CHAPTER 2. MODEL DESCRIPTION

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2.1 Introduction

The physical and biological processes describing the surface water, energy and solute balances of the plant–soil–atmosphere system are, in general, well understood. Models of that system can be formulated at almost any level of complexity with as many or as few processes as required. The level of model complexity is usually determined by the application.

Historically, physically based models have been developed to represent the real world with increasing detail. The place and use of such models, and the information contained in the data required to run them, have been debated in hydrology literature for over 20 years, most recently by Beven (1989, 1993), Hauhs (1990), Wheater and Jakeman (1993) and Barnes (1993). These authors argue that physically based models are most appropriately used in exploring the interactions between processes and fluxes under different management and/or climatic regimes, given clearly stated assumptions about which small-scale processes are relevant.

In Australia, most environmental degradation is associated with changes in the surface water balance induced by changes in land cover. The temporal and spatial scales over which these changes evince themselves precludes field experimentation as a wholly sufficient or practical investigative tool for identifying optimal or appropriate land use. Decision-makers are therefore reliant on models, especially physical process models, for predicting expected changes in the landscape; the diversity of recent hydrological modelling tools in use in Australia (see Grayson and Chiew 1994 and Hatton *et al*. 1994 for recent reviews) is testament to this need.

The WAVES model was designed to enable the simulation of soil–vegetation–atmosphere system behaviour under alternative management and climatic variation. The aim is to represent the interactions and feedbacks of the system in the simplest possible way, yet with adequate description of the key processes. The model predicts the dynamic interactions, and fluxes of energy, water, carbon, and solute within soil–vegetation–atmosphere systems.

The model adopts a one or two layer canopy representation with a soil layer underneath. The aerodynamic resistance at the top of the canopy is determined based on Monin–Obukhov surface layer similarity theory and the within canopy aerodynamic resistances are estimated using the

mixing-length approach (Raupach and Thom, 1981). The boundary layer resistances are neglected for simplicity. The model formulates the physiological control on transpiration using the canopy resistance calculated as a function of the net assimilation rate, and the vapour pressure deficit and CO_2 concentration at the canopy surface. The soil hydrology is described by the Richards equation. A distinguishing feature of the model is to couple the soil–vegetation–atmosphere system by changing the value of the saturation vapour pressure deficit of air in the canopy. The model can be used to predict plant growth using a saturation rate kinetics formulation and to simulate solute transport in the soil (Hatton *et al.* 1992, Dawes and Short, 1993, Wu *et al.* 1994, Zhang *et al.* 1996, Dawes *et al.* 1997).

This section provides a detailed technical description the conceptual framework, theoretical background, and process representation. It also describes the governing equations, parameter estimation, and assumptions within WAVES. The document is adapted from existing publications on WAVES.

2.2 General Principles

A realistic formulation of the interaction between soil–vegetation–atmosphere must represent the following physical or biological processes:

- radiation balance: to determine the available energy at the surfaces of each canopy layer and the underlying soil, to estimate sensible and latent heat fluxes;
- interception: to determine the amount of water intercepted on the canopy surface;
- atmospheric turbulence: to determine the atmospheric and boundary layer resistances for momentum, heat and mass transfer;
- canopy physiology: to determine the physiological control of transpiration;
- runoff generation: to determine surface runoff based on precipitation, evaporation and infiltration;
- soil dynamics: to determine heat and water transport in the soil, recharge to groundwater, and the available soil moisture;
- solute transport and impact on plant growth: to estimate conservative solute transport within the soil column and the impact of salinity on plants.

This list is not comprehensive, but highlights the major processes and interactions. The complexity of a model should be constrained by the questions being answered, critical interactions, probable data availability, and the need to represent the processes consistently. Any model of the soil–vegetation–atmosphere should be designed with a clear objective and used appropriately. WAVES is composed of four modules, which solve energy, water, carbon, and solute balances on a daily time step. The schematic diagram of WAVES is shown in Fig. 2.1.

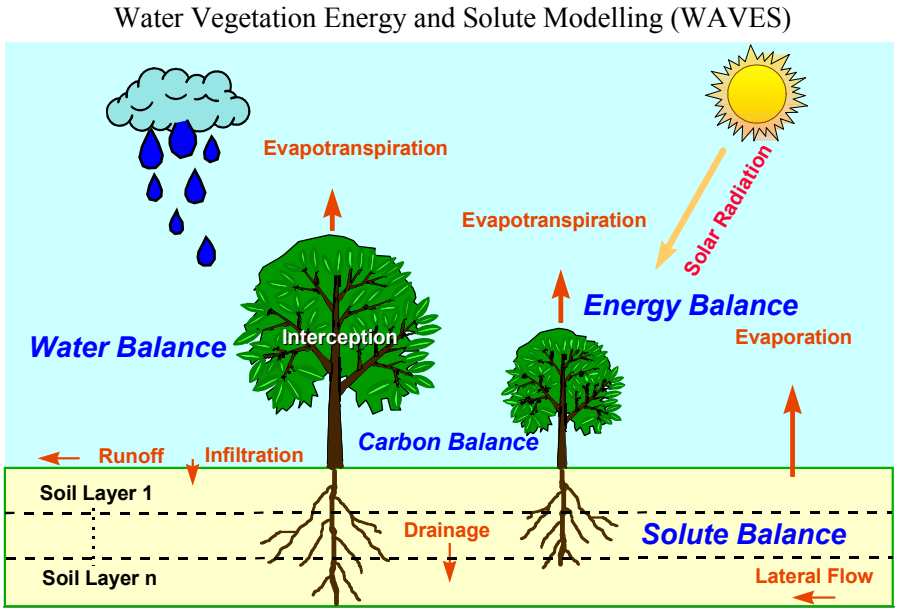

Fig. 2.1. Conceptual diagram showing the major processes modelled by WAVES.

The energy balance module calculates net radiation from incoming solar radiation, air temperature, and humidity, then partitions it into canopy and soil available energy using Beer's law. The carbon balance and plant growth module is based primarily on calculating actual daily carbon assimilation from a maximum value, and the relative availability of light, water, and nutrients; the limiting effects of temperature on light, and salt in the soil water are modelled explicitly. The soil water balance module handles rainfall infiltration, overland flow, soil and plant water extraction, moisture redistribution, drainage (recharge), and water table interactions. The solute balance module solves a convection-dispersion equation, in the same way as soil moisture dynamics (Dawes and Short, 1993). It is assumed that the solute concentration does not interact with soil hydraulic properties, so water fluxes and contents are constants with respect to the solutes. The feedback to plants of salinity is through the reduction in apparent available water due to the osmotic potential induced by dissolved common salt (sodium chloride, NaCl) alone.

These four modules are linked in the following way. At the beginning of each day time-step, the climatic forcing variables are set. The next step is to use the current values for leaf area to perform the surface energy balance, and set limits on the availability of water to plants for this day. The plant growth routines are required next to calculate gross carbon assimilation, plant respiration, and root growth. The actual assimilation rate is used to calculate canopy conductance, and plant transpiration. The soil evaporation is also calculated using the surface conditions from the start of the day. These fluxes set the surface boundary condition, and internal sinks, for solution of Richards' equation, which partitions rainfall into runoff, infiltration, drainage or uptake from a

watertable, and stored water. After this solution, the internal water fluxes are set, and conservative solute transport of common salt can be determined.

2.3 Rainfall interception

Rainfall interception for each canopy layer and the litter layer is scaled as a linear function of leaf area index as in Running and Coughlan (1988) and Hatton *et al*. (1992). When no rain occurs on a given day, evaporation is allowed to take place at the rate calculated from the Penman–Monteith equation. Where rain does occur, a sub-daily time step equal to the rainfall duration is performed with the intensity calculated from the rainfall amount and duration, and then evaporation from the surface occurs for the remainder of the day. When rain falls, any existing vegetation canopies can intercept water according to a linear relationship between leaf area and maximum interception:

$$I_{max} = K_r LAI \tag{2.1}$$

where K_r is the rainfall interception coefficient (m LAI^{-1}), and LAI is the leaf area index of the canopy layer. Any intercepted water must be evaporated before transpiration can occur, and all precipitation in excess of the interception capacity, reaches the next lowest canopy, or the ground surface. Vertessy *et al.* (1993) showed that this rainfall interception model worked well over a wide range of rainfall rates, and with a growing forest cover.

