

It has been drawn to the attention of the Journal and the authors that the paper of Larsen and Nicholls, published in the March 2012 issue, cited findings from another paper (Wood et al., 2008, *Functional Plant Biology*, 35, 483–492) which was subsequently retracted. An addendum to the Larsen and Nicholls paper to address this matter will be published in the June 2012 issue.

The nature of the recent rainfall decrease in the vicinity of Melbourne, southeastern Australia, and its impact on soil water balance and groundwater recharge

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Rainfall at three stations in the vicinity of Melbourne, southeastern Australia, was analysed to determine the nature of the approximately 20 per cent reduction in annual rainfall observed from 1997–2010 (the ‘Big Dry’). No change has occurred in the amount of rainfall per rain day, rather a decrease in the number of rain days has occurred, especially those with higher rainfalls. These are the rain days which contribute most to significant groundwater recharge and streamflow. A soil water balance model was used to investigate the effect of this change in rainfall on groundwater recharge. The increase in time between rain days, associated with the reduced number of rain days, allows for greater drying of the soil before the next rain day, thereby reducing the water available for groundwater recharge. This leads to a disproportionately greater reduction in groundwater recharge compared to the actual reduction in rainfall, consistent with observations.

Introduction

Census data from the Australian Bureau of Statistics show that the population of the city of Melbourne (38°S, 145°E), Victoria, Australia, has grown from three million in the late 1980s to some four million at present, with projections of 6.5 to 7.5 million by 2051 (Australian Bureau of Statistics). The requirements for potable water have increased commensurately. Melbourne’s water supplies are sourced from rain falling into a number of mostly forested catchments surrounding the city, with a total area of about 1550 km² (Jayasuriya et al. 1993).

Large-scale drivers of the climatic system, such as the El Niño Southern Oscillation, affect rainfall at interannual timescales, with the Southern Annular Mode (SAM) and the intensity of the sub-tropical ridge being more important at decadal and longer timescales (Nicholls 2010, Larsen and Nicholls 2009).

During the period 1997 to 2010, southeastern Australia experienced persistently drier than normal conditions (colloquially known as the ‘Big Dry’), with a reduction in

annual rainfall of some 20 per cent compared to the period 1955–1996. However, stream flow within the catchments supplying the city of Melbourne decreased by some 30 per cent (Timbal and Jones 2008). A similar pattern is evident in the Murray–Darling Basin, where mean annual rainfall decreased by 16 per cent, while stream flows decreased by 39 per cent. Potter et al. (2010) and Cai and Cowan (2008) reported that only about 33–51 per cent of the decrease in runoff in the Murray–Darling Basin was attributable to the decrease in rainfall. A similar pattern, of a larger percentage decrease in streamflow compared to rainfall, has been observed over the whole State of Victoria (Kiem and Verdon-Kidd 2010). The cause of this relatively larger streamflow decrease was unclear, and Kiem and Verdon-Kidd (2010) proposed that daily hydrological modelling was required in order to determine the cause(s). Here we report the results of such a modelling exercise for three rainfall stations located in different parts of the Melbourne catchment area. Rainfall data are used to drive a ‘bucket-type’ soil water balance model to investigate the sensitivity of groundwater recharge (GWR), and thus potentially streamflow, to various aspects of the climate.

Approximately half of the catchment area supplying Melbourne’s water is covered in *Eucalyptus regnans* forests, which occupy the more elevated (460–1100 m) and wetter (900–2000 mm yr⁻¹) regions and are the source of some 80

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per cent of the total catchment yield (Langford 1976, Moran and O'Shaughnessy 1984, Vertessy et al. 2001, Wood et al. 2008). At lower elevations and on the drier north and west-facing parts of the catchment, smaller mixed Eucalypt forest dominates. Mature *E. regnans* forest typically consists of an overstorey of even-aged *E. regnans* with a canopy height of 55–75 or more metres and a dense understorey consisting of other much smaller trees (6–30 m) and ferns which heavily shade the ground (Ashton 1976, Ashton 2000, Wood et al. 2008). Because the catchments are forested, we have included a parameterization for canopy interception and stemflow (that portion of the intercepted rainfall which runs down the tree trunk to the ground) within the model.

In this study we address two main issues: in what way did the nature of the daily rainfall change during the 'Big Dry'; and how has this interacted with the structure of the catchment (forests and soils) to alter GWR, and so ultimately the streamflow into Melbourne's water storages?

Data

Daily rainfall data for the period January 1955 to December 2007 were obtained from the Australian Bureau of Meteorology for the stations Warburton, Maroondah and Yan Yean (Table 1, Fig. 1). These are all situated in the lower parts of different catchments in the vicinity of Melbourne. The data are nearly continuous. On the few occasions where the recorded rainfall was the accumulated sum of the previous two days, a comparison was made with the other stations to decide to which day to assign the precipitation. If it rained on both days, the amount was split between the two days. Any days where trace moisture in the rain gauge was assigned to dew fall were set to zero rainfall. Data for January 1992 and 1994 for both Maroondah and Warburton were missing, as were July 2006 and November 2007 for Maroondah. Data from Yan Yean were substituted for these months. Yan Yean had a period of missing data from January 1979 to December 1980, and no interpolation was made for this period. The monthly mean precipitation for each station is shown in Fig. 2. Mid-late winter and spring are typically

the wettest periods of the year for these three stations.

