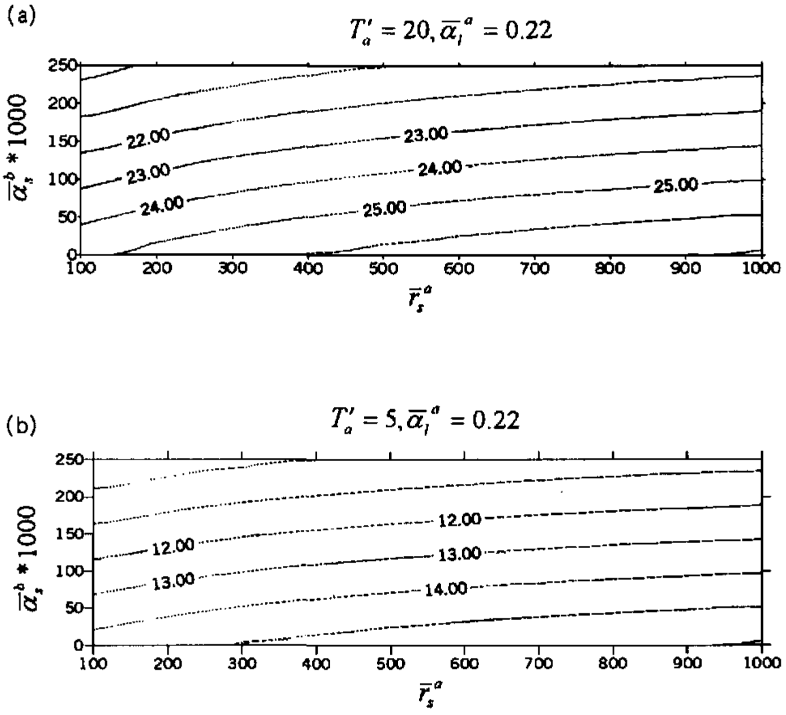

Fig. 3. The relationship of soil temperature and resistance of stomata and soil albedo (a) air temperature is 20°C; (b) air temperature is 5°C.

temperature is more important than that of air humidity on the vegetation temperature. Only when the air temperature is very high, the relative humidity is important to some extent. When the air temperature is in the range of 5–20°C, the vegetation temperature is lower than the air temperature; when the air temperature is lower than 5°C, the vegetation temperature is higher than the air temperature (figures omitted).

On the other hand, only when the air temperature is high, the resistance of stomata can increase the temperature of vegetation; when the air temperature is high, the increase of vegetation albedo reduces the vegetation temperature (figures omitted).

Figures 3a–b indicate the relationship among temperature of bare soil, soil albedo, and resistance of stomata (the vegetation albedo is 0.22, the air temperatures are 20°C and 5°C). It can be seen from the figure that the temperature of soil is generally higher than the air temperature, except when the soil albedo is large.

4. Total energy balance and total evapotranspiration of oasis

The total energy balance equation of oasis is

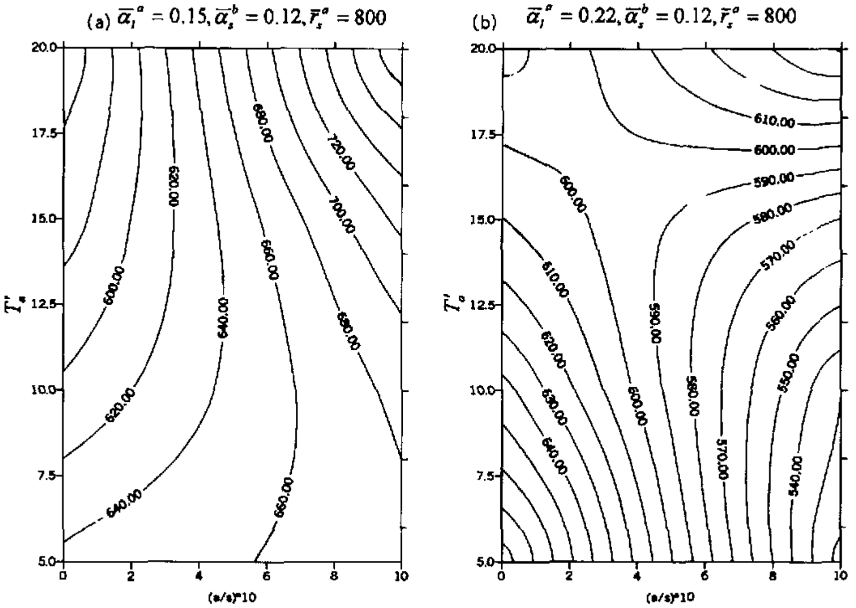

Fig.4. The relationship between evapotranspiration (mm/yr) and air temperature.