2.4 Energy balance and evapotranspiration

2.4.1 Energy balance

Evapotranspiration and sensible heat flux into the atmosphere are constrained by the available energy at the soil–vegetation–atmosphere interface. Depending on the nature of the surface, this interface may consist of water, bare soil, vegetation, or of some other substrate. For practical purposes, the energy balance equation can be written as:

$$R_n - P_s - G - A_h - \lambda E - H = S \tag{2.2}$$

where R_n is net radiation, P_s the energy flux for photosynthesis, G is the ground heat flux, A_h is the advection of energy from the surrounding, λE is evapotranspiration, H is the sensible heat flux, and S is the rate of energy storage.

The importance of the terms in the energy balance equation depends on the nature of layer for which the energy balance is written. In general, net radiation is the dominant term in the energy balance equation not only in the absolute sense, but also because the magnitudes of all the other

terms depend to some extent, directly or indirectly, on the size of net radiation. The net radiation flux can be determined from meteorological data, and the method is described in the next section. The energy absorbed for photosynthesis in day-time ranges from 6 to 16 W m^{-2}, depending on species (Monteith and Unsworth, 1990). Thus, it is usually negligible in comparison with net radiation and often be ignored in the energy balance equation, except when the objective is to determine the rate of energy absorption by photosynthesis itself. The ground heat flux is positive during the day, ranging from 2 to 20 per cent of net radiation, and negative at night with more or less the same magnitude. However, the daily mean values of the ground heat flux are often one or more order of magnitude smaller than the major terms in the energy balance equation (Brutsaert, 1982, Zhang *et al.,* 1996). The energy advection term is often neglected in the energy balance equation because it is difficult to estimate (Thom, 1975). Evapotranspiration and sensible heat flux are the most important terms in the energy balance equation. There are a number of methods for estimating these fluxes. In the next section, we will describe the combination method for calculating evapotranspiration. The rate of change in energy storage is often omitted from the energy balance equation in the case of a thin layer of water, soil or canopy, especially on a daily basis. However, this term may have to be considered in the case of a tall vegetation such as forest.

2.4.2 Radiation budget

The surface radiation balance can be written as:

$$R_n = R_{sd} - R_{su} + R_{ld} - R_{lu} \tag{2.3}$$

where R_n is the net radiation, R_{sd} is the shortwave downward radiation, R_{su} is the shortwave upward radiation, shortwave or solar radiation consists of direct and diffuse radiation, R_{ld} is the longwave downward radiation, and R_{lu} is the longwave upward radiation.

The shortwave downward radiation is the radiant flux resulting directly from the solar radiation. It is considerably modified by passage through the atmosphere. The measurement of the short-wave downward radiation can easily be made by using a calibrated pyranometer. In the event that suitable radiation data are not available, the shortwave downward radiation can be estimated from the actual number of bright sunshine hours and the number of daylight hours (Brutsaert, 1982). The shortwave upward radiation or reflected shortwave radiation is a significant term in the radiation balance and is mostly affected by the albedo of the surface. The longwave downward radiation from the atmosphere can be measured radiometrically, or calculated from knowledge of the vertical profiles of temperature and humidity, or estimated from empirical formulae. Due to the fact that necessary data of temperature and humidity profiles are not always available, it is

often convenient to express the longwave downward radiation as a function of meteorological data at screen height and the longwave upward radiation is estimated in a similar way (Brutsaert, 1982):

$$R_{ld} = \varepsilon_a \sigma T_a^4 \tag{2.4}$$

$$R_{lu} = \varepsilon_s \sigma T_a^4 \tag{2.5}$$

with $\varepsilon_a = 1.24(e_a/T_a)^{1/7}$ (2.6)

in which T_a is the air temperature at screen height in K, ε_a is the atmospheric emissivity, e_a is the vapor pressure in hPa, ε_s is the surface emissivity (equal to 0.97), and σ is the Stefan–Boltzman constant.

To calculate net radiation fluxes for different canopy layers and ground surface, we must determine longwave radiation of the atmosphere and that emitted from the canopy layers and ground surface *a priori*. WAVES is a daily time step model and at this level of complexity, it is not unreasonable to assume that the temperatures of the overstorey, understorey and ground surface are equal to air temperature (Ross, 1981). The differences in their longwave radiations are due to the emissivity. This is a good approximation for relatively dense plant stands with non-limiting water supply. WAVES is not a leaf level model and it integrates over whole canopies. The emphasis of WAVES is on water balance not leaf photosynthesis. Therefore, there is no need to separate sunlit and shaded leaves, which will complicate the model and increase the number of parameters. Attempts have been made in WAVES to represent different processes with a consistent level of complexity and this treatment can also be justified against these objectives.

When the radiation balance equation is applied to a plant canopy, the interception, reflection, transmission and absorption of radiation by vegetation have to be dealt with. Ross (1981) has shown that the theoretical equations for direct and diffuse radiation transfer in a plant canopy are complicated and cumbersome. These equations offer a theoretical treatment of the problem, but are of limited use in practice. For this reason, attempts have been made to simplify the theoretical formulae and replaced some of the functions with empirical constants. As a result, some comparatively simple equations have been derived namely semi-empirical formulae. These equations retain the key physical processes governing radiation transfer and contain bulk constants. It has been shown that these equations generally agree well with experimental data (Ross, 1981). Shortwave radiation transfer equations used in WAVES are, in a strict sense, not theoretically derived but semi-empirical. WAVES makes no distinction between direct and diffuse radiation and this can be seen as an approximation. As shown by Ross (1981, 1975) and Monsi and Saeki (1953) that the total solar radiation (direct + diffuse) can be described by a

(1953) that the total solar radiation (direct + diffuse) can be described by a simple exponential equation (Beer's Law).

In WAVES no distinctions are carried between different wave bands visible (PAR) and near-infrared (NIR). The attenuation of shortwave radiation, PAR, and NIR with depth is shown in Fig. 2.2. It can be seen that PAR decreases more rapidly than NIR and the attenuation of short-wave radiation lying between them. This suggests that using a single attenuation coefficient for shortwave radiation does not lead to any errors in irradiance for energy balance purposes. For calculating PAR for plant growth, however, it is important. WAVES currently uses 50% as a fixed PAR ratio.

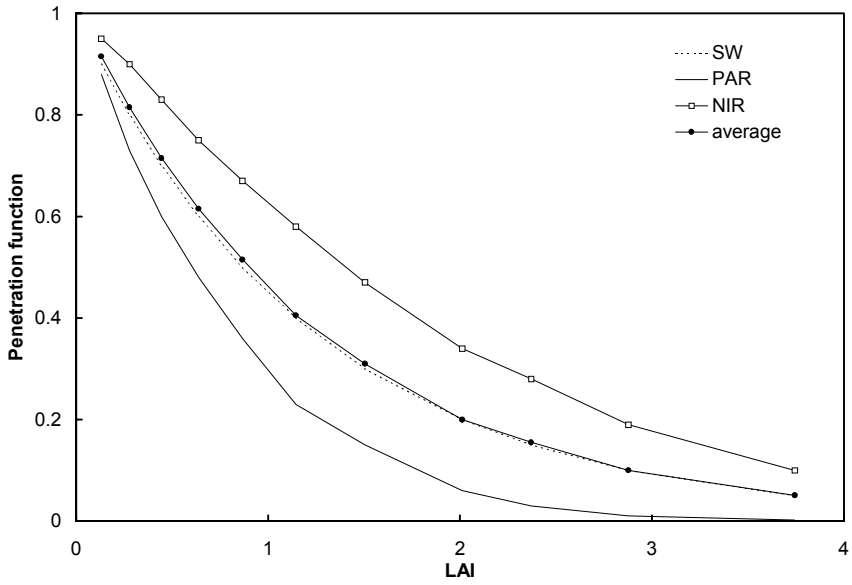

Fig. 2.2. Penetration function for shortwave (SW), visible (PAR), and near-infrared (NIS)

The longwave radiation calculations in WAVES are simplified on the basis of having daily time step, the canopy is a turbid medium, and in isothermal conditions. Ross (1981) has shown that net longwave radiation in a canopy can be dealt with in a similar way as for shortwave radiation (see Appendix A).