As discussed below, daily maximum and minimum temperature are required in order to calculate evapotranspiration. Temperature measurements are not taken at these rainfall stations, so daily data from the Melbourne Regional Office site were used. This is not ideal as this station is located in an urban environment, however the observing site was already heavily urbanised by 1955, at the start of this study. The city of Melbourne has grown since 1955 and it is likely that some warming due to an increased urban heat island effect may have occurred. We tested to see what effect this warming might have, with the results

Fig. 1 Map of the locations mentioned in the text. Maroondah (Mar), Warburton (Warb), Yan Yean (YY), Melbourne Regional Office (MRO), Tottington (Tott), East Sale (ES), Wurdiboluc Reservoir (Wurd) and Lake Eildon (LE).

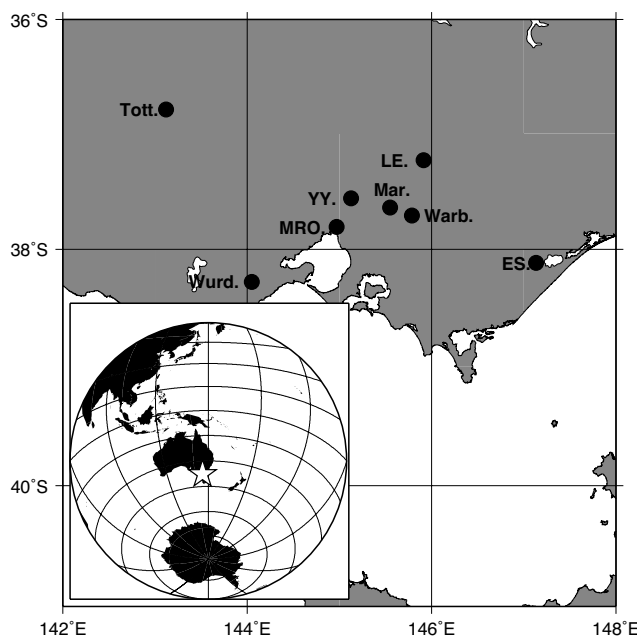


Fig. 2 Monthly mean precipitation at Warburton (filled circles), Maroondah (open circles) and Yan Yean (triangles), for the period 1955–2007, mm.

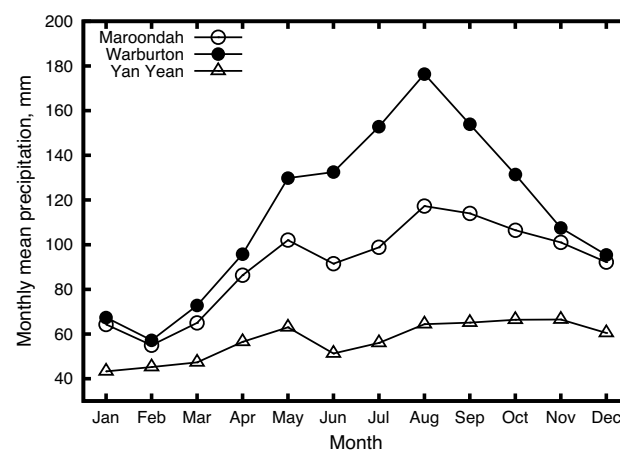


Table 1. Stations used in this study.

Station	Station number	Latitude	Longitude	Elevation (m)
Maroondah	086070	-37.64	145.55	149
Warburton	086090	-37.71	145.79	240
Yan Yean	086131	-37.56	145.13	198
Melbourne Regional Office	086071	-37.81	144.97	31
Tottington	079079	-36.79	143.12	229
East Sale	085072	-38.12	147.13	5
Wurdiboluc Reservoir	087126	-38.28	144.05	136
Lake Eildon	088023	-37.23	145.91	230

presented below. All temperature data were adjusted for elevation using the standard environmental lapse rate of 6.5 K km^{-1} .

The model introduction

A soil water balance model was used to investigate the sensitivity of GWR to various model parameters in an effort to understand how the catchment system interacts with the changing precipitation regime. The model is described in the 'Appendix'. Here we discuss general aspects of the model and explain our choices in developing and in running it.

The requirement for daily data over a long enough period to detect the difference in climatology between the 'Big Dry' of 1997–2010 and the period before (1955–1996) constrains the complexity of any model used. As there are no long-term records of wind speed, solar radiation or high-resolution (hourly or sub-hourly) rainfall data available in the catchment areas, the use of a more complex model, for example those of Hatton et al. (1993) and Donahue et al. (2010), is precluded. For this reason we have used a 'bucket-type' soil water balance model. We discuss some of the resulting limitations of this type of model below, but note that a similar approach has been used previously to model potential evapotranspiration (E_p) in *E. regnans* forests (Dunn and Connor 1993), using the method of Linacre (1977) based on temperature alone.

The model was initialised (January 1955) with the available soil water (asw) set at zero (a reasonable assumption for mid-summer in these catchments). For each day, the actual evapotranspiration (E_a) was calculated, following the method of Allen et al. (1998). If rainfall occurred, the amount of water stored in the soil was calculated as the rainfall minus the canopy interception and that day's E_a .

In the model, the soil was able to store precipitation until field capacity was reached. At that point, any extra rainfall was designated as GWR that could then contribute to streamflow. It should be noted that the high infiltration rates of the soils in these catchments are likely to reduce any surface runoff until the soils are near field capacity, at which point runoff will also occur. No distinction is made here between runoff and GWR, as we are interested in the processes leading to streamflow.

