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The water balance of a first order catchment in the montane grasslands of South Africa

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Abstract

Measurements of daily actual evaporation, streamflow, rainfall and soil water storage were measured for a two-year period in a 97 ha grassland catchment located in the Natal Drakensberg. Actual evaporation was measured using the Bowen ratio energy balance technique. Annual precipitation and discharge data were also used to estimate evaporation over a number of years. Results show that in normal years precipitation is equally split between evaporation and streamflow. In dry years evaporation is the dominant component of the water balance. The data were used to develop simple expressions to calculate annual streamflow and evaporation. Good agreements between actual and modelled trends of streamflow using the ACRU hydrological model were found ($r = 0.80$). In general ACRU underestimated streamflow by 15%. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: Evaporation; Streamflow; Bowen ratio

1. Introduction

South Africa has a highly variable rainfall and hydrology. As the rapidly expanding population puts pressure on the water resources, water deficits are becoming severe. Because of the variation in rainfall over the country it is necessary to move water from high rainfall catchments to areas where water resources are poor. Optimising the yield of water from catchments is therefore becoming an important issue in catchment management.

The major catchment areas for South Africa's water resources are covered by natural grasslands that occupy approximately 29% (350,000 km^2) of the country. These areas are relatively cool, and have an annual rainfall between 600 and 1200 mm. These areas coincide with the afforestation zones for commercial exotic tree species, whose site requirements for growth are most limited by the availability of soil water. Given that strategic planners envisage the establishment of 16,000 ha of forests per annum to supply commercial timber (Anon, 1996), and the increased demand for timber products resulting from the impact of the RDP (reconstruction and development programme), there can be no doubt that afforestation will have a further impact on the country's already scarce water supplies. Most of the research into predicting hydrological change due to afforestation in South Africa has been to quantify the water use of the commercial trees (Dye, 1996). Very little research has been directed at determining the water loss from natural communities. To quantify the effects of afforestation and agriculture, it is necessary to have information on the evaporation losses from natural vegetation to provide a management baseline. Large areas of *Themeda triandra* grasslands, growing in the

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high yielding water production areas of South Africa are currently being converted into commercial forestry plantations. Henrici (1943) has been the only worker in South Africa to attempt evaporation measurements in these grasslands and these were plant based. Measurements of evaporation in *Themeda australis* grassland using lysimeters and the Bowen ratio technique have been carried out in Southern Australia (Dunin and Reyenga, 1978). The water use of the grassland vegetation can be estimated by monitoring the difference between the annual precipitation and measured streamflow (Bosch, 1979). Paired catchment studies have also been used to determine the water use of different vegetation types (Bosch, 1992). While these studies enable broad estimates of water loss, they do not provide insights into the hydrological processes within catchments.

The atmospheric and land surface systems are dynamically coupled through the physical processes of energy and water supply, transformation (latent and sensible heat), and transport (streamflow) at a land–atmosphere interface (Dyck, 1983). In order to develop physically based water balance models which reflect the dynamic interchange of energy at the atmospheric-soil–vegetation interface, both the underlying physical determinism, and the uncertainty in the elements of the water balance dealing with their probability distributions must be included.

The present study aims to develop an understanding of the water balance of a first order grassland catchment, in order to develop or improve existing hydrologic models and to identify and quantify the principal factors (meteorological, plant and soil) controlling the processes of water loss in montane grasslands. The observed data were used to test the performance of the ACRU hydrological model (Schulze, 1997).

2. Experimental site

The work was carried out at the Cathedral Peak Forestry Research Station which lies in the northern part of the Natal Drakensberg Park (29° 00′S, 29° 15′E). A natural grassland catchment receiving a biennial spring burning treatment (catchment VI) was selected for the study. Catchment VI is $0.677\ \mathrm{km}^2$ in extent and is moderately dissected by streams (stream density $3.25\ \mathrm{km\ km}^{-2}$). The origins of the streams are obscured beneath soil and boulders, and are believed to rise above non-amygdaloidal layers of basalt. A topographic map of catchment VI is shown in Fig. 1. Elevations range from 1860 m a.s.l. at the basin outlet to 2070 m a.s.l. at the highest point. The terrain has an average slope of 19%.

The soils of the catchment are classified as Lateritic Red and Yellow earths, grading into heavy black soils (Katspruit and Champagne) in the saturated zones and along the stream banks (Granger, 1976). They are of residual and colluvial origin and derived from basalt. Characteristically these soils are acidic, highly leached and structureless. The topsoils are of friable consistence and are well suited for rapid infiltration and storage of water. The organic content of the top soil is high (6–10%), resulting in a high water holding capacity of the soils. In contrast, the subsoils have a very high clay content and poor infiltration.

3. Theory and methods

3.1. The water balance equation

The water balance of a catchment is a deterministic relationship between the water balance components that are random variables in time and space, with usually unknown probability distributions. The independent input variable is rainfall, which is transformed in the hydrological system into the dependent output variables evaporation, streamflow and change in soil storage.