We assume all leaves are randomly distributed with horizontal inclination. For a two-layer canopy, the partial coverage of the individual layers can be determined by (Monteith and Unsworth, 1990; Van De Griend and Van Boxel, 1989):

$$\Lambda_1 = 1 - exp(-kLAI_1) = 1 - \tau\ (LAI_1) \qquad (2.7)$$

$$\begin{aligned}\Lambda_2 &= \left[exp(-kLAI_1) - exp(-k(LAI_1 + LAI_2))\right] / exp(-kLAI_1) \\ &= \left[\tau\ (LAI_1) - \tau(LAI_1 + LAI_2)\right] / \tau(LAI_1)\end{aligned} \tag{2.8}$$

and the total coverage of the ground surface follows from equations (2.7) and (2.8) as:

$$\Lambda = 1 - (1 - \Lambda_1)(1 - \Lambda_2) \tag{2.9}$$

where k is the attenuation coefficient for light, and LAI_1 and LAI_2 are the leaf area indices of the overstorey and understorey canopy respectively.

Following Ross (1981), the radiation transfer equations for the soil–vegetation system are given by:

Overstorey

$$R_{sv1}\downarrow = R_{sd}\Lambda_1 \tag{2.10}$$

$$R_{sv1}\uparrow = R_{sd}\alpha_1\Lambda_1 \tag{2.11}$$

$$R_{sn1} = R_{sd}(1 - \alpha_1)\,\Lambda_1 \tag{2.12}$$

$$R_{ln1} = (R_{ld} - R_{lu1})\,\Lambda_1 \tag{2.13}$$

$$R_{n1} = \{R_{sd}(1 - \alpha_1) + (R_{ld} - R_{lu})\}\Lambda_1 \tag{2.14}$$

Understorey

$$R_{sv2}\downarrow = R_{sd}\exp(-KLAz_1)\Lambda_2 \tag{2.15}$$

$$R_{sv2}\uparrow = R_{sd}\alpha_2\exp(-KLAz_1)\Lambda_2 \tag{2.16}$$

$$R_{sn2} = R_{sd}(1 - \alpha_2)\exp(-kLAI_1)\Lambda_2 \tag{2.17}$$

$$R_{ln2} = (R_{ld} - R_{lu2})\,exp(-kLAI_1)\Lambda_2 \tag{2.18}$$

$$R_{n2} = \{R_{sd}(1 - \alpha_2) + (R_{ld} - R_{lu2})\}\exp(-kLAI_1)\Lambda_2 \tag{2.19}$$

Ground surface

$$R_{sg}\downarrow = R_{sd}\exp(-KLAI_t) \tag{2.20}$$

$$R_{sg}\uparrow = R_{sd}\alpha_g\exp(-KLAI_t) \tag{2.21}$$

$$R_{sng} = R_{sd}(1 - \alpha_g)\exp(-kLAI_t) \tag{2.22}$$

$$R_{ln2} = (R_{ld} - R_{lug})\,exp(-kLAI_t) \tag{2.23}$$

$$R_{ng} = \{R_{sd}(1 - \alpha_g) + (R_{ld} - R_{lug})\}\exp(-kLAI_t) \tag{2.24}$$

where the subscripts *d* and *u* indicate downward and upward radiations; *s* and *l* indicate shortwave and longwave radiations; *1, 2*, and *g* represent overstorey, understorey, and ground surface respectively; *n* indicates net shortwave or longwave radiation; α_i is the albedo of a particular surface; and LAI_t is the cumulative leaf area index (LAI_1+LAI_2).

2.4.3 Initial and atmospheric boundary conditions

The initial conditions are generally set up the users. The atmospheric boundary conditions necessary to run WAVES include maximum and minimum daily air temperature, vapour pressure, rainfall, and solar radiation. These data are normally measured at most meteorological stations. If the complete set of data is not available, missing data can be generated using a program called GENCLIM, which is a modification of the program MTCLIM by Running *et al.* (1987). In irrigation areas, information on amount of water applied and frequency can be considered as boundary conditions as well.

Assumptions in energy balance and radiation budget

- It is assumed that all leaves in the canopy are randomly distributed with horizontal inclination. The partitioning of radiation between the canopy layers can be described by Beer's law.
- WAVES makes no distinction between direct and diffuse radiation. No separation is made for different wave bands such as visible (PAR) and near-infrared (NIR).
- The canopy temperatures of the overstorey, understorey, and ground temperature are equal to the air temperature. It is a reasonable approximation for a daily time step model.
- The differences in longwave radiation from different surfaces are due to the emissivity.
- Net radiation, evapotranspiration, and sensible heat flux are the dominant terms in the energy balance equation.

2.4.4 Combination methods

Estimation of evapotranspiration can be based either on aerodynamic approaches or on principles of energy balance. However, both of the methods require information at two or more levels above the surface. In practice, this information is difficult to obtain. To facilitate calculation of evapotranspiration using measurements made at one level only, Penman (1948) first introduced the combination equation by combining aerodynamic and energy balance principles for open water surface or short green vegetation with adequate moisture at all times. The method was further developed by Monteith (1965) who combined aerodynamic and surface (canopy) parameters, and energy balance in an evapotranspiration equation known as the Penman–Monteith

equation for a surface of any type in any state of water supply. In what follows, the derivation of this equation is described.

Fig. 2.3 illustrates schematically the structure of a single-layer model for the partition of available energy into latent heat (evapotranspiration) and sensible heat fluxes. The resistances shown in Fig. 2.3 are stomatal resistance (r_{st}), boundary resistance (r_B), 'eddy diffusive' resistance for heat (r_H), and for water vapour (r_V). The boundary resistance is usually combined with eddy diffusive resistance to form the aerodynamic resistance (r_a).

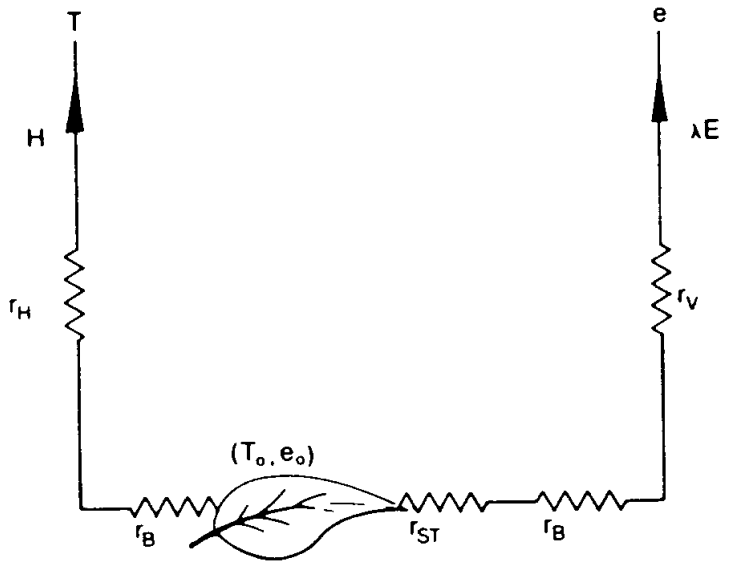

Fig. 2.3. The resistance network used in 'big leaf' model of the vegetation/atmosphere interaction (from Shuttleworth, 1979).

The transfer of sensible heat is given by

$$H = \rho\, c_p \frac{T_s - T_a}{r_a} \tag{2.25}$$

where ρ and c_p are respectively the density and the specific heat of the air at constant pressure, T_s is the surface temperature of the vegetation canopy, r_a is the aerodynamic resistance for sensible heat.

While subject to additional resistance (canopy resistance), the transfer of latent heat flux (evapotranspiration) can be expressed as

$$\lambda E = \frac{\rho\, c_p}{\gamma} \frac{e_s - e_a}{r_a + r_s} \tag{2.26}$$

where γ is the psychrometric constant, e_s is the saturated vapour pressure, e_a is the actual vapour, and r_s is the canopy resistance.

Using an average gradient (Δ) of the saturated vapour pressure versus temperature, one obtains the following expression by combining equations

$$\lambda E = \frac{\Delta(R_n - G) + \frac{\rho c_p}{r_a}(e_s - e_a)}{\Delta + \gamma(r_a + r_s)/r_a} \tag{2.27}$$

Equation (2.27) has a two-term structure suggesting that evapotranspiration has both energy and aerodynamic contributions. This equation is generally known as the Penman–Monteith equation and it is obtained by treating the entire canopy as one 'big leaf' with a bulk stomatal resistance (or canopy resistance) and a bulk aerodynamic resistance. As mentioned, the main feature of the combination equation is that it requires measurements of meteorological variables at one level only.