The modelled soil also loses water at the E_a rate, calculated for each day, and modified using the parameterization of Eagleman (1971), discussed below. This can occur until the permanent wilting point is reached. At that point no more water is evaporated until the next rainfall event.

Our model assumed that the soil was homogeneous and that any rainfall was distributed evenly within it. In reality the transport of water within the soil is more complex, with much of it occurring via macro pores which allow water to penetrate more rapidly to depth. The actual soils in these catchments are deep, with naturally high infiltration rates. *E. regnans* trees are also deep rooted, so that even soil water at five or more metres depth will be exploited by the forest. We

have also assumed that water only moves vertically within the profile. In reality, some lateral movement will also occur on the steeper parts of the catchments.

Because the catchments are forested, two key components of the model that must be parameterized are the effects of interception of precipitation by the forest canopy and stemflow: these will now be discussed.

Canopy interception

Rain falling over the catchments is first intercepted by the forest canopy, from which some may be evaporated and so not reach the ground. Previous studies of *E. regnans* forests have typically shown a mean canopy interception of about 20 per cent of the incoming precipitation (Langford 1976, Moran and O'Shaughnessy 1984, Jayasuriya et al. 1993). However, for any individual rainfall event, the amount of rainfall intercepted is a function of a number of physical and biological factors, operating on a range of timescales (Crockford and Richardson 1990b, Dunkerley 2009).

Canopy structure changes with the age and height of the forest (Ashton 1975, Ashton 1976, Connor et al. 1977, Watson et al. 1999). Following a fire, in a newly establishing forest, the total leaf area index (LAI) of *E. regnans* increases until the forest is about 15 years old, then steadily decreases throughout the next 200 years (Watson et al. 1999). This decrease in LAI for the (amphistomatous) *E. regnans* canopy is to some extent offset by the increase in LAI of the understorey canopy vegetation (Connor et al. 1977, Legge 1985, Wood et al. 2008).

These changes in forest structure in turn alter the percentage of rainfall intercepted by the canopy over time. Haydon et al. (1996) studied the mean interception losses under a range of differently aged stands of *E. regnans* forest, and found that they increased very rapidly from zero (for example, following a fire destroying all the trees) to a maximum of about 30 per cent when the forest was some 30 years old, thereafter slowly decreasing to about 17 per cent after about 200 years. Haydon et al. (1996) proposed that the percentage of canopy interception ($I\%$) varied with the age of the forest in years according to: $I\% = [1.2073 + 2.3453 \ln(\text{age}) - 0.3209(\ln(\text{age}))^2]^2$.

In addition to this long-term change in canopy structure with forest age, a common response of trees facing shorter-term drought-induced water stress is to shed leaves. This has been observed in several Eucalyptus species where individual trees lost between 50–97 per cent of their leaves in response to a severe drought (Pook 1985). A seasonal variation has also been observed in evergreen *Nothofagus* sp. forest (Rowe 1983), with 35 per cent (22 per cent) canopy interception in summer (winter). This seasonal difference may be due to warmer temperatures, lower relative humidity and greater leaf area in summer. Further, Dunin et al. (1988) found losses due to canopy interception and subsequent evaporation to be much greater during the day than at night.

In absolute terms, canopy interception increases with

rainfall intensity, in part through the more efficient wetting of both sides of the leaf from droplet splash (Dunkerley 2009). On a percentage basis, light rainfalls have a greater proportion of rain intercepted by the canopy than heavier rainfall events.

The physics of canopy interception are thus complex, varying with wind speed, the nature of the canopy and the absolute amount of rainfall, as well as spatially and temporally (Crockford and Richardson 2000, Dunkerley 2009). Detailed models of canopy interception typically require a range of meteorological data, sampled at high temporal resolution (sub-hourly) (Klingaman et al. 2007, Murakami 2007). As these data were not available for the sites studied here, a simpler method of modelling canopy interception must be used.

Based on daily observations in an *E. exerta* forest, Zhou et al. (2002) modelled canopy interception (C_i , in mm) as a function of daily rainfall (R , in mm) (Equation 2 in the 'Appendix'). A very similar relationship was observed by Crockford and Richardson (1990d) for a dry sclerophyll Eucalyptus forest, although the canopy interception values they obtained are some ten per cent (30 per cent) lower for intermittent (continuous) rain events in comparison to those of Zhou et al. (2002). The values from Equation 2 are closer to those previously reported (Langford 1976, Moran and O'Shaughnessy 1984, Jayasuriya et al. 1993, Haydon et al. 1996) for the *E. regnans* forests in the Melbourne catchment and so we have used Zhou et al.'s parameterization in the model.

Equation 2 is applicable for rainfall events greater than 0.4 mm. Below this threshold all precipitation is assumed to be intercepted by the canopy. This corresponds well with the results of Dunin et al. (1988) and Crockford and Richardson (1990c) who found a canopy storage capacity of about 0.35 mm for *E. regnans* and 0.39 mm for various Eucalyptus species respectively. The canopy storage capacity is the amount of precipitation retained on the leaf canopy from a rain event that does not reach the ground, and is subsequently evaporated.