To allow mathematical prediction of the various hydrological variables some simplifications are necessary. The most practical method is the use of the deterministic approach of applying the macroscopic version of the continuity equation. The various continuous water movement processes of the hydrological cycle are lumped over finite time intervals and areas and related by the water balance equation. The volumetric water balance per unit area may be expressed as:

$$P - E_{\mathrm{a}} - \mathrm{SS} = Q, \tag{1}$$

where P is the precipitation, E_{a} the actual evaporation,

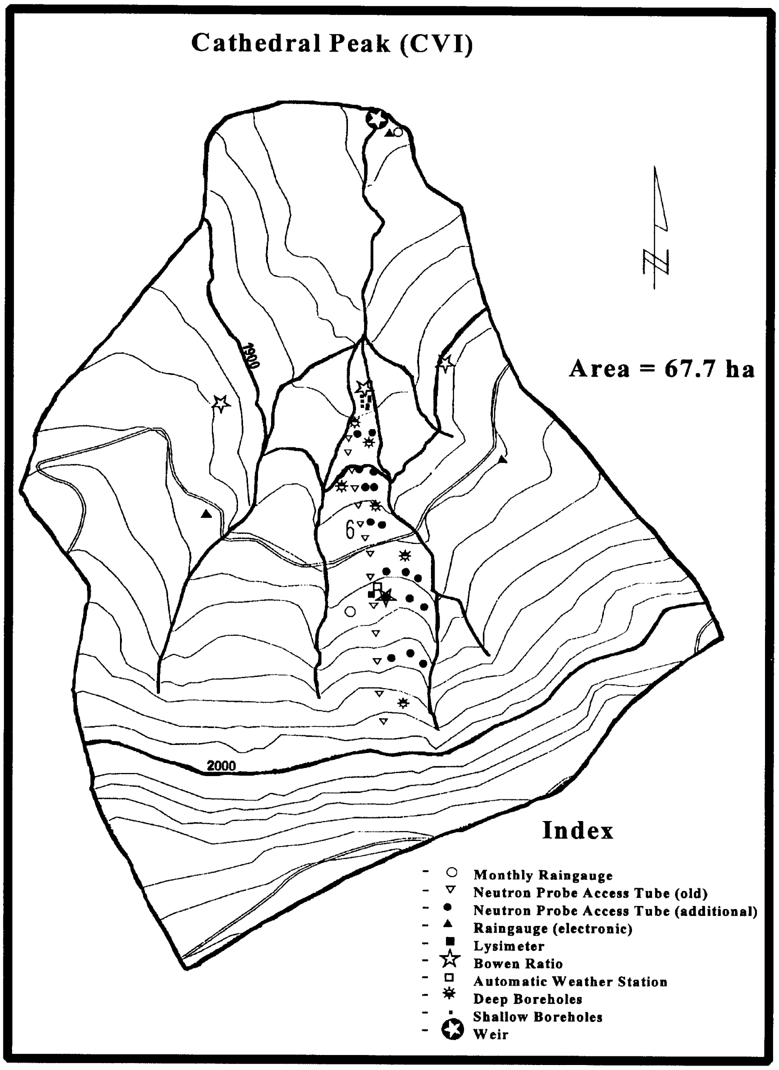

Fig. 1. Topographic map of catchment VI showing the location of the various monitoring sites.

SS the soil storage and Q is the streamflow (Eagleson, 1978). All terms except P depend upon soil moisture level and distribution which is generally not measured, the problem being overcome by assuming the system to be stationary in the mean. If the integration interval is a full year and expected values are substituted, the change of storage is negligible and the average annual water balance equation is as follows:

$$\bar{P} - \overline{E_a} = \bar{Q} \text{ mm year}^{-1}. \tag{2}$$

All components of the water balance equation were measured in catchment VI from 1990/91 to 1994/95. The validity of the data collected could be checked by balancing the equation. Between 1980 and 1989 actual evaporation (E_a) was calculated in the traditional way using the annual water balance equation. This assumes that deep losses of water are negligible and that the soil water content is identical at the beginning and at the end of the hydrological year (1 October and 30 September, respectively). With respect to the first assumption, an impervious layer of basalt underlies catchment VI, and there is no evidence of faulting, and no deep water losses occur. The second assumption is tested as part of the present study.

3.2. Rainfall

Precipitation was recorded continuously with three tipping bucket raingauges located within the catchment. Areal catchment rainfall was estimated using the Thiessens polygon method (Thiessen, 1911). The gauging network was expanded in 1992 to include six tipping bucket raingauges (five 0.2 mm resolution gauges at 1.2 m height and one 0.1 mm resolution ground surface raingauge) for the continuous monitoring of rainfall within the catchment.

3.3. Streamflow

Streamflow was monitored continuously in catchment VI for the duration of the project. A 457.2 mm, 90° V-notch weir was used with a Belfort streamflow recorder that had been recently modified with an MCS 250-01 streamflow encoder (MC Systems, CT, SA). The Department of Water Affairs and Forestry verified the calibration of the weir in 1994.

3.4. Evaporation

Evaporation was monitored using the Bowen ratio energy balance technique for the five-year study period (1990/91–1994/95). A complete description of how the technique was used in this study is given by Savage et al. (1997). The Bowen ratio (β) is obtained from the equation

$$\beta = \frac{\rho c_p (T_1 - T_2)}{\lambda \epsilon (e_1 - e_2)}, \tag{3}$$

where $\rho c_p / \lambda \epsilon$ is the psychrometric constant, $(T_1 - T_2)$ the difference in air temperature and $(e_1 - e_2)$ is the difference in vapour pressure between the sensors. The surface energy budget is given by

$$R_n - G - H - L_e = 0, \tag{4}$$

where R_n is the net radiation for the surface, G the rate of soil heat flux and H and L_e are the sensible and latent heat flux, respectively. The latent heat of evaporation is obtained from the equation

$$L_e = \frac{R_n - G}{1 + \beta}. \tag{5}$$

In this study the Bowen ratio latent heat flux density (L_e) values were converted to millimetres of water and totalled for each day.

In addition to the measurement of vapour pressure and temperature gradients, the calculation of evaporation required the measurement of net radiation (Fritchen type model Q5 (REBS, Seattle, WA, USA) and soil heat flux (Middleton Instuments Model CN3, Australia).