2.4.5 Evapotranspiration

Fig. 2.4 shows the schematic resistance network for WAVES. The transpiration from the overstorey and understorey canopy layers, and evaporation from the soil can be separately calculated using equations of the Penman–Monteith type

$$\lambda E_1 = \frac{sR_{nv1} + \rho c_p D_a/(r_{a1} + r_{b1})}{s + \gamma\left[1 + r_{c1}/\left(r_{a1} + r_{b1}\right)\right]} \tag{2.28}$$

$$\lambda E_2 = \frac{sR_{nv2} + \rho c_p D_{c1}/(r_{a2} + r_{b2})}{s + \gamma\left[1 + r_{c2}/\left(r_{a2} + r_{b2}\right)\right]} \tag{2.29}$$

$$\lambda E_s = \frac{sR_{ng} + \rho c_p D_{c2}/r_{as}}{s + \gamma\left(1 + r_s/r_{as}\right)} \tag{2.30}$$

where s is the slope of the saturation vapor pressure versus temperature curve, ρ is the air density, c_p is the specific heat of air at constant pressure, D_a is the vapor pressure deficit at reference height, D_{c1} and D_{c2} are the vapor pressure deficit at canopy source height for the overstorey and understorey respectively, r_{a1}, r_{a2} and r_{as} are the aerodynamic resistances between the overstorey canopy source height and a reference level, between the understorey source height and the overstorey canopy source height, and between the soil surface and the air within the understorey canopy, respectively; r_{b1}, r_{b2} are bulk boundary layer resistances of the vegetation elements in the

overstorey and understorey canopies, r_{c1}, r_{c2} are the canopy resistances for overstorey and understorey canopies, r_s is the soil resistance, γ is the psychrometric constant.

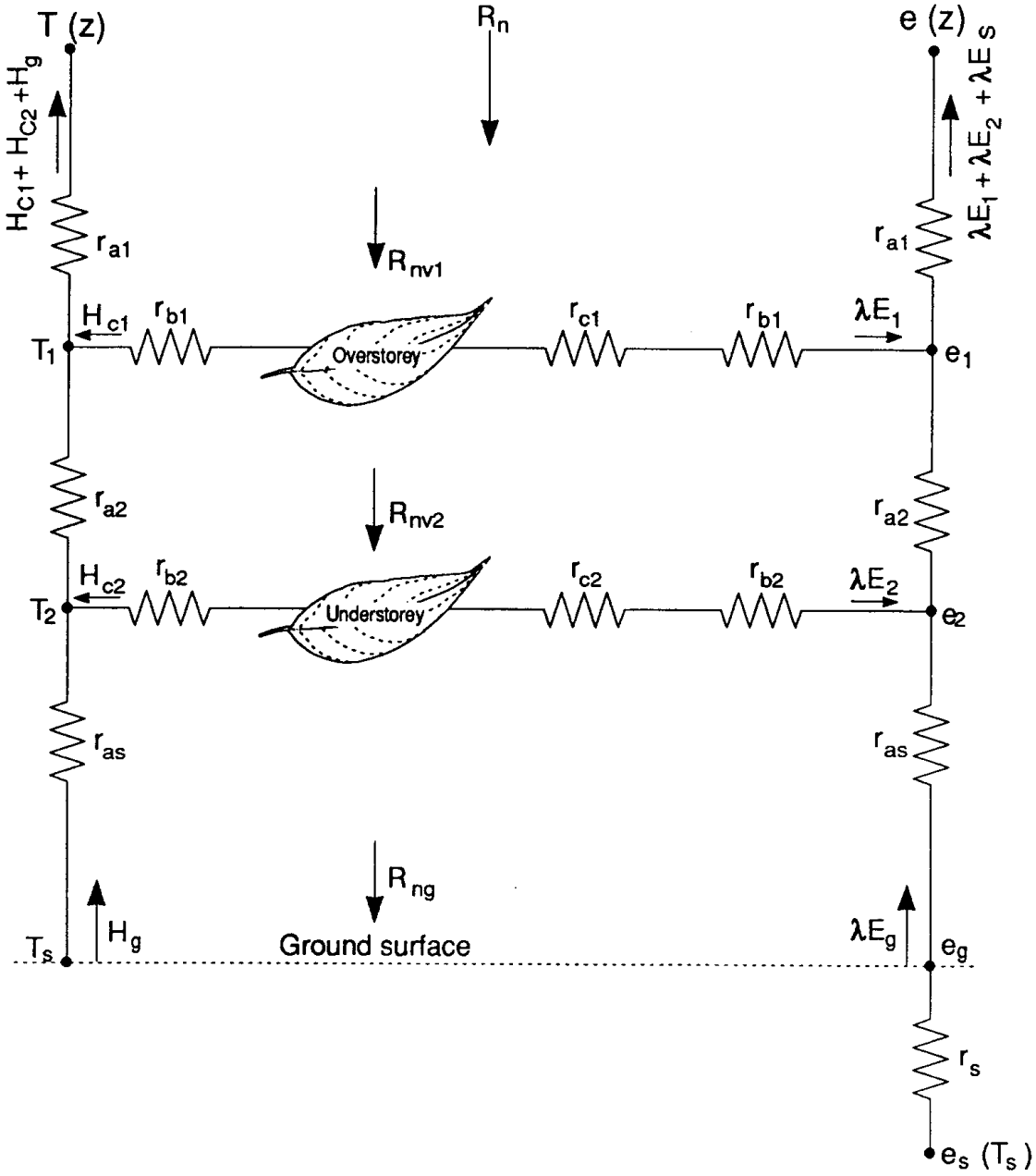

Fig. 2.4. A schematic resistance network for the WAVES model. The symbols are defined in the text.

The ground heat flux is neglected in equation (2.30) because over land surfaces the daily mean value of the ground heat flux is one or more orders of magnitude smaller than the net radiation (Brutsaert, 1982). The boundary layer resistances are generally much smaller than the corresponding aerodynamic resistances, especially if the leaf area index is large (Shuttleworth and Wallace, 1985; Choudhury and Monteith, 1988). As well, the Penman–Monteith equation is found to be rather insensitive to the values of the boundary layer resistance (Shuttleworth and

Wallace, 1985). Based on these observations, the boundary layer resistances are neglected and equations (2.24) and (2.25) can now be expressed as:

$$\lambda E_1 = \frac{sR_{nv1} + \rho c_p D_a / r_{a1}}{s + \gamma(1 + r_{c1} / r_{a1})} \tag{2.31}$$

$$\lambda E_2 = \frac{sR_{nv2} + \rho c_p D_{c1} / r_{a1}}{s + \gamma(1 + r_{c1} / r_{a1})} \tag{2.32}$$

Assumptions in evapotranspiration sub-model

- The canopy can be represented as a 'big leaf' using bulk aerodynamic resistance and bulk canopy resistance.
- Ground heat flux can be neglected over land surface on a daily time step.
- The boundary resistances are much smaller than the corresponding aerodynamic resistances, especially when the leaf area index is large.

2.4.6 Feedback processes between canopy and atmosphere

There exist feedback responses between canopy transpiration, vapour pressure deficit at the canopy surface, and canopy resistance (Jarvis and McNaughton, 1986). For example, an increase in the canopy vapour pressure deficit can affect the canopy resistance, transpiration, and photosynthesis. These primary responses can result in secondary responses. As transpiration increases, the water potential of the mesophyll cells will increase, which has a feedback on canopy resistance. On the other hand, increases in transpiration will also affect canopy vapour pressure deficit. The vegetation canopy and the atmosphere are coupled and the feedback between the processes can be explained using the omega coefficient proposed by Jarvis and McNaughton (1986):

$$D_{ci} = \Omega_{ci} D_{eqi} + (1 - \Omega_{ci}) D_a \tag{2.33}$$

with

$$D_{eqi} = \gamma \varepsilon R_{nvi} \left(\frac{r_{ci}}{c_p} \right) / (\varepsilon + 1) \tag{2.34}$$

$$\Omega_{ci} = (\varepsilon + 1) / (\varepsilon + 1 + r_{ci} / r_{ai}) \tag{2.35}$$

where the subscript i is equal to 1 for the overstorey and 2 for the understorey, Ω_{ci} is the decoupling coefficient, D_{eqi} is the equilibrium saturation deficit, ε is s/γ. Tall vegetation canopies, such as coniferous and deciduous forest, have small omega values and they are well coupled to their environment. As a result, transpiration rate is sensitive to changes in the vapour pressure deficit and canopy resistance. On the other hand, short vegetation canopies, such as crops and pasture, have high omega values and are decoupled from their external environment. The vapour pressure deficit at the canopy surface tends to a equilibrium value and the transpiration is controlled strongly by net radiation.

