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LAND-USE CHANGE AND THE WATER BALANCE — AN EXAMPLE OF AN EVAPOTRANSPIRATION SIMULATION MODEL

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ABSTRACT

Land-use change in New Zealand may alter runoff, because of changes in evapotranspiration. The effects of changing albedo, surface roughness and supply of water on evapotranspiration are simulated with the Myrup numerical model of the energy balance. The model is applied to a hypothetical example in Central Otago, where a scrub/tussock area is converted to improved pasture. Investigation of the model over a wider range of conditions indicates that a change from low vegetation with an albedo of 0.35 to young coniferous forest will increase the daily potential evapotranspiration by about 75 percent. Differences between land-use types tend to be reduced as soil water deficit increases. The behaviour of the model on an annual basis is illustrated for climatological data for Alexandra. The simulated results need to be confirmed by actual measurements of evapotranspiration, but use of the model could have several advantages.

INTRODUCTION

Since the beginnings of European settlement New Zealand has undergone substantial land-use changes involving large areas (see Fig. 1 for an example). These changes from native forest, scrub, tussock and swamp to introduced pasture, ploughed fields, coniferous forests, gorse and urban areas have been essential for New Zealand's development as an agricultural economy, but they have also probably affected runoff. The exact magnitude of runoff changes is seldom known, although we can often hypothesize on the type of change. Not only are estimates of runoff change confounded by possible climatic changes over the last 100 years, but often we are unable to quantify critical processes in the hydrological cycle.

Land-use changes are continuing, and as the demands for flood protection or for water grow, some of these changes become more

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FIG. 1 — Land-use change by forest clearance, 1840–1965, showing the major areas cleared of their forest covers during European settlement. *From Cumberland (1967).*

controversial. Further extensive changes are contemplated from beech to coniferous forest, from rural to urban land use, and from tussock grassland to introduced grasses. These changes may alter runoff. This would be especially important for rivers where the supply of water is already committed for irrigation schemes, growing urban areas, or hydro-electric power development, and should be taken into account in the planning of land development.

In part, land-use change alters runoff because significant changes in evapotranspiration may be produced. For example, Denmead (1969) has measured 41 percent greater evapotranspiration from a *Pinus radiata* forest than from a nearby wheat field. Our ability to predict changes in water supply induced by land-use change depends on how well we can understand and model the processes involved. This may be done experimentally as in the IHD experimental basin programme, and/or by the development of simulation models.

CHANGES IN COMPONENTS OF THE WATER AND ENERGY BALANCES PRODUCED BY LAND-USE CHANGE

A change in land use may influence many subsystems of the hydrological cycle, and hence the amount of water available for evapotranspiration, as well as water yield:

(a) A change in vegetation type may alter interception storage because of different plant size and architecture.

(b) A change in vegetation may alter transpiration because of different plant physiological restrictions on water loss.

(c) Changes at the air/ground interface may alter surface storage.

(d) Cultivation, logging, stock trampling, etc., may alter soil structure and hence the ability of the soil to store and transmit water.

As well, land-use change may alter the energy available to power evapotranspiration:

(a) Albedo changes will influence the amount of radiant energy available for partitioning among heating the air (the sensible heat flux), evapotranspiration (the latent heat flux*), or heating the ground (the ground heat flux).

(b) Different plant architectures with different foliage density and foliage angles will absorb different amounts of radiation.

(c) A change in vegetation may alter the emmissivity of the surface and hence the loss of longwave radiation.

(d) A change in vegetation will alter the surface roughness and hence mechanical turbulence in the air above. This may alter evapotranspiration because the degree of mechanical turbulence partly determines the rate of turbulent diffusion of water vapour from the vegetation or soil to the atmosphere.

APPROACH

It is the purpose of this paper to examine changes in evapotranspiration produced by different land-use types. It is convenient to draw on the well developed theory of energy transfer at the air/ground interface to use a model to compute the fluxes of the energy balance. Since the water balance and the energy balance are linked by the evapotranspiration term, this strategy allows some understanding of changes in the hydrological cycle.

The rate of evapotranspiration will be controlled by:

(a) The availability of water from vegetation, interception, surface, and soil storage. In the model presented here, this is crudely repre-

* The latent heat flux (LE) = latent heat of vaporization of water × evapotranspiration.

sented by a gross term known as the surface wet fraction (w). This will be inversely related to the resistance to water movement in plants and soils. Apart from changes in this term the model disregards any change in soil hydrological or thermal properties.