(a) vegetation albedo is 0.15 (b)vegetation albedo is 0.22

Fig. 4. The relationship between evapotranspiration (mm / yr) and air temperature. (a) vegetation albedo is 0.15 (b) vegetation albedo is 0.22.

$$\left[1-\left(\frac{a}{S}\right)\overline{\alpha_l}^{a}-\left(1-\left(\frac{a}{S}\right)\right)\overline{\alpha_s}^{b}\right]Q_a-4\varepsilon\sigma T^{*3}\left[\left(\frac{a}{S}\right)\overline{T'_l}^{a}+\left(1-\frac{a}{S}\right)\overline{T'_s}^{b}-T'_a\right]$$

$$=(\rho_a C_p C_d V)\left[\left(\frac{a}{S}\right)\overline{T'_l}^{a}+\left(1-\frac{a}{S}\right)\overline{T'_s}^{b}-T'_a\right]+L_v\overline{E}^{s} \tag{24}$$

According to this equation, the total evapotranspiration rate of oasis is calculated, for the average temperatures of vegetation and bare soil have been known.

Now we are attempting to get the relationship between oasis evapotranspiration rate and climate variables as well as ecological parameters.

Firstly, the effects of air temperature are analyzed when $r=0.50$, $\overline{r_s}^a=800$ s / m, $\overline{\alpha_s}^b=0.12$. Figure 4a shows that the effect of the air temperature on the evapotranspiration rate depends on the vegetation albedo. When $\overline{\alpha_l}^a=0.15$, the dependence is not very obvious. But when the vegetation albedo is up to $\overline{\alpha_l}^a=0.22$, the oasis evapotranspiration rate distribution changes obviously (Fig. 4b). The evapotranspiration rate is 590 mm / yr at the air temperature of 10°C, and the corresponding vegetation covering rate is about 55%. But when the air temperature increases to 17°C, vegetation has almost covered the whole oasis. We notice that the same value lines of evapotranspiration rate higher than 590 mm / yr appear double. In one of the constant evapotranspiration rate lines, the vegetation covering area reduces as the air temperature increases, for example, it is 600 mm / yr at the air temperature of 10°C, and the corresponding covering area is 45%. It reduces rapidly as air temperature increases, and at about 17°C the vegetation area is almost zero, i.e. desert appears. But there is another 600 mm / yr constant line of evapotranspiration rate that appears at high air temperature. When the air temperature is about 20°C, the corresponding area covered by vegetation is 25% and increases as the air temperature does. At about 17°C, the area covered by vegetation expands rapidly, even up to the whole oasis. The changing rule is the same as that when the evapotranspiration rate is larger than 600 mm / yr.

The above results indicate that under some environmental humidity condition (such as relative humidity is 50%, semi–arid) and some ecological condition (the vegetation albedo is 0.22, the resistance of stomata is 800 s / m), the evapotranspiration rate has bi–equilibrium state. One is in low air temperature region and another is in high air temperature region. In the low temperature region, when the air temperature increases, the area covered by vegetation reduces, oasis develops into desert; in the high temperature region, when the air temperature reduces, the area covered by vegetation increases, desert develops into oasis. The appearance of bi–equilibrium state depends sensitively on the physical and ecological conditions.

5. Multi–equilibrium state and bifurcation of evapotranspiration rate

The distribution of evapotranspiration rate in Fig. 4b can be divided into three regions. In region I and region II, the constant evapotranspiration rate lines separately lead to the direction of small and large areas covered by vegetation. The 590 mm / yr and 600 mm / yr constant lines are very close in the starting state (the air temperature is 5°C), but they go to the opposite direction, which is called quasi–bifurcation of solutions. It is understandable in physics. As to 590 mm / yr line, the area covered by vegetation is a bit more than the bare soil area. When the air temperature is low, the vegetation temperature is higher than the air temperature, the air temperature increases more than the vegetation temperature does and the evapotranspiration rate increases. To maintain the total value of evapotranspiration rate in the same value, the area covered by vegetation should reduce correspondingly. But when the air temperature is up to the some critical value(for example 17°C), as the air temperature increases, the vegetation temperature will be lower than the air temperature and the evapotranspiration rate will reduce. To maintain the same evapotranspiration rate, the area covered by vegetation should increase. This is the physical explanation to the region II. Similarly, in region III if the initial air temperature is 20°C, the vegetation temperature is generally lower than 20°C, but it still has large evapotranspiration rate. When the air temperature reduces, the vegetation temperature reduces too, and the evapotranspirature rate reduces correspondingly . To maintain the same evapotranspiration rate, the area covered by vegetation should increase. The areas covered by vegetation in region II and region III are large, even to the whole oasis, so they can be called oasis solution domain, warm domain and cold domain.