2.4.7 Aerodynamic resistances

In describing transfer of latent heat and sensible heat fluxes, it is convenient to use the so-called aerodynamic resistance approach. By analogy to Ohm's law in electricity, the aerodynamic resistance can be defined as:

$$aerodynamic\ resistance\ = \frac{concentration\ difference\ of\ an\ entity}{flux\ density\ of\ the\ entity} \tag{2.36}$$

This definition is arrived at by replacing in Ohm's law 'potential difference' by concentration difference and 'current' by flux density. The aerodynamic resistance represents the time in which a unit volume of air exchanges energy with a unit area of surface. It can be calculated from wind speed and surface roughness by assuming a logarithmic wind profile (Thom, 1975). The dimension of aerodynamic resistance is (velocity)$^{-1}$.

In WAVES, separate aerodynamic resistances are calculated for the overstorey, understorey, and ground surface. These resistances vary with wind speed and roughness length. Above the overstorey, the wind speed profile is assumed to be logarithmic and mean wind speed decreases exponentially through the overstorey and understorey. The aerodynamic resistance between overstorey canopy source height and reference level is determined by Monin–Obukhov surface layer similarity theory:

$$r_{a1} = \left\{\ln\left[\left(z-d\right)/z_{01}\right]-\psi\right\}^2/\left(k^2u\right) \tag{2.37}$$

where u is the wind speed at the reference height z, k is the von Karman constant, z_{01} is the roughness length of the overstorey canopy, d is the zero plane displacement, ψ is the atmospheric stability function. It is reasonable to approximate the roughness length and the zero plane displacement as fractions of the canopy height (Brutsaert, 1982; Monteith and Unsworth, 1990). The effect of atmospheric stability on the aerodynamic resistance is not considered in WAVES.

The aerodynamic resistance between understorey source height (z_2) and overstorey canopy source height (z_1+z_2) is defined as:

$$r_{a2} = \int_{z_2/2}^{(z_1+z_2)/2} dz / K(z) \tag{2.38}$$

The increase of the diffusivity $K(z)$ with height is approximately exponential and the relation may be expressed by (Thom, 1975):

$$K(z) = K(H)\, exp[-\alpha(1 - z/H)] \tag{2.39}$$

where H is the height of the overstorey canopy, α is an attenuation coefficient, and

$$K(H) = k^2 (H - d)u / \ln[(z - d)/ z_{01}] \tag{2.40}$$

The following expression can be obtained from equations (2.38) and (2.39):

$$r_{a2} = \frac{H \exp(\alpha)}{\alpha K(H)} [\exp(-\alpha z_2/2H) - \exp[-\alpha (z_1 + z_2)/2H]] \tag{2.41}$$

The aerodynamic resistance between the soil surface and the air within the understorey canopy is defined in a similar way as for r_{a2}:

$$r_{as} = \frac{H \exp(\alpha)}{\alpha K(H)} [\exp(-\alpha z_{02}/2H) - \exp[-\alpha z_2/2H]] \tag{2.42}$$

where z_{02} is the roughness length for the understorey canopy.

2.4.8 Surface resistance to soil evaporation

The surface resistance r_s in WAVES is assumed to be zero until the water content in the first node depth drops below the air dry soil moisture content; commonly called stage 1 soil evaporation (Ritchie, 1972). The soil resistance during stage 2 evaporation is determined by the method of Choudhury and Monteith (1988):

$$r_s = \tau l / (p D_m) \tag{2.43}$$

where p is the porosity of the soil, D_m is the molecular diffusion coefficient for water vapour, τ is a tortuosity factor and l is the depth of the air-dry soil layer. The depth of the soil layer is determined dynamically by the finite difference Richards' equation of water content, by finding how deep below the surface the soil is at air-dry potential.

Assumptions in aerodynamic resistances

- The conventional log-linear wind profile is assumed to be valid above the canopy.
- It is assumed that on a daily time-step the effect of atmospheric stability on the aerodynamic resistance is not significant.
- The vertical profile of wind speed within a plant canopy is assumed to be similar to that of eddy diffusivity, which tends to decrease exponentially with depth.

2.5 Soil water dynamics and runoff

2.5.1 Richards' equation

Movement of water through a soil matrix is governed Darcy's law:

$$q = -k\frac{dH}{dz} = -k\left(1 - \frac{\partial \psi}{\partial z}\right) \tag{2.44}$$

where q is water flux density, H is the total hydraulic head, which is the sum of the gravitational potential, and the matric potential ψ, z is the vertical distance from the soil surface downward, and k is the hydraulic conductivity. Darcy's law states that water flux density is proportional to the hydraulic gradient, which is the driving force. The proportionality factor k is generally known as the hydraulic conductivity.

Darcy's law assumes that the soil is homogeneous and isotropic so that the hydraulic conductivity is uniform and has no dependence on the direction of water movement. It further assumes that the soil is isothermal, isotropic, incompressible. Darcy's law is valid for most range of flow velocities observed in soil.

The continuity equation of water flow can be expressed as:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} + S \tag{2.45}$$

where θ is volumetric water content, t is time, and S is a source/sink term.

Combining equations (2.44) and (2.45) gives the general flow equation:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} + S = -\frac{\partial}{\partial z}\left(k - k\frac{\partial \psi}{\partial z}\right) + S \tag{2.46}$$

Equation (2.46) is the fundamental mixed form of the Richards equation.

For analytic solutions and common numerical solution techniques, equation (2.46) is often cast with a single dependent variable. The ψ-based form was used by Richards and has traditionally been seen as mandatory to treat soils that may become saturated. A criterion for choosing a form of the equation is the need to minimise nonlinearity in time and space. Redinger *et al.* (1984) and Ross and Bristow (1990) minimised nonlinearity in time by using θ on the left hand side of the equation, and reduced nonlinearity in space by using the Kirchhoff transform in the flux term:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial}{\partial z}\left[K - \frac{\partial U}{\partial z}\right] + S \tag{2.47}$$

where θ is the soil water content, t is time, K is the hydraulic conductivity and U is the Kirchhoff transform variable defined as:

$$U = \int_{-\infty}^{\psi} K d\Psi \tag{2.48}$$

where Ψ is the matric potential of the soil water.

The sink term S which accounts for the rate of water extraction from the soil is modelled as follows. For evapotranspiration, the water is extracted from the entire root zone according to a weighting function which depends on the rooting density and availability of soil moisture. The model first calculates total root water uptake potential (*RWUP*):

$$RWUP = \sum_{i=1}^{n} r_i(\psi_{l\,max} - \psi_i)\Delta z_i \tag{2.49}$$

where r_i is the root biomass, ψ_{lmax} is the most negative water potential under which roots can still extract water (its value can be estimated from the minimum leaf water potential), ψ_i is the total potential (matric and osmotic), Δz_i is the depth of the discretised soil layer, and subscript i represents the soil layer. Then water uptake from each discretised soil layer is determined by

$$S_i = \frac{r_i(\psi_{l\,max} - \psi_i)\Delta z_i}{RWUP}\lambda E \tag{2.50}$$

The calculated water uptake from equation (2.50) is balanced with the available water which can be obtained by

$$S'_i = (\theta_i - \theta_d)\Delta z_i \tag{2.51}$$

The actual water uptake from each discretised soil layer is determined by

$$S_i = min\{S_i, S'_i\} \quad (2.52)$$

This method does not require a functional form of the equation for root water extraction and is suitable for investigating the impact of salinity on water extraction and plant growth.