(b) The availability of energy. Initially, this is assumed to be totally supplied by clear-sky radiation, although the model does sometimes augment this with a supply of sensible heat. Changes in albedo with changes in land use are examined, but the model disregards changes in plant architecture, foliage density and surface emmissivity.

(c) The degree of turbulent transfer. Only changes in mechanical turbulence induced by changes in surface roughness are examined. Changes in turbulent transfer induced by stability or instability of the atmosphere are not included, as these are not usually influenced by land-use change.

In summary, when comparing the effects of land-use change, this model investigates changes in albedo, surface roughness, and the surface wet fraction. Denmead (1972) has examined land-use change in terms of some other properties such as heat transfer coefficient and crop structure. His analysis is based on the combination formula of Slatyer and McIlroy (1961), a related but different model to that discussed here.

THE MODEL

The energy-balance model uses the simplest set of equations which still retain the essential physics of the atmospheric surface layer. The model was originally designed by Halstead *et al.* (1957) and developed further by Myrup (1969) and Outcalt (1972).

The model seeks to compute each flux of the energy balance equation:

$$R_n + LE + H + G = 0 \qquad (1)$$

where R_n is the net radiation flux, LE the latent heat flux (L = latent heat of vaporization of water, E = evaporation), H the sensible heat flux, and G the flux of heat into the soil. All terms are defined to be positive for transfer toward the ground/air interface. By using eight further equations it is possible to compute each of the above fluxes over a 24-hour period.

The net radiation can be computed from

$$R_n = R_0 T_r \, (\sin\varphi \sin\delta + \cos\varphi \cos\delta \cos\gamma)(1-\alpha) - IR_n \qquad (2)$$

where α is the albedo, T_r the transmission coefficient of the atmosphere (a function of the dust content and precipitable water), R_0

the solar constant, φ the latitude, δ the solar declination, and γ the solar hour angle. In this version of the model IR_n, the net longwave radiation, is computed with the universal formula given by Idso and Jackson (1969).

The turbulent fluxes of sensible heat (H) and latent heat (LE) may be written as

$$H = -\rho C_p K_h \frac{\delta \theta}{\delta z} \tag{3}$$

$$LE = -\rho L K_v \frac{\delta q}{\delta z} \tag{4}$$

where θ is potential temperature, ρ air density, C_p the specific heat at constant pressure, q the specific humidity, K_h and K_v the turbulent diffusivities for heat and water vapour respectively, z the distance from the interface, and L the latent heat of vaporization.

For neutral and near-neutral stability, the diffusivity (K_m) for the flux of momentum is given by

$$K_m = \frac{k^2 u z}{\ln(z/z_0)} \tag{5}$$

where k is the von Kármán constant, z_0 is a function of surface roughness, and u is wind speed at height z. This assumes a logarithmic wind profile.

Specific humidity near the surface is calculated as

$$q_0 = (w/L)[3.74 + 2.64\ (T_0/10)^2]10^{-3} \tag{6}$$

where q_0 is the specific humidity and T_0 the temperature at height z_0. The surface wet fraction, w, can be regarded as the fraction of the surface occupied by freely evaporating surfaces (Myrup, 1969), or alternatively as the ratio of actual to potential evapotranspiration in conditions where water supply is limiting.

If it is assumed that

$$K_m = K_h = K_v \tag{7}$$

then H and LE can be computed by means of equations (3) and (4).

The soil heat flux (G) is given by

$$G = -K_s \frac{\delta T}{\delta z} \tag{8}$$

where T is soil temperature, and K_s the soil thermal conductivity. To calculate the soil heat flux at any time t, the one-dimensional form of the Fourier heat conduction equation must be solved, i.e.

$$\frac{\partial T}{\partial t} = \lambda \frac{\delta^2 T}{\delta z^2} \tag{9}$$

where λ is the soil thermal diffusivity.

Using equations (1) to (9) it is possible to obtain a continuous simultaneous solution, and to calculate each component flux of the energy balance given in equation (1). The major boundary conditions are provided by a constant temperature and specific humidity upper-airstream boundary in the atmosphere, to which a clear-day radiation input is applied. Given the input data and boundary conditions, there is a unique surface temperature which will balance equation (1). That temperature is termed the equilibrium surface temperature, and with ideal modelling, it will converge with the observed diurnal thermal regime. For further details of the model see Myrup (1969: p. 909–910), and Outcalt (1972: p. 629–632).