3.5. Soil water storage using the neutron probe

A 300 m transect for soil water assessment was laid out along a topographic gradient from the upslope region into the margin of the saturated zone (Fig. 1). A Troxler neutron moisture meter (Troxler Electronic Laboratories, NC, USA) was used for the determination of soil water content. Fourteen aluminium access tubes were placed along the transect, approximately 20 m apart and to a depth of 2 m. Weekly count ratios were routinely taken from these sites at 0.25 m depth intervals. The count rate was translated into the soil moisture content (by volume) using field derived calibration curves. Mass water content was converted to volumetric water content from previously determined values of the soil bulk density. From these data the total profile water content was calculated from each of the 14 access tubes. For brevity only data from a single representative mid-slope site are presented here.

3.6. The ACRU hydrological model

The ACRU model was developed as a simple decision making tool for agrohydrological problems (Schulze, 1995) and has been used extensively in South Africa for water resource assessments (Schulze et al., 1990), landuse management (Kienze and Schulze, 1992; Tarboton and Schulze, 1993) and irrigation supply (Dent, 1988). ACRU is a physical conceptual model that conceives a one-dimensional system in which important processes are included in discrete time units. The model represents the ability of the soil to store and transmit water, while vegetation water use is simulated using hydrological variables. The generation of stormflow is based on the assumption that after initial abstractions the runoff is a function of the magnitude of the rainfall and the soil water deficit from a critical depth of soil. The soil water

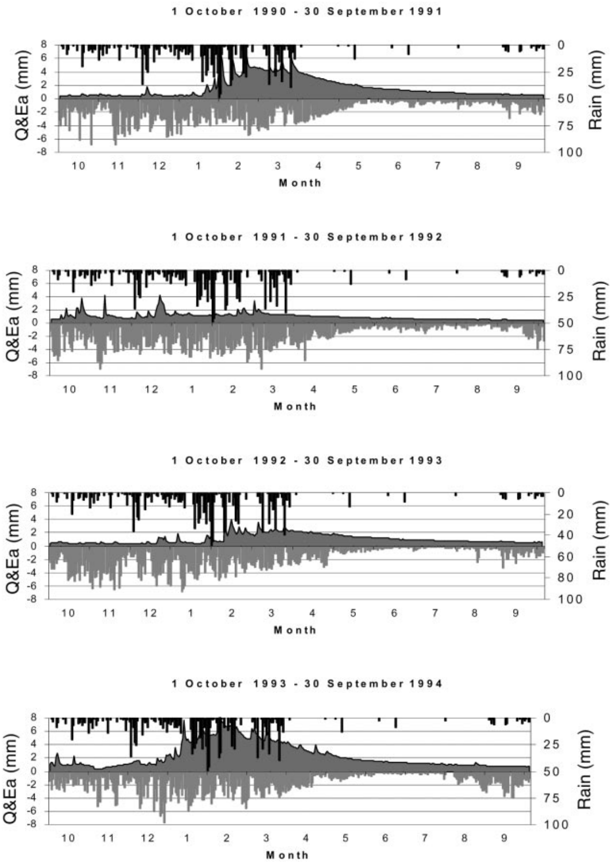

Fig. 2. Daily streamflow (primary *y*-axis: positive numbers), rainfall (secondary *y*-axis) and evaporation (primary *y*-axis: negative numbers) for the period September 1990–September 1994.

deficit antecedent to a rainfall event is simulated by ACRU's multi-layer water budgeting routines on a daily basis. The critical response depth depends on the dominant runoff-producing mechanism. Not all stormflow generated by a rainfall event is the same day response at the catchment outlet; stormflow is therefore split into quickflow and delayed stormflow, with the 'lag' dependent on soil properties, catchment size, slope and drainage density. A composite daily input data file was created from the five years of hourly meteorological and ancillary data collected in catchment VI. Variables included: observed rainfall, maximum temperature, minimum temperature, A-pan, leaf area index, incoming radiation flux density (MJ m^{-2} day^{-1}), relative humidity (%) and wind run (km day^{-1}).

4. Results and discussion

4.1. Rainfall

The variable annual rainfall during the study period provided an ideal opportunity for contrasting the water balance of wet, dry and average hydrological years. Daily precipitation in the five-year study period showed the characteristic pattern found in the summer rainfall regions, with the summer months being wet and the winters very dry (Fig. 2). There was a high frequency of dry days during the period April–September. During 1992 no rainfall was recorded in the months of May, June and July, whilst <3.9 mm fell during this period in 1993 and 1994. During the dry period, the grassland is dormant and evaporation is minimal. However, dry periods during the growing season can be expected to have a significant influence on plant water stress. For example, November 1994 was an exceptionally dry month, which received only 33.6 mm rainfall when compared to the long-term average of 190 mm.

4.2. Streamflow

The effect of the contrasting rain years on the stream discharge is clearly shown in Fig. 2. Total streamflow for the study period corresponded closely to annual rainfall. The effects of the good rains in 1993/94 were reflected in the high annual streamflow (863 mm). In contrast, the low discharge in 1994/95 (342 mm) reflected the effect of the unusually dry year.In every year of the study there was a steady decrease in streamflow from April to the end of September, which corresponded with the dry winter period. The advent of the spring rains in October had little impact on streamflow in the catchment. The increase in streamflow was only apparent in January/February when the soil moisture storage was recharged.The annual runoff during the driest year (1994/95) was 342 mm which represented 34% of the annual precipitation. The effect of this dry year was a continuous low discharge, which is depicted by the 'flat' graph with few peaks (Fig. 2). Streamflow seldom exceeded 2.0 mm day^{-1} throughout the year. In spite of some rain from October onwards (Fig. 2), the catchment was unable to recharge, and the runoff remained low.