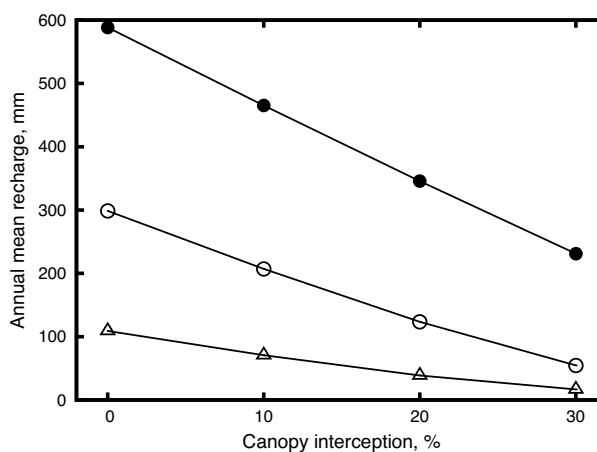
While Equation 2 describes a variable canopy interception as a function of daily rainfall, canopy interception will also vary with the age and condition of the forest, as well as seasonally. These other factors cannot be incorporated into our model directly, and so in order to investigate the sensitivity of our modelled GWR to changes in canopy interception with the age of the forest, we ran the model with fixed canopy interception factors of 0, 10, 20 and 30 per cent.

Sensitivity to canopy interception

The modelled mean annual GWR, as a function of various fixed canopy interception factors covering the potential range from bare ground, as after a fire, to very dense regrowth forest are plotted in Fig. 3.

GWR is seen to be very sensitive to variations in the canopy

Fig. 3 The effect of a fixed canopy interception factor (per cent) on annual mean GWR (solid lines, mm) 1955–2007 for Maroondah (open circles), Warburton (filled circles) and Yan Yean (triangles).



interception efficiency, and thus to the age and condition of the forest. For example, compared to a canopy interception factor of zero per cent (such as might exist following a fire), our model, run with a fixed 20 per cent interception factor as typically reported, reduced the mean annual GWR by about 60 per cent at Maroondah and Yan Yean (which have lower rainfalls) and by some 40 per cent at Warburton, where the rainfall is greater. This is consistent with the expectation that canopy interception becomes an increasingly significant factor under light or intermittent rainfalls.

Interestingly, using a fixed canopy interception value of 20 per cent, as previous studies have done, gives values for annual GWR (345, 123, 39 mm for Warburton, Maroondah and Yan Yean respectively, for the period 1955–2007) that are close to what we model using the rainfall-dependent variable canopy interception factor of Zhou et al. (2002), suggesting that 20 per cent is a reasonable approximation. However, under a varying rainfall regime, it cannot be assumed that the canopy interception will remain at this value, hence the importance of using a variable canopy interception term.

Stemflow

Of the precipitation intercepted by the canopy, a small proportion drains onto the branches and then down the trunk. This proportion, called the stemflow, needs to be added back into the model as it represents intercepted rainfall that eventually reaches the surface. Crockford and Richardson (1990b) provide observational data for stemflow (S) as a percentage of rainfall in eucalyptus forests. There is little difference in their data for continuous or discontinuous rainfall events. We have parameterized their data as follows: for rainfalls ≤ 2.5 mm there is no stemflow, for rainfalls ≥ 10.0 mm, stemflow accounts for 5.5 per cent of the rainfall. For rainfalls between 2.5 and 10.0 mm, we have used a linear relationship, $S = 0.73(\text{rainfall}) - 1.82$

Soil water holding capacity

The water holding capacity of the soils in the catchment varies with soil type. The soils are generally Kraznozems (red ferrosols) or Podzols developed on igneous rocks. These soils are typically deep (3–4 m) although on ridges they may be more shallow (0.6 m) (Ashton 1975, Langford 1976, McKenzie et al. 2004, Wood et al. 2008). They are characterised by rapid infiltration rates (13 to >90 mm hr⁻¹), so surface runoff is infrequent, only being important in the few metres either side of streams in the catchment or on very steep slopes.

The plant-available soil water holding capacity varies from small (70 mm in semiaquic Podzols) to large/very large (300–400 mm in red ferrosol) (Williams 1983, McKenzie et al. 2004). The roots of *E. regnans* show significant lateral development and in mature trees may extend to depths of five or more metres (Ashton 1975), and are thus able to fully exploit these soil water reserves.

An additional source of water storage may occur where there is deep leaf litter on the surface of the soils, although there appear to be no quantitative observations available of this term for *E. regnans* forest. *E. regnans* has an unusually high average rate of litter fall (64 g m⁻² month⁻¹) compared with other temperate trees, and this increases with stand age (Ashton 1975, Pook 1985, Martin et al. 2007).

Leaf-litter depth varies with the age and type of forest, as well as forest management practices. Under drought conditions, the higher rate of leaf fall may lead to an increased litter layer, which, as well as allowing more precipitation to penetrate to the soil (due to less canopy interception), may also help to reduce soil evaporation rates due to a mulching effect (Pook 1985). Under a mature forest canopy, soil evaporation is generally considered to be very small (Wood et al. 2008). However, using microlysimeter measurements, McJannet et al. (1996) estimated the annual evaporation rate from the soil-litter layer of old (235 years) and young (ten years) *E. regnans* forest as 110 and 150 mm yr⁻¹, or some 13 per cent and 14 per cent of the annual total evapotranspiration, respectively. The smaller evaporation rate in the older forest was attributed to lower night-time temperatures and reduced wind speed near the surface, both a function of the greater canopy height.

The available soil water-holding capacity is thus a function of the physical properties (mineralogy, depth) of the soil, which are essentially constant for a given location, and of the surface organic matter, which varies with forest age, management practices and the occurrence of fire.