The fundamental assumptions made in formulating the model are:

1. All surfaces are level and homogeneous. Meteorological and soil parameters are also considered to be horizontally homogeneous.

2. The turbulent diffusivities for the sensible heat and latent heat fluxes are the same as for the momentum flux (equation 7). This implies that the atmosphere should be in a state of near-neutral stability, a condition that is not always met.

3. The energy fluxes between a height just above the surface z_0 and some height z_2 are constant.

4. Temperature, wind speed and specific humidity are constant at some height z_2.

5. Clear sky conditions prevail, and no snow is present.

6. The vegetative cover is uniquely characterized by the roughness length z_0.

It must be recognized that the model is too simple to allow for feedback effects of biological processes on the micro-climatology of the vegetative cover. The only feedback of this kind permitted is through the effect of canopy growth (by increasing z_0), or through reduction of water supply for evapotranspiration (by decreasing the surface wet fraction).

To illustrate the model, suppose an area in Central Otago were to be converted from a scrub/tussock association to a pasture made up of introduced grasses. This example is appropriate, because a large area of Teviot soils is slated for such development in the near future. The model is to compute the clear-day diurnal variation in fluxes of the energy balance for 1 January. At this time of the year the amount of incoming radiation, and hence energy for evapotranspiration, is likely to be greatest. The model would be supplied with the following input, with the values for this example given in brackets:

1. Latitude (46° S).
2. Solar declination (23°, corresponding to 1 January).
3. Atmospheric pressure (1015 mbar).
4. Precipitable water in the atmosphere (14.00 mm, a mean value for 46° S as given by Sellers, 1965).
5. Dust particles in the atmosphere ($0.2/cm^3$, a representative value given for rural areas by Gates, 1962).
6. Radius vector of the earth's orbit (0.98324, see List, 1968, Table 169, p. 495).
7. Vapour pressure (11.0 mbar, a representative value for summer at 46° S).
8. Mean diurnal air temperature (15° C).
9. Wind speed (5 m/s).
10. Soil thermal diffusivity ($0.0037\ cm^2/s$, a typical value for a moist, sandy clay).
11. Soil heat capacity ($0.59\ cal\ cm^{-3}\ degC^{-1}$).

The above variables were held constant for the two types of land use. Albedo, roughness length and the surface wet fraction were varied for each case considered. For pasture grass the values of these parameters are well documented (e.g. Sellers, 1965). The albedo was taken as 0.15 and the roughness length as 0.75, corresponding to a vegetation height of 6 cm. The grass was assumed to completely cover a soil surface at field capacity so the surface wet fraction was set equal to 1.0.

For the scrub/tussock association, the vegetation was taken as 150 cm high. The roughness length was computed as 20 cm by using the formula of Tanner and Pelton (1960). The surface wet fraction was set equal to 1.0. No known measurements of albedo for this type of vegetation in Central Otago were available, so a value of 0.35 was assumed. This is probably on the high side, but similar albedos have been recorded over savanna and dry soil (Sellers,