The annual runoff during the wet year (863 mm) was 59% of the annual precipitation. Seasonal discharge from the stream reached a peak in February and gradually receded by the end of September.

4.3. Evaporation

Evaporation from the study site (moderate slope, north aspect) was measured continuously from September 1990 to June 1995. Evaporation ranged between 3 and 7 mm per day during the wet season and dropped to <1 mm per day during the dry winter months (Fig. 2). The annual evaporation totals for the four year study period (1990/91–1993/94) were 681, 752, 698 and 651 mm, respectively (Fig. 2). Although the rainfall varied by 377 mm, there was only 100 mm difference in evaporation (E_a). The low variation in evaporation measured over the four-year study period suggests that at this site the total annual evaporation is not limited by soil moisture, even in low rainfall years. The higher evaporation in the drought years is explained by the increase in the net solar radiation in these years brought about by the higher incidence of clear sunny skies.

Data collected from the Bowen ratio apparatus were used to model evaporation for the grassland vegetation in catchment VI using site-specific Penman–Monteith and equilibrium evaporation equations (Savage et al., 1997). Comparisons were made for selected periods during summer, autumn and winter. A potential problem in the use of these formulations is that neither net radiation nor soil heat flux is measured routinely at weather stations in South Africa. However, solar radiation is monitored at most sites (Reid, 1981). Estimates of isothermal net radiation were made from the sum of the net solar radiation and the net isothermal long-wave radiation as described by Campbell (undated):

$$R_{ni} = a_s R_t + L_{ni},$$

where a_s is the albedo (absorptivity of the vegetation for solar radiation), R_s the measured solar irradiance and L_{ni} is the atmospheric radiant emittance minus vegetation emittance at air temperature.

The relationship between the measured and modelled net radiation was very good (Fig. 3). A critical factor in the quantification of the net radiation was selection of the correct albedo. The 'best' albedo values were 0.25, 0.30 and 0.45 for summer, autumn and winter, respectively. These data show that the net radiation can be modelled very accurately from standard weather station data if the albedo can be adjusted to suit the prevailing conditions.

Values of stomatal resistance calculated from the inverse solution of the Penman–Monteith equation (using evaporation calculated from the Bowen ratio data) showed that average summer values were 20–100 s m^{-1}. This shows that the grasses were transpiring freely during the day and soil water was not limiting. These values are close to the standard for reference crops (70 s m^{-1}, Smith, 1991). In contrast, winter stomatal resistances were 100–500 s m^{-1}, reflecting the dormant state of the grasses. Stomatal resistances of 50 (summer), 70 (autumn) and 200 s m^{-1} (winter) were used for calculating the site-specific Penman–Monteith evaporation (Fig. 4).

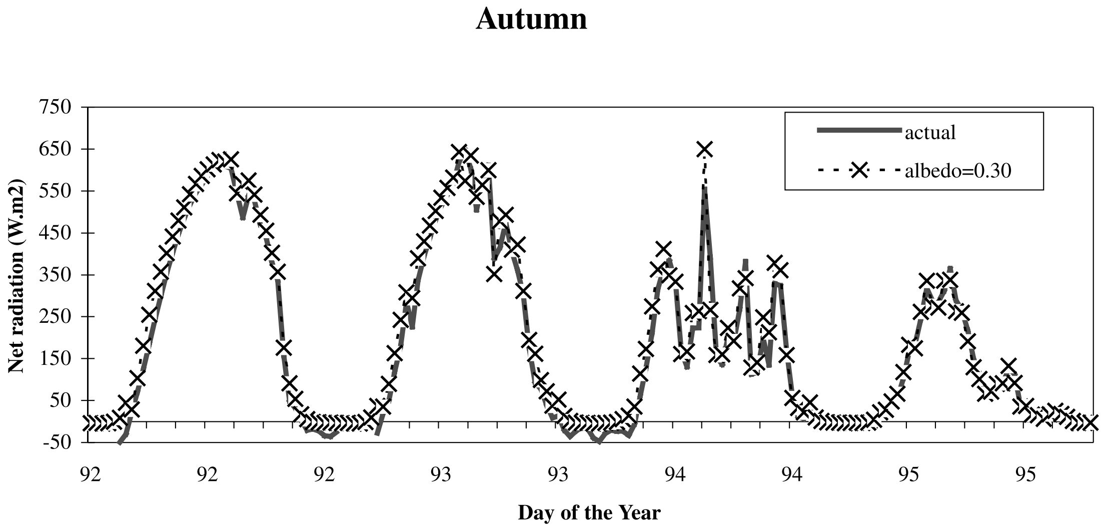

Fig. 3. Comparisons of modelled and measured net radiation in catchment VI over four days in autumn.

The diurnal trends in the Penman–Monteith data were smooth in comparison to the Bowen ratio technique. All three sets of data exhibited the same temporal variations, which closely followed daily trends of net radiation. The diurnal trends of the Bowen ratio and Penman–Monteith equation showed closest agreement (Fig. 4). Although the equilibrium evaporation followed a similar pattern, they were always higher, particularly in autumn and winter (Table 1). For example, the total evaporation in the winter month of June for the Bowen ratio (9.9 mm) showed little agreement to that of the equilibrium evaporation (47.6 mm, Table 1). This over-estimation is attributed to the fact that the equilibrium assumes a humid airflow over the canopy. In winter when the canopy is dormant, this condition is not met, resulting in poor agreement with actual evaporation. In the six-month period from January to June 1993 the Penman–Monteith under-estimated evaporation by 13% while the equilibrium evaporation rate was 55% higher than the Bowen ratio estimates.