In order to investigate the sensitivity of the modelled GWR to soil water capacity, values of 50–500 mm were used here, a typical range for the soils in the catchment zone.

Water availability under increasingly dry conditions

Studies of the transpiration rate of *E. regnans* under water limitation, by Connor et al. (1977) and subsequently Legge

(1985), found no significant correlation between available soil water and stomatal closure, but that the best relationship was with air temperatures >31 °C (Connor et al. 1977). Higher air temperatures are associated with high vapour-pressure deficits, which are known to trigger stomatal closure (Kallarackal and Somen 1997), and it is likely that vapour-pressure deficit, rather than air temperature, is the trigger.

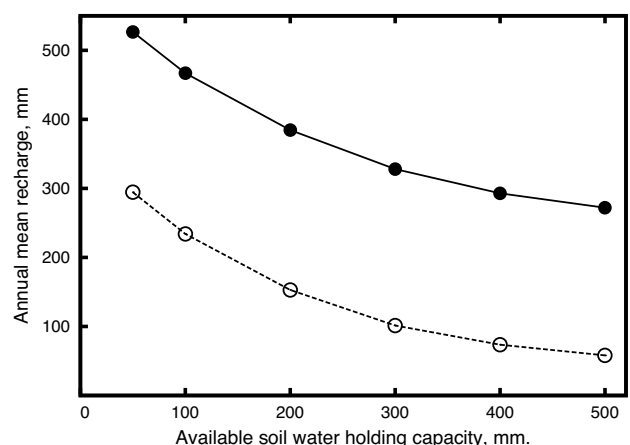
Some water is able to be stored by the tree in the sapwood of the trunk. Both Legge (1985) and Connor et al. (1977) found that even under very dry conditions *E. regnans* was able to re-equilibrate its water status overnight, so that there was no evidence of water shortage in the leaves at sunrise. It may be then that the trees can, to a limited extent, call on internal water reserves in the sapwood to smooth out the extremes of evapotranspiration, while the deep and extensive root systems of the forests allow them to exploit the available soil water. As the soil dries out, however, it becomes increasingly difficult for the roots to extract the remaining soil water and we have used the parameterization of Eagleman (1971) to account for this.

Sensitivity to soil water capacity

As noted above, the available soil water holding capacity (swc) typically varies between 70–400 mm for soils within the catchment (McKenzie et al. 2004). The model was run with swc values of 50, 100, 200, 300, 400 and 500 mm. These values incorporate any water retention by the surface litter layer. As expected, GWR is highly sensitive to swc (see Fig. 4). Increasing swc resulted in decreased GWR, as the soil is able to store more precipitation which is then subsequently able to be exploited by the forest community.

Whilst swc for any given part of the catchment is a function of the soil type, and so largely fixed, it may vary due to changes in the depth of the surface litter layer. Figure 4 indicates that any changes in the depth of such a water-retentive litter layer will have a proportionately larger effect

Fig. 4 Annual GWR at Warburton (solid) and Maroondah (dashed) as a function of soil water holding capacity, averaged over the period 1955–2007.



on the soils having a lower swc, with soils in the 100–300 mm range losing (gaining) about 0.7 mm of GWR for every 1 mm of water retention added (lost) from the soil. Under drought conditions, enhanced rates of leaf shedding may lead to a greater build-up of this water retentive litter, further reducing the GWR to the catchments.

Comparison of our model with other studies

Figure 5 shows the range of daily actual evapotranspiration (E_a) values calculated using the model. For Yan Yean, which has a lower rainfall regime, GWR was negligible for swc greater than 100 mm, and so this value was used for the soil there. For Maroondah and Warburton, swc of 300 mm were used, and these values are likely to be more typical of those found in the catchments. As the rainfall at Yan Yean is low, the soil is dry over the warmer months and into autumn. As a result, while the potential evapotranspiration (E_p) values are high, E_a is low, because there is little moisture in the soil and it becomes increasingly difficult for this to be evapotranspired.

Zhang et al. (2001) reviewed the relationship between annual evapotranspiration as a function of mean annual rainfall, for more than 250 locations globally. Wetter regions can develop more complex and dense vegetation types and so tend to have higher evapotranspiration rates. They proposed an empirical relationship to describe this, which is plotted in Fig. 6, for both a fully forested and a fully grassland catchment. Also plotted is the calculated mean annual evapotranspiration for Maroondah, Warburton and Yan Yean using the methodology described here. From Fig. 6 it would appear that our model does a reasonable job of estimating annual E_p , though with some underestimation compared to the global mean for forested ecosystems. This may reflect the nature of these Eucalyptus forests, which grow in a highly drought-prone environment.

Our results in Fig. 6 are somewhat sensitive to the soil water capacity chosen; using higher values would give higher annual evapotranspiration, at the expense of annual GWR. As noted above, a low soil water capacity (100 mm) was chosen for Yan Yean in the analysis presented here because a higher value resulted in little or no annual GWR signal, but given the current rainfall regime, and the likelihood of greater soil water capacities than this, it is unlikely that any significant GWR currently occurs at the lower elevation Yan Yean site.