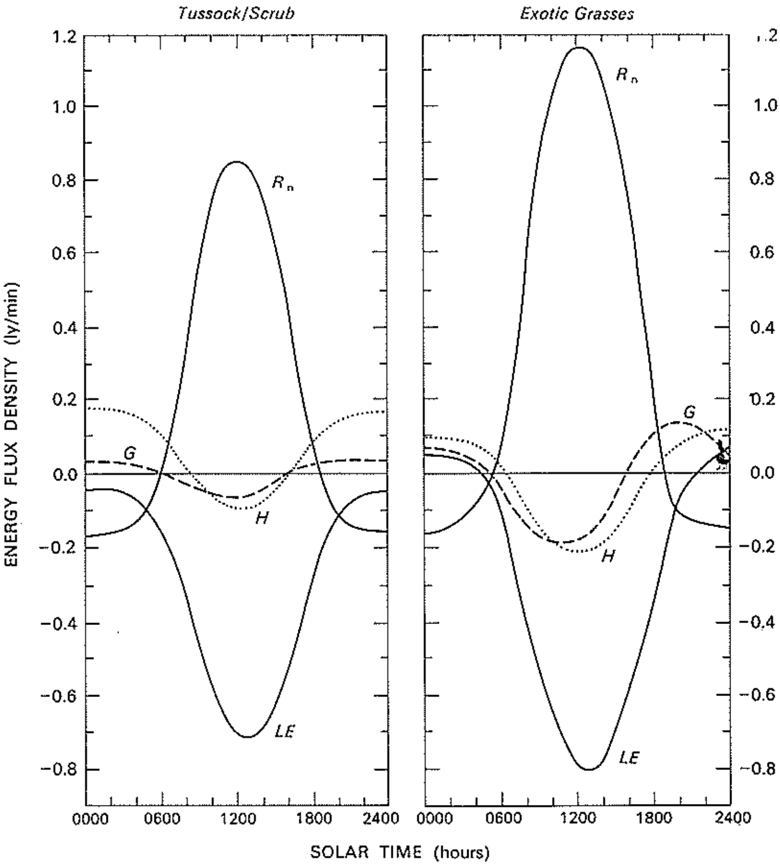

FIG. 2 — Diurnal variation in simulated fluxes of the energy balance, 1 January at 46° S.

1965). The comparison between the energy balance fluxes for clear sky conditions and adequate moisture supply is shown in Fig. 2. The magnitude of the fluxes and their diurnal behaviour is consistent with grass-like surfaces measured in similar environments (e.g. Sellers, 1965).

The simulation suggests that when a change from scrub/tussock to exotic grasses is made, the net radiation increases. Since this flux is the driving force for the other fluxes, they too increase. The change in the latent heat flux is most important here, for it indicates that the conversion to improved pasture will increase evapotranspiration in the middle of the day and hence may decrease runoff.

At night, a small amount of the lost water may be regained by condensation on to the grass surface (latent heat flux directed towards the surface). Water yield may be further reduced by this land-use change at high altitudes, because there is some evidence to suggest that snow tussocks augment natural precipitation by interception of cloud and fog water droplets (Mark and Rowley, 1969).

BEHAVIOUR OF THE MODEL

The behaviour of the model was investigated over a wider range of conditions. The albedo was varied between 0.15 and 0.35, which covers most natural situations. Many vegetative surfaces have

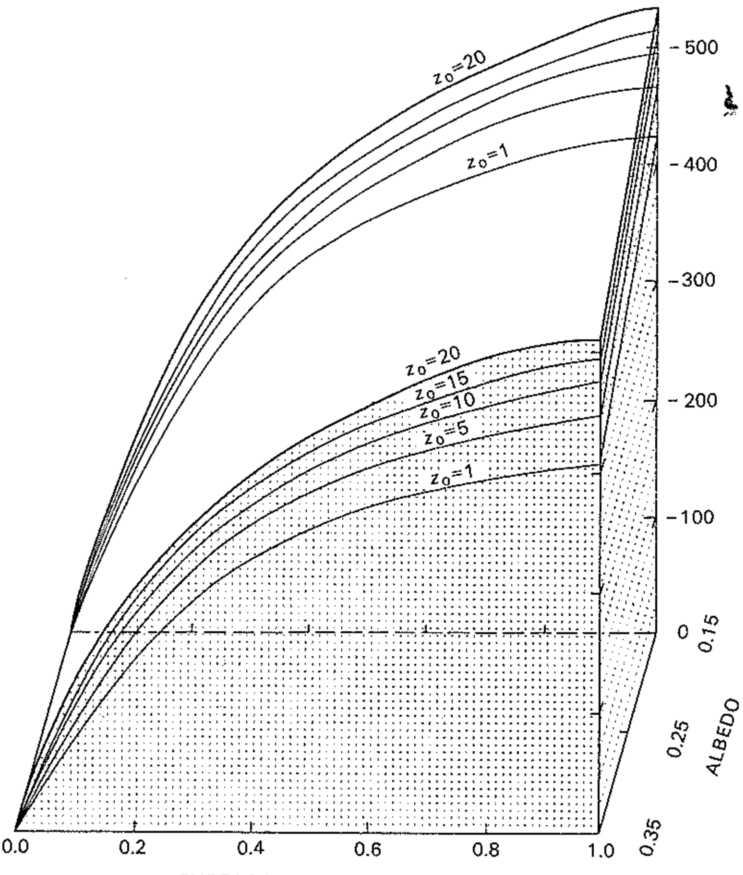

FIG. 3 — Variation of simulated daily latent heat flux density with albedo, roughness length and surface wet fraction. For 1 January at 46° S.

an albedo close to 0.25 (Sellers, 1965), although very few measurements have been made in New Zealand. The surface wet fraction was varied from 1.0 (moisture readily available) to 0.0 (no moisture available). The height of the vegetation was varied from 7 cm to 150 cm, corresponding to a roughness length from 1 cm to 20 cm. Other input was the same as the previous example.