4.3. Soil water storage

Soil moisture storage in the upper 2000 mm of the profile for the study years (Fig. 5) indicated that in summer, rainfall recharged the soil profile, while during the dry winter period it was depleted. Total profile water contents ranged from 850 to 960 mm for the 2000 mm sample depth. The maximum value of 960 mm of equivalent water is representative of the catchment in a fully charged state, while the 850 mm minimum is likely to be closest to the lowest value ever recorded for this catchment (due to the extreme drought conditions). Differences in soil water storage at the beginning and end of the four-year period were small (−11, −45, +7 and −14 mm), the only noticeable effect being the inability of the catchment to recharge at the end of 1991/92. At this time there was approximately 45 mm less water in the profile than in the previous year. These results indicate that differences in soil water storage in average and wet years are not likely to effect water balance

Table 1
Monthly totals of evaporation estimated using the Bowen ratio, equilibrium equation and Penman–Monteith model

Month	January	February	March	April	May	June	Total
Bowen ratio	100.6	85.8	75.7	46.4	20.2	9.9	338.6
Equilibrium equation	139.5	96.2	89.1	85.9	68.2	47.6	526.5
Penman–Monteith	093.2	59.2	50.5	44.5	28.8	16.4	292.7

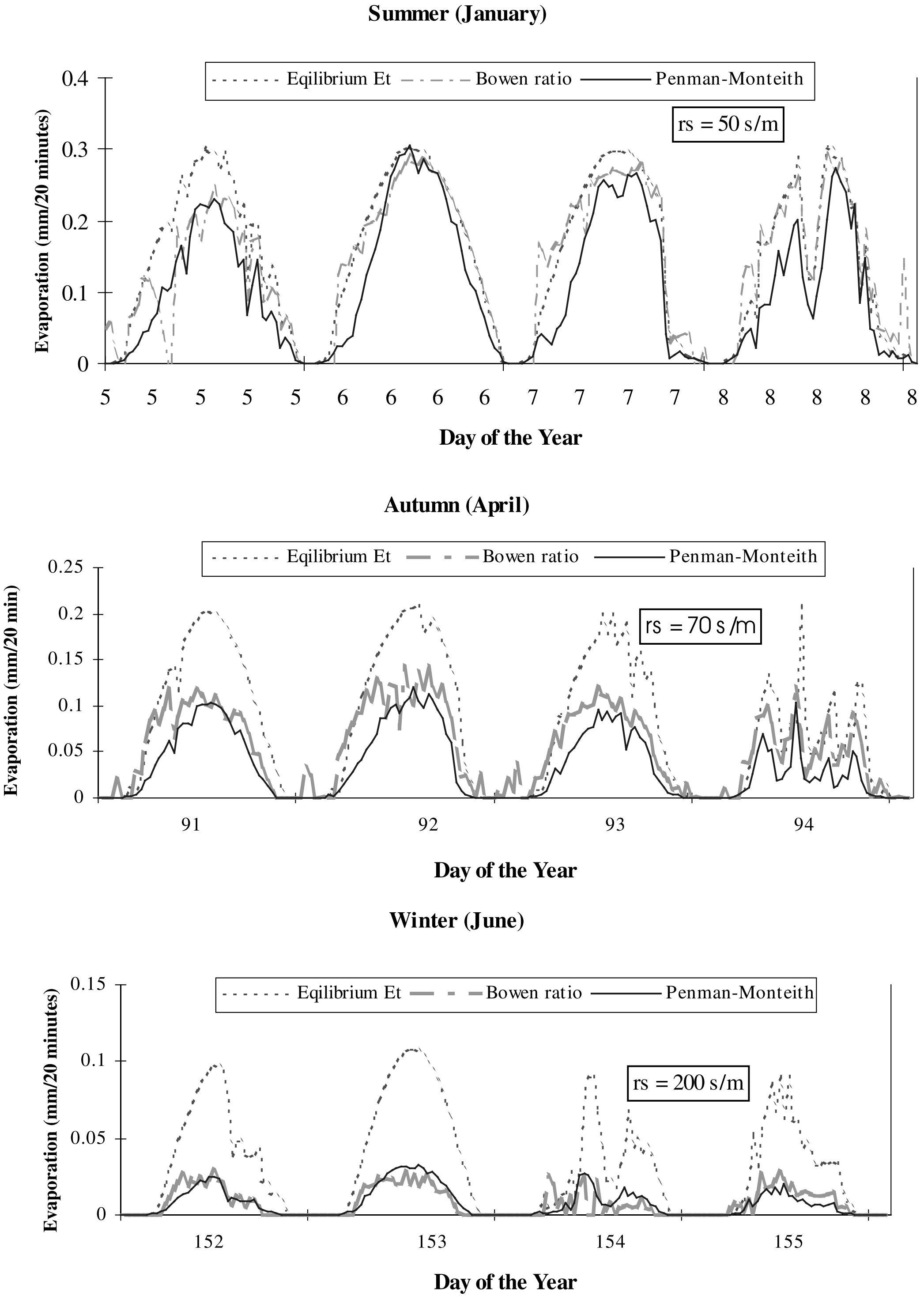

Fig. 4. Comparisons of actual and modelled evaporation for the summer, autumn and winter seasons.

Table 2
Components of the water balance equation measured in catchment VI (1990/91–1993/94). P = precipitation; E_a = actual evaporation; SS = soil storage; Q = streamflow calculated using the water balance equation; and Q_a = actual streamflow

Year	$P - E_a - SS = Q$	Q_a	Difference (mm)	% of P
1990/91	$1223 - 681 - (-11) = 553$	603	− 50	4.1
1991/92	$1092 - 752 - (-45) = 385$	366	+ 19	1.7
1992/93	$1093 - 698 - (+07) = 388$	391	− 03	0.3
1993/94	$1469 - 651 - (-14) = 822$	863	− 41	3.0

computations of E_a by more than 2%. However, the summer drought in 1991/92 would result in a 12% error.