Wood et al. (2008) provides monthly mean measurements of whole-forest evapotranspiration from a mature (296 yr old) *E. regnans* forest with a well-developed understorey, located near Kinglake 45 km northeast of Melbourne. Their observations range from 3.35 mm day⁻¹ in December to 0.68 mm day⁻¹ in June, with an annual total E_a of around 840 mm at a location with an annual precipitation total of 1190 mm. We used our model to simulate the same period (December 2005 – January 2007) observed by Wood et al. and the results are presented in Fig. 7 for Warburton and

Fig. 5 Monthly means of daily E_a (mm day⁻¹) calculated for the period 1955–2007 for Warburton (filled circles) and Maroondah (open circles), both with available soil water capacities of 300 mm, and Yan Yean (1955–1978, 1981–2007; triangles) with an available soil water capacity of 100 mm.

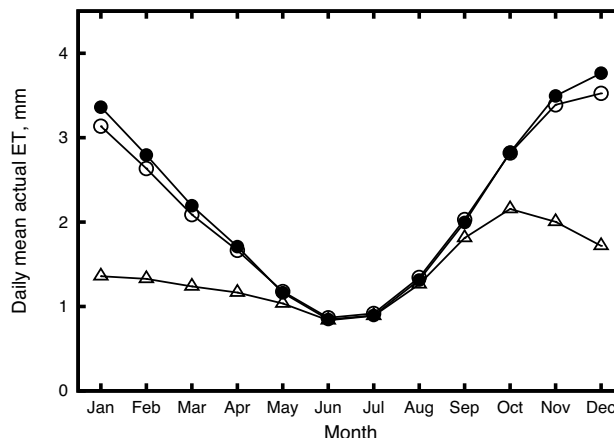
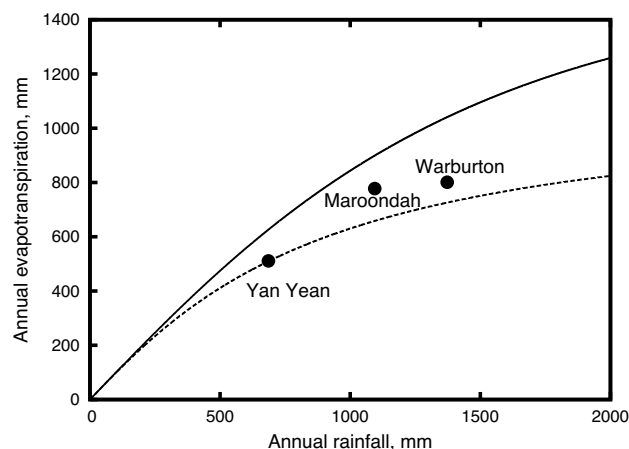


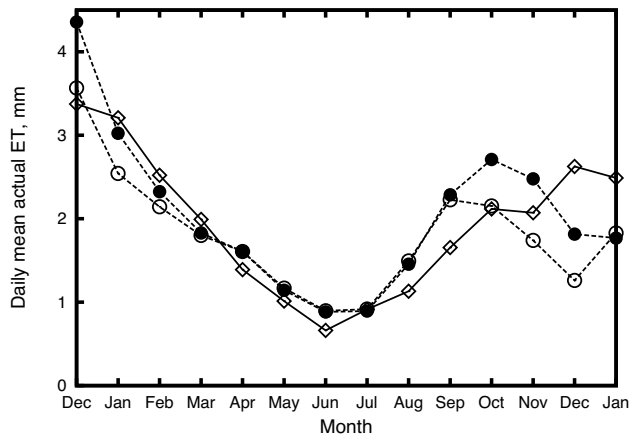
Fig. 6 Comparison of the annual actual E_a calculated for Yan Yean, Maroondah and Warburton, using available soil water capacities of 100, 300 and 300 mm respectively, compared to the empirical relationship of Zhang et al. (2001) for forest (solid) and grassland (dashed).



Maroondah (which has a similar mean annual rainfall of 1094 mm). Figure 7 indicates that our model does a good job of simulating E_a and so our calculated values for GWR are likely to be realistic.

The study period of Wood et al. (2008) spans most of two summers, one winter and the driest year (2006) in the forest's recent history (see Fig. 8). It is probable that under these conditions the trees had shed a significant amount of leaves. Our model predicted a slightly lower E_a rate in the last and driest months (see Fig. 7), as expected given that the soils were very dry and the remaining water more difficult to extract. That the trees were observed to maintain slightly higher evapotranspiration rates may indicate that they were able to exploit even deeper water reserves and potentially re-equilibrate their internal water stores overnight.

Fig. 7 Monthly means of daily actual evapotranspiration (E_a , mm day⁻¹) for the period Dec. 2005 – Jan. 2007 for Wallaby Creek forest (diamonds) after Wood et al. (2008), and the modelled E_a for Warburton (filled circles) and Maroondah (open circles), assuming a soil water holding capacity of 300 mm.