The resultant daily evapotranspiration as a function of albedo, surface wet fraction and roughness is given as a three-dimensional perspective diagram (Fig. 3). The effect of varying albedo by ± 0.10 about 0.25 is to change net radiation by ± 20 percent, which in turn produces a change in evapotranspiration of about ± 17 percent. These results are similar to those found by Denmead (1972) with a different model. As albedo increases, the evapotranspiration decreases. The range of albedo for most agricultural crops is small, so that it is unlikely that large differences in potential evapotranspiration will result from a change from one crop to another. On the other hand, differences of some significance could result from a change in land use involving beech and coniferous forest, urban areas, tussock, and introduced grasses where the differences in albedo may be larger.

An increase in vegetation roughness increases turbulence above the surface and hence can increase the transport of water vapour to the atmosphere. Here, a change in vegetation height from 7 cm to 150 cm increases evapotranspiration by about one-third. This analysis suggests that a change from low vegetation with an albedo of 0.35 (e.g. Central Otago scabweed land) to a taller vegetation with albedo of 0.15 (e.g. a young coniferous forest) will increase the daily potential evapotranspiration by about 75 percent.

Increasing the vegetation height can have the effect of reducing its surface temperature, even to the extent of making this less than air temperature. Under these conditions, the sensible heat flux is directed toward, rather than away from the surface, adding more energy to that of net radiation to power evapotranspiration. This results in negative Bowen ratios (Fig. 4). The Bowen ratio (H/LE) is an index of the relative amounts of energy used to heat the air and to evaporate water. The analysis suggests Bowen ratios increase with lower albedo, or as the surface wet fraction decreases. Eventually, when the surface wet fraction becomes zero, there is no evapotranspiration and hence no latent heat flux. Most of the energy absorbed at the surface then goes to heating the air (H) or to heating the ground (G).

This serves to illustrate that the large changes in evapotranspiration which may be produced by changes in albedo and rough-