Measured values of P, E_a and SS and the actual runoff measured at the weir (Q_a) were substituted in the water balance equation to solve for Q (Table 2). The differences between the measured runoff and that calculated using the water balance equation were small (<4.1% of the mean annual rainfall). The agreement of these results confirms the validity of the data collected in this study.

Table 3 shows the annual values for P, Q, E_T and E_a for 11 hydrological years at Cathedral Peak's catchment VI. On average 43% of annual precipitation becomes streamflow and the remaining 57% evaporates. The response ratios (Rr in Table 3) are similar to others reported in wetter and colder climates of the world. For example in New Hampshire (USA), Likens et al. (1977) calculated a response ratio of 0.61 for the Hubbard Brook catchments, while in Greece, Nakos and Vouzaras (1988), estimated a response ratio of 0.51. In the drier and more arid areas response ratios between 0.07 and 0.25 are common (Lewis, 1968; Burch et al., 1987; Pinõl et al., 1991).

Annual variability of Q and E_a for catchment VI are high. Streamflow (Q) ranged from 293 to 745 mm, while E_a ranged from 493 to 892 mm (Table 3). Fig. 6 (a) and (b) show the relation between annual values of Q, E_a and P in catchment VI. The results of the regression analysis showed that there was a significant positive correlation between streamflow and rain ($r = 0.87 : P = <0.001$) while there was a poor correlation ($r = 0.45 : P = \mathrm{NS}$) between evaporation and rain. This situation can be contrasted to drier climates, where E_a increases with P while there is no significant correlation between Q and P (Pinõl et al., 1991). This

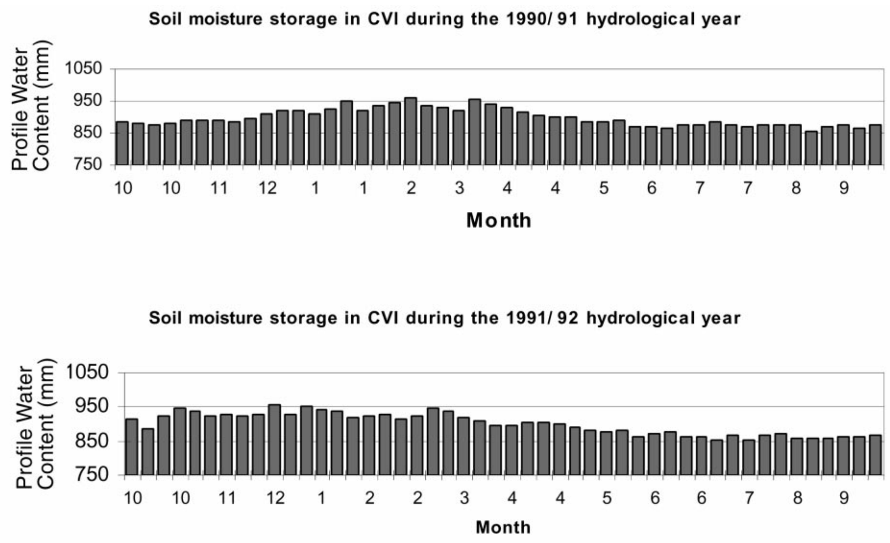

Fig. 5. Weekly total profile water content over 2000 mm in catchment VI (October 1990–September 1992).

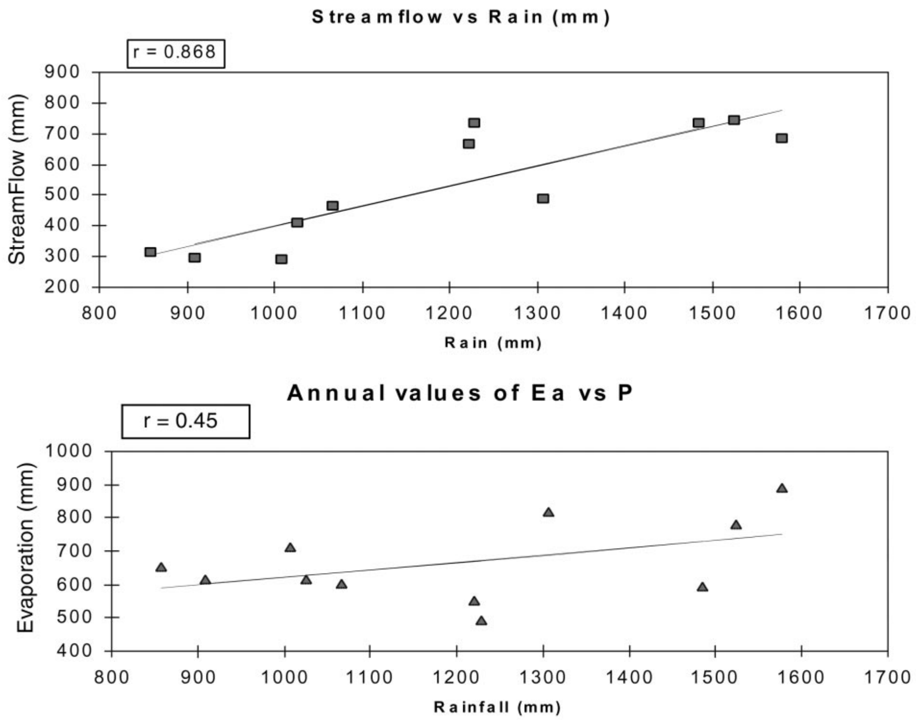

Fig. 6. Annual values of Q vs P and P vs E_a for catchment VI for an 11-year period. The lines of best fit are described by the equations $Q = 0.65P - 251$ and $E_t = 0.21P + 402.2$.