In the region I, when the air temperature increases, the vegetation temperature increases more, then the evapotranspiration rate will increase. To maintain the same evapotranspiration value, the vegetation area should reduce, even changes into desert. This region is called cold desert domain. Since the soil temperature is higher than the vegetation temperature, the evapotranspiration in this region is larger than that in region II.

At the same time, we notice that region I and region III have common evapotranspiration rate (600 mm / yr), which indicates that with the given parameters there are two solutions, one is warm oasis solution and another is cold desert solution.

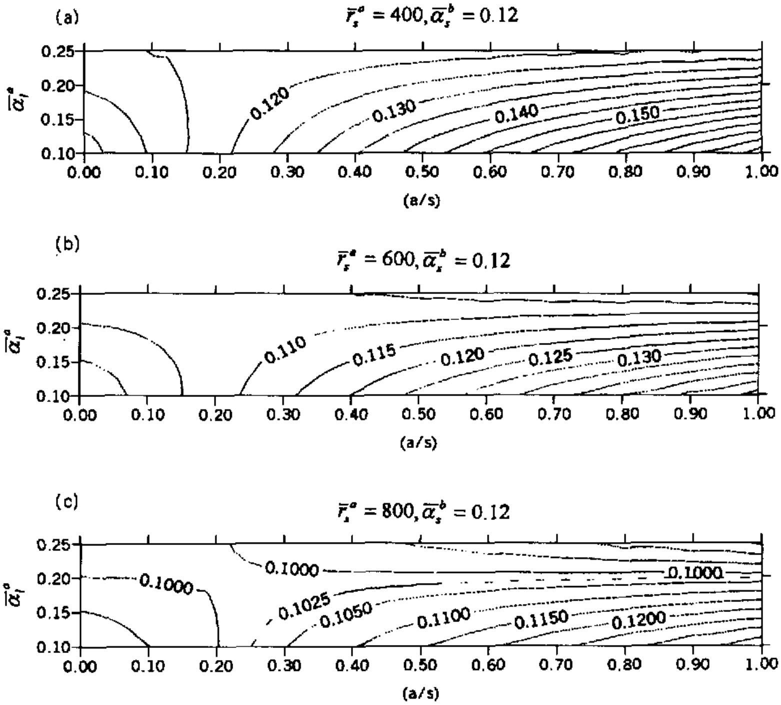

Fig. 5. The relationship of evapotranspiration (mm / h) and vegetation albedo (a) resistance of stomata is 400 s / m; (b) resistance of stomata is 600 s / m; (c) resistance of stomata is 800 s / m.

The appearances of quasi–bifurcation solutions and multi–equilibrium state solutions depend on climatic conditions of environment and ecological condition of vegetation. In Figs. 5a–c, the air temperature is 15°C, the soil albedo is 0.12, other parameters are the same as before, but the resistance of stomata is 400 s / m, 600 s / m, 800 s / m respectively. Fig. 5a shows that there is a semi–bifurcation of oasis solutions and desert solution in a small zone covered by vegetation, but the bi–eqilibrium state solution does not exist. Fig.5b indicates that evapotranspiration rate has bi–equilibrium state solutions due to increasing of resistance of stomata, one is desert solution with small vegetation albedo, the other is oasis solution with large vegetation albedo. Fig. 5c shows that in the same covering area of vegetation, one quasi–bifurcation solution and two equilibrium state solutions coexist in high and low value part of vegetation albedo. One of the bi–equilibrium state solutions is 0.1000 mm / h, the other is 0.1025 mm / h. They are respectively corresponding to one desert solution and one oasis solution.

Figs. 6a–c indicate the effect of soil albedo. The air temperature is 15°C, the resistance of stomata is 800 s / m, the soil albedo is 0.15, 0.20, 0.25 respectively. In these parameters, the desert zone has only bifurcation and does not have bi–eqilibrium state. But if the soil albedo reduces to 0.10 (Fig. 6d), two pairs of bi–equilibrium state solutions appear, similar to Fig. 5c.