2.5.2 Initial and boundary conditions

The initial conditions can be specified by the user as matric potential at specified depths. The lower boundary condition in equation (2.47) is defined as a fraction (β) of hydraulic conductivity: β is set to 1 when the lower boundary condition is free drainage, and β is set equal to zero for an impermeable boundary. For the upper boundary condition, three conditions may exist. Firstly, all rainfall or evaporation can be transmitted through the soil surface; in this case the flux of rain or evaporation is set. Rainfall may exceed the capacity of the soil either because the rate is too high (Hortonian runoff), or the soil becomes saturated (Hewlett or Dunne runoff); in this case the surface is set to a constant saturated potential, and all rainfall in excess becomes runoff. Finally evaporation may be limited by dry soil; in this case the soil surface is set to a constant air-dry potential and the flux passing the surface node is returned as the daily evaporation.

To solve Richards' equation, an analytical soil model of Broadbridge and White (1988) is used to describe the relationships among water potential, volumetric water content and hydraulic conductivity. This soil model has five parameters, including the saturated hydraulic conductivity, the volumetric soil moisture content at saturation, air-dry volumetric water content, the soil capillary length scale which is a function of sorptivity, and a soil structure parameter. The Broadbridge and White soil model can realistically represent a comprehensive range of soil moisture characteristics, from the highly nonlinear associated with a well developed capillary fringe, to the weakly nonlinear associated with highly structured soils and macropores. The model is subject to two levels of dimensionless scaling that lead to simple rules for guaranteed numerical performance (Short *et al*., 1995).

Assumptions in the Richards' equation

- The soil is rigid, incompressible, non-hysteretic, and isothermal.
- Water flow is via the soil matrix only, and not via macropores and larger preferred pathways.
- The soil is isotropic so that the saturated hydraulic conductivity of the soil is used in the formulation of Darcy's law for lateral movement.
- Vapour flow within the soil is not modelled explicitly, but included in the soil hydraulic model if possible.

2.5.3 Runoff generation

Overland flow can be generated from the excess of rainfall intensity over soil infiltrability, and the occurrence of precipitation over saturated surfaces. Both of the mechanisms are considered explicitly in WAVES. Water tables may develop anywhere within the soil profile. If non zero topographic slope is specified as input, then lateral subsurface flow occurs via the saturated water table and is described by Darcy's law.

2.6 Solute Transport

2.6.1 Theory

Soil water contains dissolved salts which may range from 5 mg per litre in rainwater to as high as 10 000 mg per litre in drainage from saline soil (Hillel, 1980). Solute transport in the soil is governed by three mechanisms, namely convection, diffusion, and hydrodynamic dispersion.

The convection of soil water carries with it a convective flux of solutes q_{sc} , which is proportional to their concentration c:

$$q_{sc} = qc = -c(k\, dH/dz) \tag{2.53}$$

The average apparent velocity $\bar{v}$ of the flowing solution can be calculated as:

$$\bar{v} = q/\theta \tag{2.54}$$

where θ is volumetric water content. Combining eq. (2.53) and (2.54) yields:

$$q_{sc} = qc = \bar{v}\,\theta\, c \tag{2.55}$$

Diffusion processes commonly occur within gaseous and liquid phases. The net effect is a tendency to equalise the spatial distribution of diffusible components in any mixed or multicomponent fluid. As such, diffusion processes are extremely important in the soil. If solutes do not happen to be distributed uniformly throughout a solution, concentration gradients will exist and solutes will tend to diffuse from high concentration to low concentration. In bulk water at rest, the rate of diffusion q_{sd} is related by Fick's law to the gradient of the concentration c:

$$q_{sd} = -D_s(\theta)dc/dz \tag{2.56}$$

in which D_s is the diffusion coefficient in the soil and dc/dz is the concentration gradient.

Equation (2.56) can only describe steady-state diffusion processes. For transient-state processes, we must invoke the mass conservation law, as formulated in the continuity equation. Let us assume that there are no sources or sinks for the diffusing solute in the soil and consider a rectangular volume element of soil which contains a liquid phase and which is bounded by two parallel square planes, of area A, separated by a distance Δz. The amount of solute diffusing through one of these planes into the volume element per unit time is Aq_{sd}, and the amount diffusing out the volume element through the second plane is $A[q_{sd} + (\partial q_{sd}/\partial z)\Delta z]$. The rate of accumulation of the solute in the volume elements is $-A(\partial c/\partial t)\Delta z$, where $\partial c/\partial t$ is the time rate of change of concentration. Thus,

$$A\left(\partial c \big/ \partial t\right)\Delta z = A\left[q_{sd} + \left(\partial q_{sd} \big/ \partial z\right)\Delta z\right] - Aq_{sd} \tag{2.57}$$

Combining this equation with eq. (2.56), we obtain a second-order equation as follows:

$$\frac{\partial c}{\partial t} = \partial\left(D_s \, \partial c \big/ \partial z\right) \big/ \partial z \tag{2.58}$$

The motion of any inhomogeneous solution in a porous body brings another process which differs from diffusion in its mechanism but which tends to produce an analogous or synergetic effect, which is to mix and eventually even out the concentration or composition difference between different portions of the flowing solution. This process is called *hydrodynamic dispersion*. It results from the microscopic nonuniformity of flow velocity in the soil's conducting pores. Mathematically, hydrodynamic dispersion is formulated in a manner analogous to the formulation of diffusion as give by eq. (2.56) and (2.58), except that, instead of a diffusion coefficient a *dispersion coefficient* is introduced. This coefficient, which we will designate D_h, has been found to depend more or less linearly on the average velocity:

$$D_h = a\bar{v} \tag{2.59}$$

with a an empirical parameter.

Because of the similarity in effect (though not in mechanism) between diffusion and dispersion, it is tempting to assume the two effects to be additive. Accordingly, the diffusion and dispersion coefficients are often combined into a single term, namely *the diffusion-dispersion coefficient* D_{sh}, which is a function of both the fractional water content and the average velocity:

$$D_{sh}(\theta, \bar{v}) = D_s(\theta) + D_h(\bar{v}) \tag{2.60}$$

To take into account the three mechanisms of solute movement, we can combine these equations to obtain:

$$q_s = \bar{v}\,\theta\,c - [D_s(\theta)\,dc/dz + D_h(\theta)\,dc/dz] \tag{2.61}$$

Since in practice the diffusion and dispersion phenomena can not be separated, the foregoing equation is usually written in the form:

$$q_s = \bar{v}\,\theta\,c - D_{sh}(\theta)\,dc/dz \tag{2.62}$$

Here q_s is the total mass of a solute transport across a unit cross-sectional area for soil per unit time.

For transient-state processes, we once again invoke the continuity condition, which for combined convective-diffusive-dispersive transport can be written:

$$\partial(c\theta)/\partial t = -\partial q_s / \partial z \tag{2.63}$$

For the rate of change of the solute mass present in a volume element of soil equal to the difference between the incoming and outgoing fluxes of the solute for that volume element.

Combining eq. (2.61) and (2.62), one obtains:

$$\partial(c\theta)/\partial t = -\partial(\bar{v}\theta c)/\partial z + \partial\left(D_{sh}(\theta)dc/dz\right)/\partial z \tag{2.64}$$

or

$$\partial(c\theta)/\partial t = -\partial\left(qc\right)/\partial z + \partial\left(\theta\,D_s(\theta)dc/dz\right)/\partial z \tag{2.65}$$

where q is the flux of water and $D_s(\theta)$ is the diffusion-dispersion coefficient of solute in the soil.

2.6.2 Initial and boundary conditions

The initial solute concentration in the profile is specified in an input file, in units of kg/ℓ. There are two other concentrations required: the solute concentration in the rain (and flood) water, and in the groundwater. When there is leakage from the soil column, the concentration of solute at the bottom node is used to calculate the amount of solute leached, and if there is uptake from the groundwater, then the solute concentration in the groundwater is used to calculate the amount of solute accumulated.

Assumptions in solute transport

- The solute is conservative, i.e. the solute is non-volatile, does not adsorb to, or desorb from, the soil matrix, and that solute concentration does not affect soil hydraulic properties.
- The soil water solution is perfectly mixed, is completely mobile, and is at concentrations that do not cause a precipitate.
- The solute is not taken up by root water extraction, or lost by soil evaporation.

2.7 Carbon allocation and plant growth

2.7.1 Integrated rate methodology

Plant growth is a complex process and dependent on a number of factors such as light, water, and nutrients. Attempts have been made to develop models that incorporate detailed biochemical and biophysical processes (Farquhar *et al.*, 1980, Collatz *et al*., 1991). An important aspect of such models is their value in helping to improve our process understanding, but they have limited application in practice because of the model complexity and data requirements. On the other hand, purely empirical plant growth models have been developed by statistical means (Hunt, 1982) and they contain few physically based functions to relate input to output. Within the range of data analysed, such a model may be highly successful. However, these models can not be used to make any predictions beyond the range of actual experience.