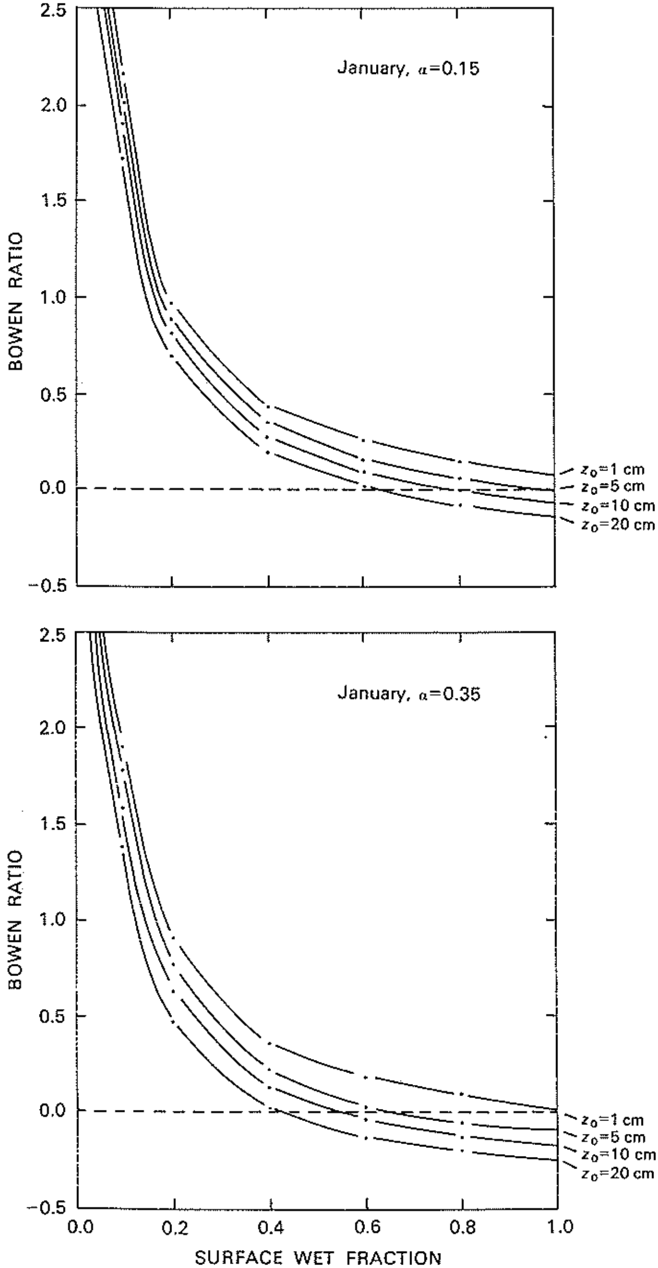

FIG. 4 — Variation of daily Bowen ratio with albedo, roughness length and surface wet fraction. For 1 January at 46° S.

ness length, are still less than those induced by a reduction in the water supply. This reduction can occur through land-use change by decreasing the interception and surface storage, or by altering the properties of the soil which affect the transmission and storage of water. However, reduction of the water supply occurs naturally, without land-use change, as the soil dries out from field capacity towards the wilting point. The model only crudely represents this process through the surface wet fraction, and there is need to improve this function by calibrating it against soil water deficit. In any event, the effect of increasing soil water deficit is to smooth out differences in evapotranspiration between different land-use types.

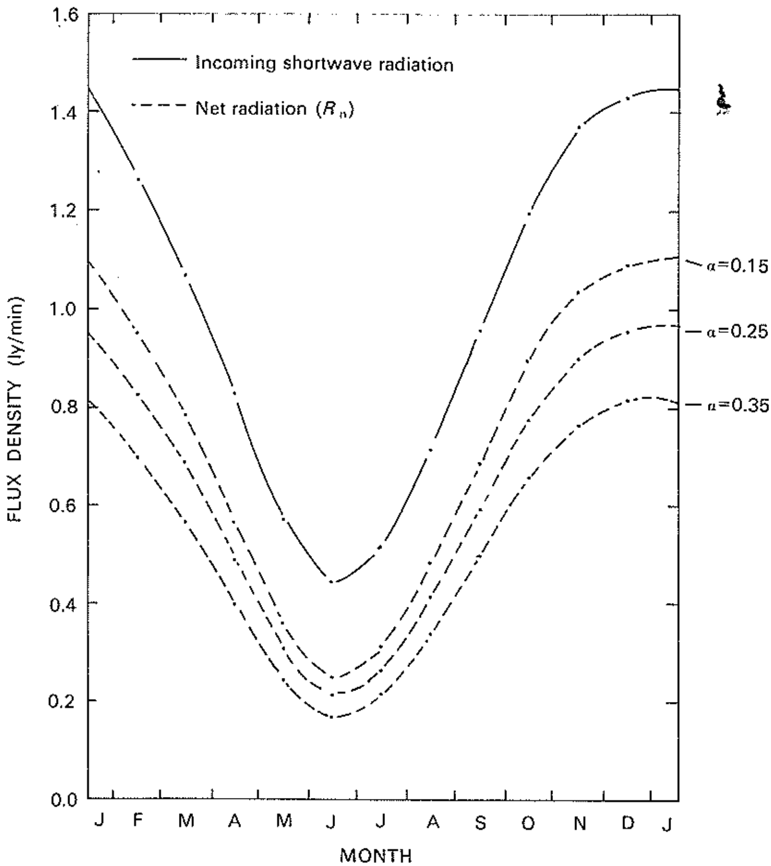

FIG. 5 — Simulated annual variation of incoming shortwave radiation and net radiation for various albedos. Values for clear sky conditions at midday for Alexandra.