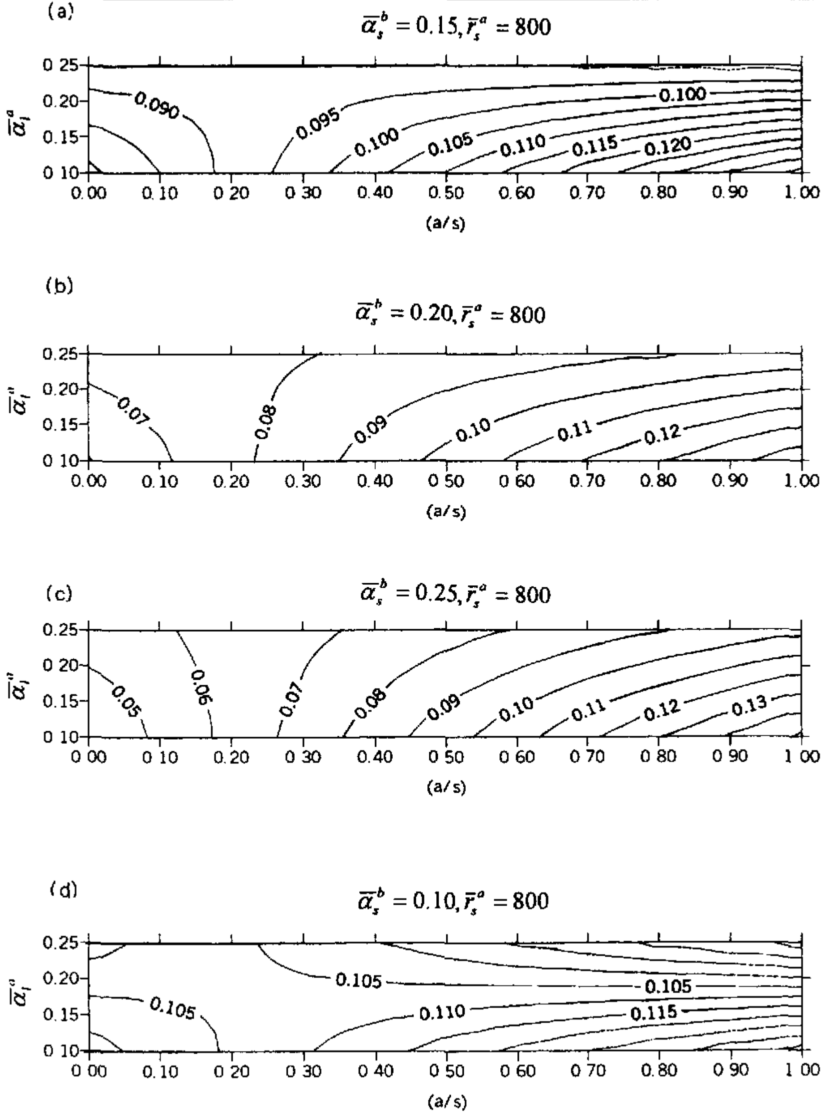

Fig. 6. The relationship of evapotranspiration (mm / h) and vegetation albedo (a) soil albedo is 0.15; (b) soil albedo is 0.20; (c) soil albedo is 0.25 (d) soil albedo is 0.10.

Under what kind of environmental and ecological conditions, the bi–equilibrium state will appear, which is a very important problem and will provide us with some references and guidance to choose different kind of plants to control the desert and should be further studied. In fact, it can be developed into an artificial control theory to modify oasis after thorough researches carried out.

6. Conclusion

Biosphere–atmosphere interaction is one of the main projects in the earth environmental science. Strengthening observation and developing numerical simulation are no doubt important, but some theoretic analysis in simple models is helpful to develop observation and numerical simulation and is practical for people to control oasis.

By using a simple thermal energy balance model, in this paper it is shown that there significantly exist bifurcation solutions and bi–equilibrium state or multi–equilibrium state solutions. It is worth to pay much attention to and put more effort into this phenomenon.

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绿洲生态系统的发展和气候的关系

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摘　要

在植被和裸地共存的情况下，考虑了植物的某些生态条件，利用热量平衡方程，对于不同气候和生态条件计算了绿洲总的蒸散率以及植被和土壤的温度．研究表明，在某些条件下，蒸散率的解可出现准分岔和多平衡态的形式，它们主要发生在植被覆盖小的地区。

关键词: 能量守衡方程, 分岔, 双平衡态形式