In WAVES, an intermediate position between the above two approaches was taken. The method defines a potential growth rate and modifies it by the availability of resources using an integrated rate methodology (IRM). The assimilation rate A_i is expressed as:

$$A_i = A_{\max} r_i \tag{2.66}$$

where A_{max} is the maximum carbon assimilation rate (see Table 2.1) and r_i is the relative carbon assimilation rate given by Wu *et al.* 1994 as:

$$r = \frac{1 + w_H + w_N}{1/\eta_T \chi_L + w_H / \chi_H + w_N / \chi_N} \tag{2.67}$$

where w_W is the weighting of water relative to light, w_N is the weighting of nutrients relative to light, χ_H, χ_N, and χ_L are the relative resource availabilities for water, nutrient, and light respectively, and η_T is the modifier of light availability due to temperature. The availability of light is

determined from intercepted radiation and air temperature, the availability of nutrients is an explicit constant for a growing season. The availability of water is a depth-weighted integral of the soil matric and osmotic potentials in the root zone. This value is made relative by dividing by the maximum soil water potential at which the plants can extract water, and subtracting from 1.

A normalised index of light availability (χ_L) is calculated as the ratio of the average PAR per unit leaf area to the light saturation value of a unit leaf:

$$\chi_L = \frac{R_P}{R_{max}} \tag{2.68}$$

where R_p is the average PAR per unit leaf area of the canopy calculated from available energy, R_{max} is the light availability at which maximum growth is obtained. The temperature modifier (η_T) is set to 1 at the optimum growth rate temperature (T_{opt}) and 0.5 at the half optimum temperature (T_h):

$$\eta_T = \exp\left(\frac{(T_a - T_{opt})^2}{(T_{opt} - T_h)^2} * \ln(0.5)\right) \tag{2.69}$$

where T_a is the average daily temperature. The availability of soil water for transpiration is estimated from the total soil water potential, *i.e.* the sum of the matric (ψ) and osmotic (π) potential of the soil water. The matric potential is calculated directly from the moisture characteristic function, whilst the osmotic potential is estimated from the sodium chloride concentration of the soil water. This assumes that the soil water is chloride dominated. The total soil water potential of each soil layer is normalised using the lowest soil water matric potential (ψ_{wilt}) against which the plant can transpire. A weighting factor adjusts for different effects of osmotic and matric potential on water availability. For the *i*-th soil layer:

$$\chi_{wi} = 1 - \frac{\psi_i + w_{osm}\pi_i}{\psi_{wilt}} \tag{2.70}$$

where w_{osm} is an osmotic potential weighting factor representing the ratio of the lowest matric potential to the lowest osmotic potential against which transpiration can occur. The osmotic potential of the soil water is calculated by

$$\pi_i = -2C_{NaCl}RT \tag{2.71}$$

where C_{NaCl} is the molarity of sodium chloride in the soil water, R is the universal gas constant, T is temperature. The index of water availability to the plant is estimated as the water uptake weighted average soil water availability:

$$\chi_w = \frac{\sum_{i=1}^{n} W_{ri}\chi_{wi}}{\sum_{i=1}^{n} W_{ri}} \tag{2.72}$$

where W_{ri} is a layer weighting factor and assumed to be related to the relative amount of root carbon in each soil layer, n is the number of soil layers in the potentially active root zone. The relative availability of nutrients (χ_N) is an input parameter (0–1).

The zero to one scalar r multiplied by the vegetation's maximum carbon assimilation rate gives the daily gross assimilation. After meeting growth respiration, any remaining carbon is dynamically allocated between leaf, stem and root carbon pools. If the plant has a net deficit during a day, the carbon is removed from these pools. Because WAVES concentrates only on the hydrological aspects of plants growth, e.g., leaf area index, the model does not fill grain or otherwise account for reproductive material of crops, grass, or trees.

The IRM framework provides an explicit means of integrating the net effect of multiple limiting factors. It also provides a means of taking into account not only the relative availability of resources, but also other possible factors such as salinity. IRM retains a mechanistic representation of relative plant growth response to resources availability in the form of its enzyme kinetics origins. A fully detailed description of the plant growth model in WAVES, and how to calculate the weighting factors for (2.67), can be found in Hatton *et al.* (1992) and Wu *et al.* (1994).

2.7.2 Carbon allocation and plant growth

The simulated carbon is partitioned to leaves, stems, and roots on a daily basis. The partitioning coefficients depend on both genotype and environment. The amount of leaf, stem, and root carbon allocated is reduced by growth and maintenance respiration and mortality rates. The daily carbon increment is given by:

$$\Delta C_L = n_L Y_L (A_i - C_L R_L) - C_L M_L \tag{2.73}$$

$$\Delta C_{S,R} = n_{S,R} Y_{S,R} \{(A_i - C_L R_L) - C_{S,L} R_{S,R}\} - C_{S,R} M_{S,R} \tag{2.74}$$

where subscripts L, S, and R refer to leaves, stems, and roots respectively, C is the carbon content of the biomass, n_L is the proportion of net canopy assimilation allocated to the leaves, n_S, and n_R

are the proportions allocated to stem and root growth of the remaining carbon assimilation, Y are respiration coefficients which account for conversion of assimilated carbon to biomass, R are maintenance respiration coefficients and M are mortality coefficients. Note that leaf maintenance respiration is subtracted before carbon is allocated to stems or the roots and maintenance respiration is deducted before assimilation is used for growth.

Carbon allocated to leaves is assumed to increase leaf area by an amount determined by the specific leaf area and the carbon allocated to roots is distributed amongst soil layers using the same weight function used for the soil water availability (2.72).

2.7.3 Canopy resistance

The canopy resistance of vegetation plays a major role in partitioning the available energy into evapotranspiration. It is governed by stomatal function and is primarily dependent on the photosynthetic rate and environmental factors, such as light interception, temperature, and vapour pressure deficit. For vegetation transpiration, the canopy resistance for the overstorey and the understorey is calculated using the empirical model of Ball *et al*. (1987) as modified by Leuning (1995):

$$r_{si} = (g_{si})^{-1} = \{g_0 + g_1 A_i / [(c_{si} - \Gamma)(1 + D_{ci} / D_{co)}]\}^{-1} \tag{2.75}$$

where the subscript i equals to 1 for the overstorey and 2 for the understorey, g_{si} is the leaf stomatal conductance, g_0 is a residual stomatal conductance, g_1 is an empirical coefficient, C_{si} is CO_2 mole fraction of the air at the canopy surface, Γ is the CO_2 compensation point, D_{ci} is the vapor pressure deficit at the canopy surface, D_{co} is an empirical coefficient. The values of these coefficients are given in Table 2.1.

The leaf stomatal conductance model (2.75) can be integrated to yield canopy conductance model (Sellers *et al.*, 1992a):

$$r_{ci} = (g_{ci})^{-1} = \{g_0 LAI + g_1 A_i / [(c_{si} - \Gamma)(1 + D_{ci} / D_{co)}]\Pi\}^{-1} \tag{2.76}$$

with

$$\Pi = (1 - exp(-kLAI))/k \tag{2.77}$$

The canopy resistance for water vapour can be estimated from (2.76) by considering the gaseous pathway and adjusting for the diffusivity of CO_2 and water vapor:

$$r_{wi} = r_{ci} / [1.6(1 + \chi_w / w\chi_L\chi_T)] \tag{2.78}$$

Assumptions in carbon balance and plant growth

- It is assumed that plants are greedy, constantly germinate, and attempt to fully stock an area if possible, i.e., when conditions are good, the vegetation attempts to close the canopy and develop a maximum leaf area.
- Carbon assimilation rate is controlled mainly by the availability of light, water, and nutrients.
- It is assumed that temperature and CO_2 concentration are invariant with depth in the canopy.