Results

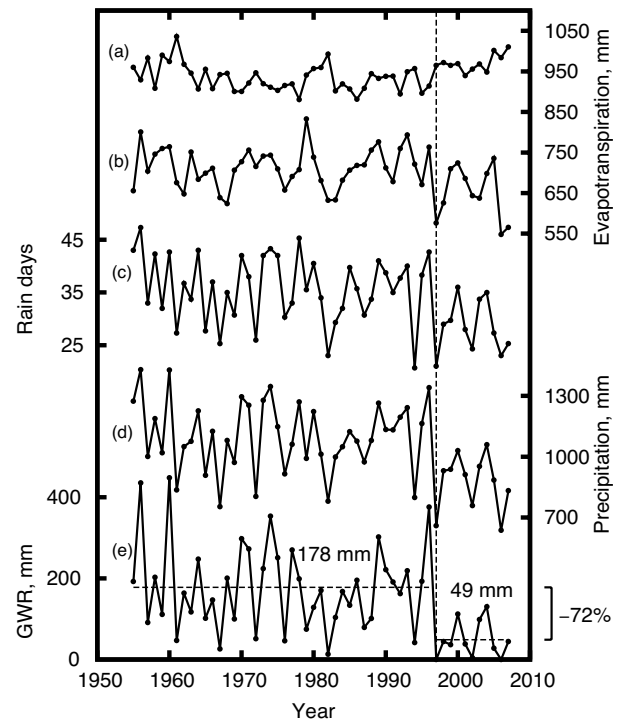
Changes in the nature of rainfall

Figure 8 summarizes how rainfall has varied at the three stations, all of which show similar patterns. A decreasing trend in rainfall is apparent since the start of the record, although it should be noted that there were some wetter-than-normal years in the 1950s. Of particular note in the context of this study is the ‘step-like’ decrease in rainfall occurring in 1997. The annual mean precipitation for the three stations in the period 1955–1996 was 1085 mm, while for the period 1997–2007 it was 872 mm, a decrease of 20 per cent. There has also been a decrease in the number of rain days with heavier rainfalls. Modelled E_p has increased post-1997, while E_a has decreased. This decrease has occurred because the soil is either too dry to evaporate water at the E_p rate, or else the soil moisture is not present to evaporate in the first place.

The daily rainfall data were sorted into specific ranges of rainfall per rain day, with these results presented in Figs 9 and 10 for the periods 1955–1996 and 1997–2007. A general decrease in rainfall is observed across all the ranges, and in most months of the year, except in the lowest (0.1–4.9 mm day⁻¹) category, where little change has occurred. The decrease is greater for the higher rainfall per rain day categories. Figure 11 plots the mean annual amount of precipitation from each rainfall class at Warburton for the periods 1955–1996, and 1997–2007. The largest decrease in rainfall is due to fewer rain days with rainfalls of 20 mm or more per day.

In order to understand further the nature of the rainfall decrease, means were taken of all the monthly rainfalls, and the number of occurrences, for each rainfall per rain day class, for the periods 1955–1996 and 1997–2007. These are shown for Warburton in Fig. 12, with Maroondah and Yan Yean having a similar pattern (not shown). The total mean rainfall for 1997–2007 is observed to decrease for each

Fig. 8 Time series of the average of Warburton, Maroondah and Yan Yean: (a) E_p , mm; (b) E_a , mm; (c) Days of rainfall with >10 mm; (d) Precipitation, mm; and (e) Modelled GWR, mm. All data are annual means. Data for 1979 and 1980 are means of Warburton and Maroondah only. The vertical dotted line indicates 1997 and the onset of the drought.



class relative to the 1955–1996 period. The mean number of occurrences also decreased for all but the 0.1–4.9 mm class of rainfall per rain day. Dividing the total mean rainfall by the number of occurrences gives an index of the mean daily rainfall rate on rain days, and little change or only a slight decrease in this index is observed between the 1955–1996 and 1997–2007 periods.

The primary cause of the decrease in rainfall, then, is a decline in the number of rain days, rather than a decrease in the intensity of the rain on a given rain day. This is most notable for the rain days with heavier rainfall. For example, in the >20 mm event⁻¹ class, the reduction from 1.1 to 0.7 rain days per month (i.e. 4.8 fewer rain days per year) each with a mean rate of about 30 mm rain day⁻¹, results in some 144 mm less precipitation per year. This is equivalent to about a ten per cent decrease compared to the 1955–1996 period, or half of the observed total reduction in precipitation. The remainder of the reduction comes from fewer rain days in the 5–19.9 mm day⁻¹ range.

Consequences for GWR

Using soil water capacities of 300 mm for Maroondah and Warburton, and 100 mm for Yan Yean, and the variable canopy interception factor of Equation 2, our model predicts a substantial reduction in GWR (Fig. 8). Modelled GWR at Warburton in the period 1997–2007 decreased by 68 per cent compared to the period 1955–1996, while at Maroondah

and Yan Yean it decreased by 85 per cent and 81 per cent respectively. These are significantly greater reductions than the observed catchment-wide streamflow decrease of some 30–40 per cent reported by Timbal and Jones (2008) and Potter et al. (2010). However Potter et al. (2010) note that in southeastern Australia the percentage decrease in catchment-wide runoff may be 2–3 times larger than the rainfall decrease (implying a 40–60 per cent reduction in streamflow for the 20 per cent reduction in rainfall we observe), and this is closer

Fig. 9 Monthly mean precipitation averaged over the three stations, Maroondah, Warburton and Yan Yean, falling in the range of (a) 0.1–4.9 mm day⁻¹; (b) 5.0–9.9 mm day⁻¹; (c) 10.0–19.9 mm day⁻¹; and (d) 20+ mm day⁻¹ for the period 1955–1996 (solid line) and 1997–2007 (dashed line).