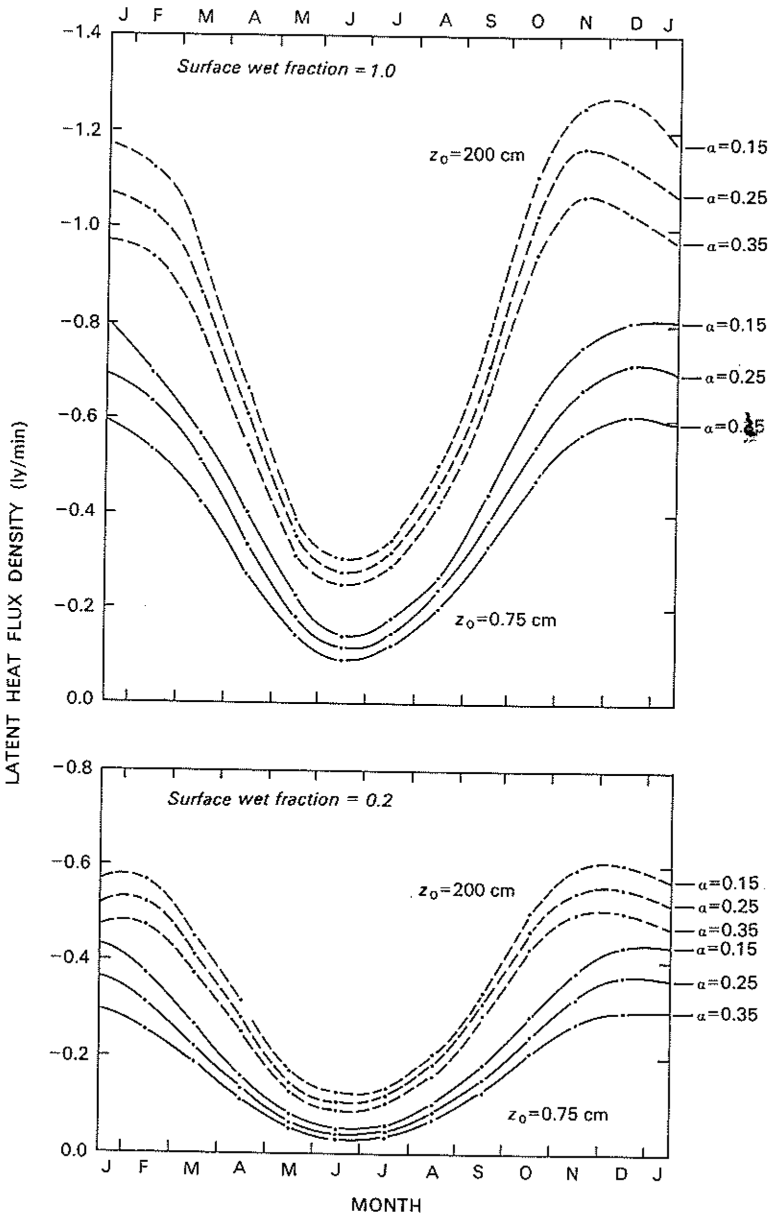

FIG. 6 — Latent heat flux density for conditions of Fig. 5.

BEHAVIOUR OF THE MODEL ON AN ANNUAL BASIS

The model is used to compute evaporation throughout the year based on available meteorological data from Alexandra (N.Z. Meteorological Service, 1966). Other input variables were obtained from List (1968), or were estimated. The calculations represent solutions for midday on clear days at the middle of each month for various albedos, roughness lengths and surface wet fractions.