2.8 Parameter Estimation

As a physically based model, WAVES involves a number of parameters to describe processes occurring in the soil–vegetation–atmosphere system. Generally, two types of model parameters can be identified: *physical parameters* and *process parameters*. Physical parameters are well-defined and physically measurable properties such as albedo, saturation water content, hydraulic conductivity, saturation light intensity, etc. Process parameters are used to represent processes that are not well-defined; these parameters are not directly measurable properties such as capillary length, root mortality coefficient, relaxation coefficient, etc. A necessary step in applying the model is to determine the values of these model parameters or constants for the site under consideration. Parameter estimation may have at least as great an effect on the accuracy of the model results as the intrinsic accuracy of the model itself, e.g. even a model that perfectly describes a system will produce the wrong answers if the parameters are wrong. Three techniques are employed for estimating parameter values: *direct estimation, knowledge-based estimation,* and *model calibration*. Direct estimation or measurement of parameter values from field observation is logically the best approach, but it requires well-defined parameters which have a physical meaning and relate to the biophysical processes. Parameter values can also be estimated on the basis of published data, knowledge of likely values, and previous experience. The third technique is to calibrate the model against experimental data, or to fit model parameters. This is usually necessary in the use of any model, but must be treated with caution and the number of fitted parameters should be kept to a minimum. Calibrated parameters are often affected by deficiencies in the model structure and conceptualisation, as well as by measurement errors. In what follows, a detailed description of parameter estimation for WAVES is given.

The input variables and principal parameters used in WAVES are summarised in Table 2.1. WAVES requires three types of data: a) meteorological data; b) soil parameters; and c) vegetation parameters.

2.8.1 Meteorological variables

The meteorological data provide the atmospheric boundary conditions necessary to drive WAVES. They include maximum and minimum daily air temperature, daily average vapour pressure deficit, rainfall, rainfall duration, daily solar radiation, and wind speed. Some or all of these data are available from weather stations, and with only the temperatures and rainfall, realistic estimates of the other data can be made.

2.8.2 Soil parameters

The soil data required is knowledge of the soil layering, and the parameters that describe the relationships between ψ (soil water potential), θ (volumetric water content), and K (hydraulic conductivity). Before WAVES can be used, users need to estimate the parameter values of the soil water retention and hydraulic conductivity functions. In idea cases, these parameter values can be obtained by fitting experimental data to specified functions (*e.g*. Cresswell and Paydar, 1996). When no such data are available, the parameter values can be estimated based on particle size distribution and textural description Salama *et al.,* 1999). In some cases, inverse modelling techniques can also be used (Ross, 1993, Hume *et al.,* 1996). WAVES reads soil hydraulic properties from a table generated by an external program and thus allows the users to choose which soil water retention and hydraulic conductivity functions they want to use.

2.8.3 Vegetation parameters

WAVES requires 22 vegetation parameters to describe canopy energy and carbon balance, and interactions between soil and vegetation. Most of these parameters can be measured directly or taken from plant physiological literature, with only a few remaining for fitting, or adaptating to local conditions. Canopy albedo (α_v) is considered as a constant for a given vegetation and can be obtained from literature (Monteith and Unsworth, 1990). Roughness length (z_{ov}) affects the turbulence transport of water between canopy and the atmosphere. It is generally estimated from the height of vegetation canopy (Brutsaert, 1982). Light extinction coefficient (K) depends on the geometry of radiation with respect to the architecture of canopy. In practice, values of light extinction coefficient can be determined by measuring the attenuation of radiation in a plant canopy or taken from literature (Monteith and Unsworth, 1990). Rainfall interception coefficient (K_r) defines the maximum rainfall interception for a given vegetation canopy. Dunin *et al*.(1988) and Leuning *et al.* (1994) reported values for Eucalyptus and wheat crops. Specific leaf area (SLA) is used to convert leaf carbon to leaf area and its value can be found in Charles-Edwards (1982) for crop species and Raison *et al.* (1992) and Read and Busby (1990) for forests. Maximum assimilation rate of carbon (A_{max}) is a species dependent parameter and Collatz *et al.* (1991)

and (1992) reported values for C_3 and C_4 plants. It should be mentioned that the unit for A_{max} has changed from μmol m^{-2} s^{-1} to kg C m^{-2} d^{-1}. The availability of water for plant transpiration decreases as the matric potential of the soil water decreases. The maximum plant available water potential (LWP_{max}) is 100–150 m for most plants (Hillel, 1971, Marshall *et al.*, 1996). Drought-tolerant plants can use soil water at a lower matric potential over long periods. Saturation light intensity (L_{max}) defines the irradiance beyond which photosynthesis is not limited by light and its value varies between 1000 to 2000 μmol m^{-2} d^{-1}(Monteith, 1979, Wu *et al*., 1994). Maximum rooting depth (RD_{max}) depends on plant species and soil properties. It can be measured directly in the field (Incerti and O'Leary, 1990) or obtained from literature (Canadell *et al.,* 1996, Jackson *et al.,* 1996). Temperatures when growth is optimum and half-optimum were estimated for a range of plant species by Slavich et al (1998) based on Verteeg and Keulen (1986) and Larcher (1980). Respiration coefficients for crops vary from 0.0036 to 0.0095 kg C kg^{-1} C d^{-1} and are approximately 0.00084 kg C kg^{-1} C d^{-1} for evergreen trees (Larcher, 1980). Most of the parameters listed in Table 2.1 may be considered constants and set at representative values. When specific information is available, these parameters can be adjusted, and this is part of model calibration exercise. It should be emphasised that most of the parameters used in WAVES are well defined and can be measured directly. However, others are less well defined and have to be determined or inferred indirectly from field measurements or other source of information.

Table 2.1. List of input meteorological variables and principal model parameters of WAVES

Definition	Symbol	Unit	Parameter estimation
a. Meteorological inputs			
Total solar radiation	R_s	kJ m^2 day^{-1}	field measurements
Maximum daily temperature	T_{max}	°C	field measurements
Minimum daily temperature	T_{min}	°C	field measurements
Mean daily vapour pressure deficit	D_a	hPa	field measurements
Total daily precipitation	P	mm	field measurements
b. Soil parameters			
Soil albedo	α_s	—	Brutsaert (1982)
Soil roughness length	z_{0s}	m	Brutsaert (1982)
Saturated hydraulic conductivity	K_s	m day^{-1}	Clapp and Hornberger (1978)
Volumetric water content at saturation	θ_s	cm^3 cm^{-3}	Clapp and Hornberger (1978)
Air-dry soil moisture content	θ_d	cm^3 cm^{-3}	Clapp and Hornberger (1978)
Capillary length scale	λ_c	m	Estimated based on soil texture
Shape parameter	C	—	Estimated based on soil texture
c. Vegetation parameters			
Canopy albedo	α_v	—	Brutsaert (1982)
Rainfall interception coefficient	K_r	m day^{-1} LAI^{-1}	Vertessy *et al*. (1996)
Light extinction coefficient	K	—	Monteith and Unsworth (1990)
Specific leaf area	SLA	—	Measured
Maximum assimilation rate of carbon	A_{max}	μmol m^{-2} s^{-1}	Collatz *et al*. (1992)
Maximum plant available water potential	LWP_{max}	m	Hillel (1971)
Saturation light intensity	L_{max}	μmol m^{-2} day^{-1}	Wu *et al* (1994)
Maximum rooting depth	RD_{max}	m	Measured (soil depth limited)
Canopy roughness length	z_{ov}	m	Brutsaert (1982)
Residual stomatal conductance	g_0	mol m^{-2} s^{-1}	Leuning (1995)
Slope parameter of the conductance model	a_1	—	Leuning (1995)
CO_2 mole fraction of the air	C_s	μmol mol^{-1}	Measured
CO_2 compensation point	Γ	μPa Pa^{-1}	Leuning (1995)
Temperature when growth is optimum	T_{opt}	°C	
Temperature when growth is half optimum	T_h	°C	
Leaf maintenance respiration coefficient	R_l	kg C kg^{-1} C d^{-1}	Running and Coughlan (1988)
Stem maintenance respiration coefficient	R_s	kg C kg^{-1} C d^{-1}	
Root maintenance respiration coefficient	R_r	kg C kg^{-1} C d^{-1}	
Leaf Mortality coefficient	M_l	kg C kg^{-1} C d^{-1}	
Stem Mortality coefficient	M_s	kg C kg^{-1} C d^{-1}	
Root Mortality coefficient	M_r	kg C kg^{-1} C d^{-1}	
Vapour pressure coefficient	D_{co}	hPa	Leuning (1995)