difference in behaviour can be explained by the relative magnitude of precipitation and potential evapotranspiration (E_T). When precipitation is much greater than E_T, the evaporative demand can be totally satisfied and the remaining water becomes streamflow. In contrast, when rainfall is lower than E_T, the water supply in the wetter years is still not enough to satisfy all the transpiration and potential evaporation. As a result, streamflow is more dependent on the rainfall distribution in time than on the annual volume. The E_T

Table 3
Annual water balance (mm) for catchment VI for 11 years. Data is for each hydrological year (October–September)

Year	Rain (P)	Actual evaporation (E_a)	Run-off (Q)	Potential (E_T)	Response ratio (Rr)
80	1221	552	668	1503	0.55
81	1067	601	465	1612	0.44
82	909	614	294	1521	0.32
83	1228	493	735	1734	0.60
84	1025	614	411	1578	0.40
85	1306	818	487	–	0.37
86	1578	892	686	1437	0.43
87	1524	779	745	1550	0.49
89	1008	714	293	–	0.29
91	1484	594	736	–	0.49
92	858	655	314	–	0.37
Mean	1200	666	530	1562	0.43

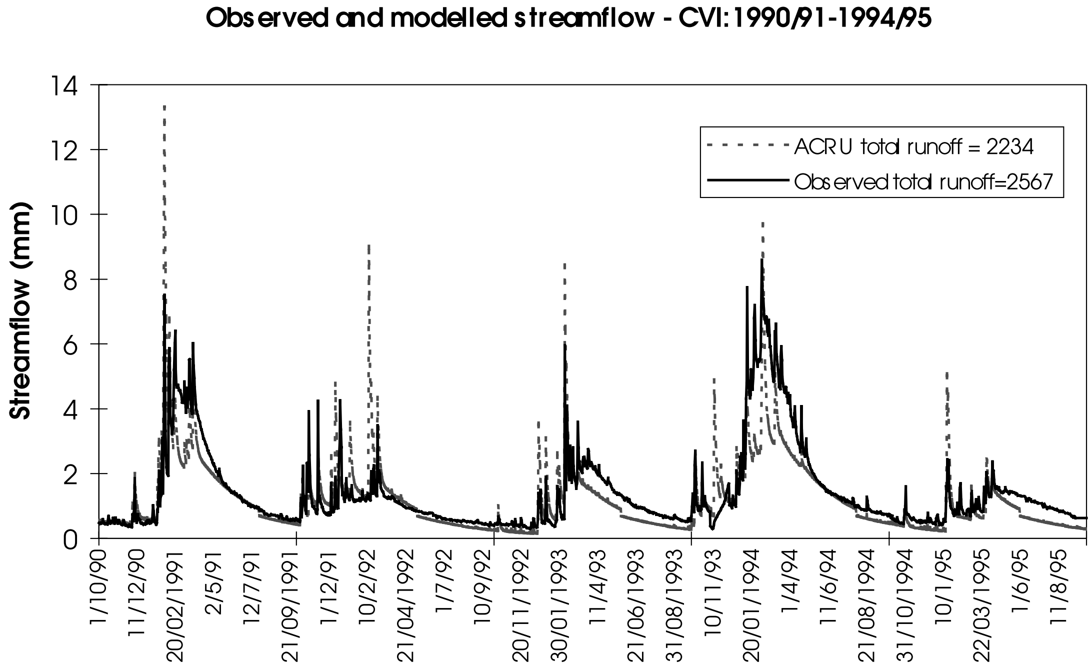

Fig. 7. Observed and ACRU modelled streamflow for catchment VI at Cathedral Peak for the period January 1990–September 1995.

estimated for Cathedral Peak using A-pan data, shows that rainfall seldom exceeds E_T, although the magnitude of this difference is not large (Table 3). Thus catchment VI is intermediate between the two climatic extremes, but tending towards the situation found in the wetter and colder climates. The slope for linear regression of streamflow on precipitation for CVI was 0.65, indicating that runoff is largely dependent on precipitation. The slope for evaporation versus rainfall was only 0.21, demonstrating that evaporation is largely independent of the annual rainfall.

4.4. Water balance modelling

The time series graph of actual and modelled streamflow showed good agreement (coefficient of agreement = 0.87; correlation coefficient = 0.80) (Fig. 7). In general, ACRU under-estimated streamflow by 15%. This resulted in 358 mm under-estimation of runoff over the five-year simulation. The model performed better in the average rainfall and dry years of 1991/92, 1992/93 and 1994/95. The biggest deviation was in the very wet year of 1993/94 when ACRU was clearly under-estimating the catchment runoff. A preliminary analysis of the unsaturated flow from the B horizon to the groundwater zone and saturated drainage showed that ACRU was unable to account for the predominance of subsurface flow found in catchment VI (Everson et al., 1998), which may explain this under-estimation.

The accurate estimation of daily E_r (reference evaporation) is a critical factor in the successful application of ACRU (Schulze, 1995). In this study monthly means of A-pan data were selected to calculate the reference evaporation (disaggregated to provide daily values) (Fig. 8). Daily A-pan potential evaporation on rainless days ranged between 5.5–6.5 and 3–4 mm in summer and winter, respectively. Estimates of ACRU total evaporation (actual evaporation − AET) over a representative period (1/7/91–30/5/93) closely followed the values using the more highly technical Bowen ratio technique. For the period 1/10/90–30/9/94 the total AET and E_t was 2664 and 2796 mm, respectively. This is a good agreement considering the generally accepted poor performance of the A-pan and crop factor approach when compared to the highly technical Bowen ratio technique.