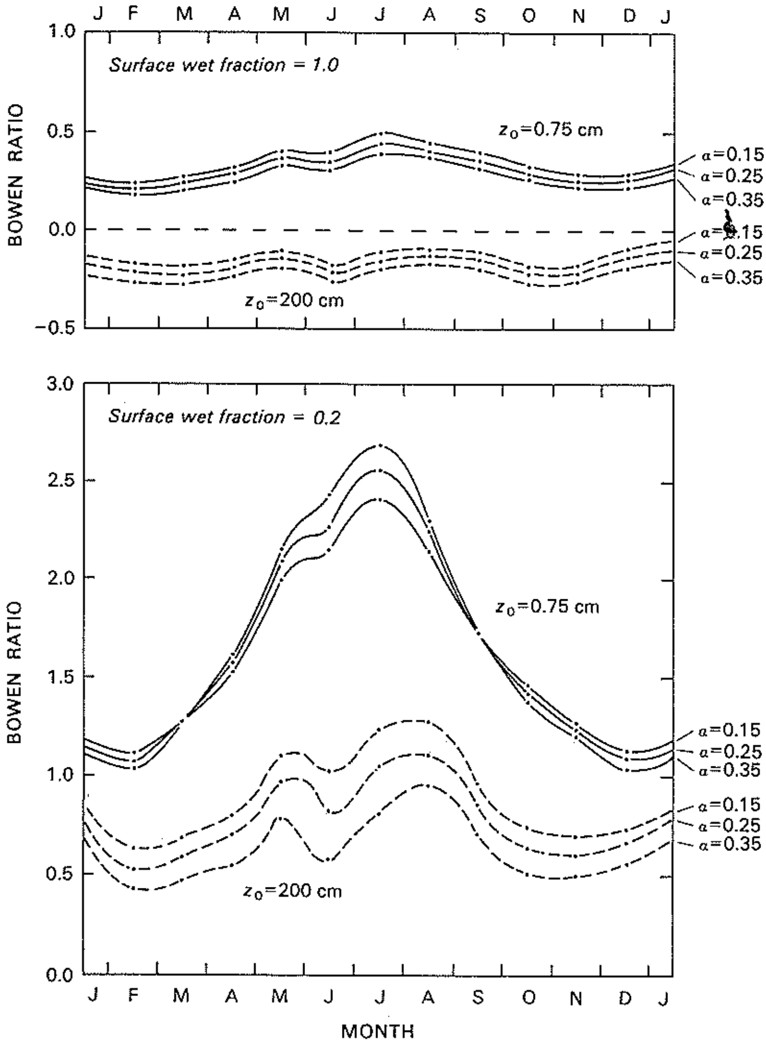

FIG. 7 — Bowen ratios for conditions of Fig. 5.

Incoming shortwave and net radiation vary sinusoidally throughout the year (Fig. 5). The previously noted decrease in net radiation with increasing albedo is clear. The shape of the radiation curves is reflected in the curves for evapotranspiration (Fig. 6). Because of low precipitation, a surface wet fraction of 1.0 is probably seldom attained near Alexandra, except for irrigated areas. The graphs reaffirm the increase in evapotranspiration associated with a decrease in albedo, an increase in vegetation height, and an increase in available water.

Bowen ratios are always negative for tall vegetation if there is a plentiful supply of water (Fig. 7). This results from a flux of sensible heat towards the surface which helps power evapotranspiration, sometimes to values greater than net radiation. Otherwise, Bowen ratios are positive, with the value of 0.2 to 0.3 typical of low vegetation with adequate water supply. As water availability decreases, Bowen ratios rise to greater than 1.0. It is interesting to note that the model indicates that relatively more energy goes to power evapotranspiration, rather than to heating the air, in summer when the radiation input is higher.

CONCLUSIONS

The application of a numerical model to simulate the fluxes of the energy balance indicates that land-use changes will alter evapotranspiration, and hence runoff. This may have important consequences for downstream river flow, where water is already committed for use. An increase in evapotranspiration, and hence decrease in runoff, is produced if land use is changed in such a way as to decrease albedo, increase surface roughness, or increase water availability.

It is stressed that the assumptions of the model will not always apply. In addition, many relevant processes of the real world are modelled in a very simple fashion, if at all. For these reasons, the actual results given here may not be valid, although they probably indicate the general nature of the changes. Hence this paper should not be regarded as a definitive statement about the effects of a particular land-use change, but as a useful approach to the problem. However, it is worth noting that Brazel and Outcalt (1973) have shown that at an Alaskan alpine pass the simulated evaporation using the Myrup model compared favourably with actual observations for clear weather, especially for sites where water availability is not a limiting factor. Drier sites produced some differences between simulation and actual evaporation calculations, but these differences did not exceed ± 0.10 ly/min, which is within the error limit of their observations.

The model could be applied to land-use changes in experimental basins. For example, it can identify areas of the hydrological cycle where more information is needed, and can provide a framework for research. Conversely, the experimental basin could provide data with which to test the model. Ultimately, models such as this might provide assessment of the hydrological impact of land-use change based largely on theory, rather than on empiricism. Such theoretical models are more desirable because they tend to be more general and less sensitive to local conditions. Thus they are more liable to extrapolation outside the immediate experimental basin.

There is need to have measures of albedo of typical New Zealand vegetation associations such as tussock grassland, beech forest and urban areas. These albedos are almost unknown at present, yet are important in determining evapotranspiration rates. The model presented here needs to be tested for New Zealand conditions against actual values of evapotranspiration measured by lysimeter or micrometeorological methods. Finally, there is need to calibrate and expand the surface wet fraction in order that it realistically models the amount of water available for evapotranspiration.

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