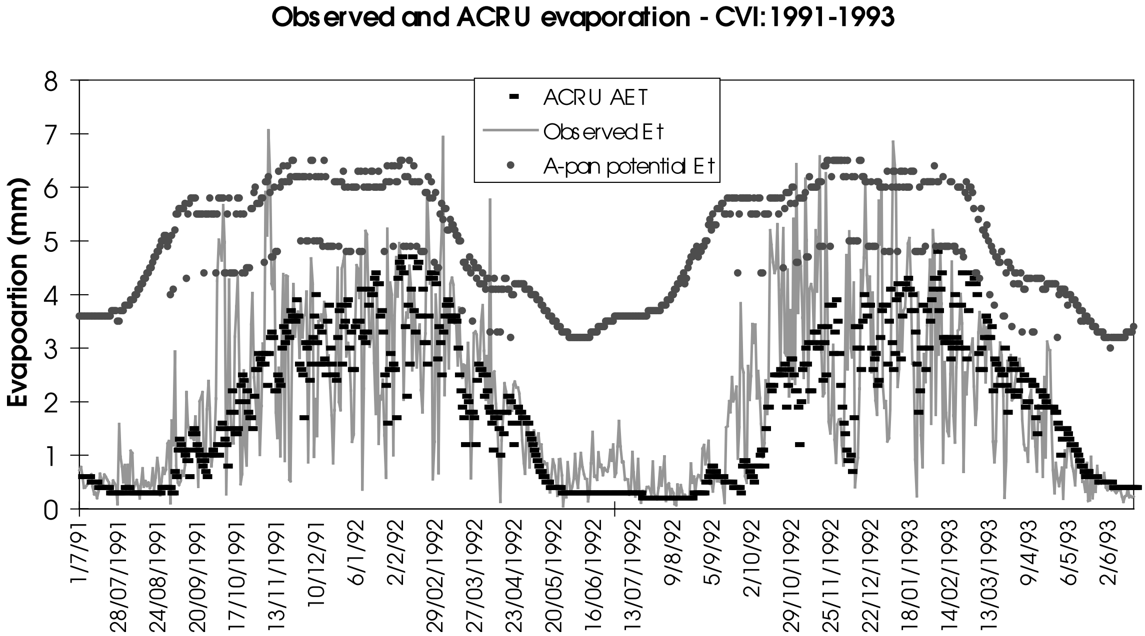

Fig. 8. Daily values of A-pan potential evaporation, ACRU modelled actual evaporation (AET) and Bowen ratio evaporation for catchment VI between July 1991 and June 1993.

The influence of the high altitudes (severe frost) and summer rainfall on shortening the growing season of these highland grasses is clearly reflected in the E_a data, where values in winter were <0.5 mm day^{-1} and there was little or no actual evaporation until late spring (Fig. 8). The crop factor approach used in the ACRU model to calculate AET (Doorenbos and Pruitt, 1977) was clearly sensitive enough to reflect this seasonality. This is surprising as crop factors are a complex mixture of surface and aerodynamic resistances of the crop and also the climate within which they were derived (Shuttleworth, 1979). The method generally works best for irrigated crops, especially if locally derived crop factors are used. The application of the method is more hazardous when the vegetation experiences stress. The lack of stress found during the growth phase in this study and the fact that the crop factors were locally derived would explain the good agreement found between the complex micrometeorological technique and the simple crop factor approach. During the winter, potential evaporation is much greater than AET or E_a (Fig. 8), demonstrating the importance of taking care when comparing evergreen forest water losses with actual grass E_a rather than assuming potential evaporation from grass, which will overestimate E_a.

5. Conclusions

Complete information on all the terms of the water balance equation is rarely available to catchment hydrologists (Wallace and Oliver, 1990). The data collected in this study therefore represent one of the few long-term data sets where all the components of the water balance equation have been measured at a daily resolution. The data show that the partition of the main hydrological fluxes into streamflow and evaporation is dependent on the wetness of the hydrological year. In average to wet years the hydrological flux is equally split between evaporation and runoff, while in drier years evaporation becomes the dominating component of the water balance. Streamflow was largely dependent on rainfall, and good predictions of streamflow from precipitation and potential evaporation were possible. Evaporation was not closely linked to rainfall, and as a result poor relationships were found.

In this study two principal techniques for estimating Et were used, i.e., the mass water balance and the measurement of the rate of water vapour into the air. The water balance method is often subject to the problem that the accuracy of the determination of evaporation is dependent on the cumulative errors in the three unknown variables in the equation (Eq. (1)). Furthermore, since evaporation is usually only a fraction of rainfall, its estimation is often only a small difference between several large terms. In our case, evaporation was accurately estimated using the mass balance approach, since it represented a large proportion of the annual rainfall, differences in storage at the beginning and end of the hydrological year were small, and losses by deep drainage were negligible. The more data intensive Bowen ratio energy balance technique provided little advantage at the scale of the annual time step, but was essential for understanding the process of evaporation at short time intervals (hours, days, months). The development of site-specific parameters for use in the Penman–Monteith formulation, using standard weather station data, demonstrated the high accuracy with which evaporation can be estimated using this technique. This is clearly a much cheaper alternative than micrometeorological techniques, which also have fetch limitations.

The ACRU model generally predicted streamflow well except in wet years. The model was successfully validated against the catchment VI data set. The empirical ‘crop factor’ approach, used to estimate the actual evaporation in ACRU, showed good agreement with the Bowen ratio derived estimates. One of the limitations of ACRU is the inability to account for subsurface soil water flow in a catchment. The data sets collected during the course of this study provide a unique data set that will be used for future model testing and refinement.

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