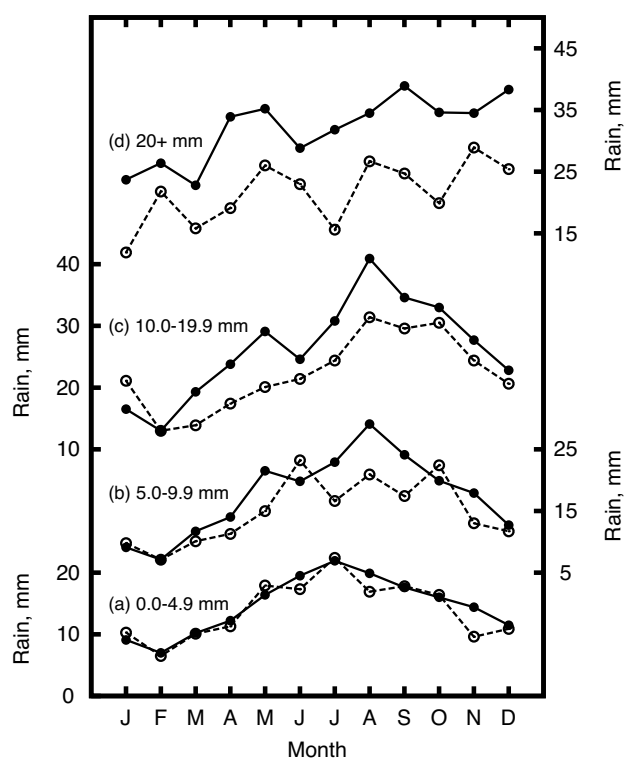
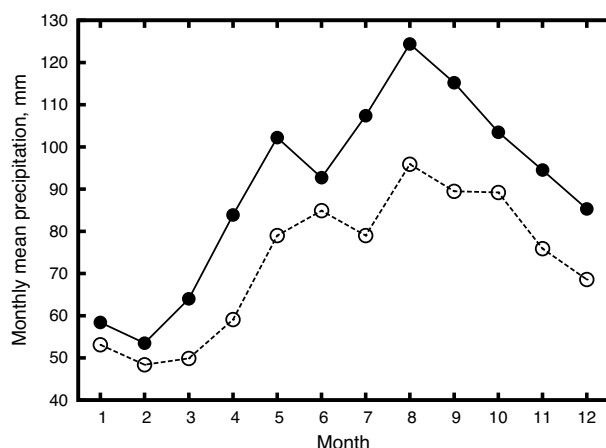


Fig. 10 As for Fig. 9 but for all the intensities combined.



to what we model. This more severe decrease in streamflow may in part be due to the fact that the elevations (Table 1), and so rainfall, of the stations we have used are lower than the higher, cooler and wetter parts of the catchment (>1000 metres). In the latter case, the catchments may be steeper, with shallower soils having lower soil water holding capacities and so are likely to be nearer to field capacity for more of the time, all of which will facilitate a greater and more reliable contribution to GWR and streamflow. By implication, this suggests also that the lower-elevation *E. regnans* forest will become drier (and so more prone to fire or drought-induced ecosystem change) before the higher-elevation forest of these catchments. A further contributing factor may be due to the steeper nature of some parts of the catchments, resulting in some movement of water downhill within the soil profile (for which, as we note above, our model does not account), and so into the streams.

Fig. 11 The annual mean rainfall at Warburton due to each category of rainfall per rain day, for the period 1955–1996 (grey) and 1997–2007 (black).

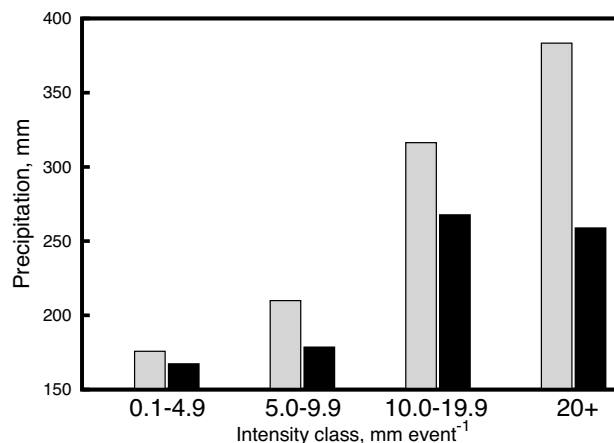
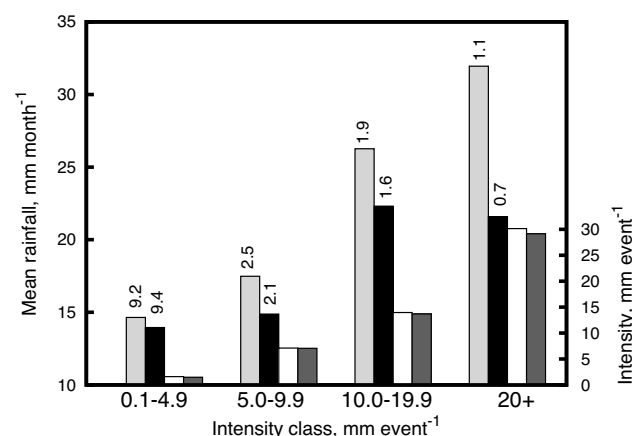


Fig. 12 Monthly precipitation, averaged for all months in the period 1955–1996 (light grey) and 1997–2007 (black). The mean number of rain days per month for each class, to one significant figure, are shown over each bar. Also plotted are the monthly means of rainfall intensity (mm per event), for each intensity class, for the period 1955–1996 (white) and 1997–2007 (dark grey). Data are for Warburton.



Effect of temperature

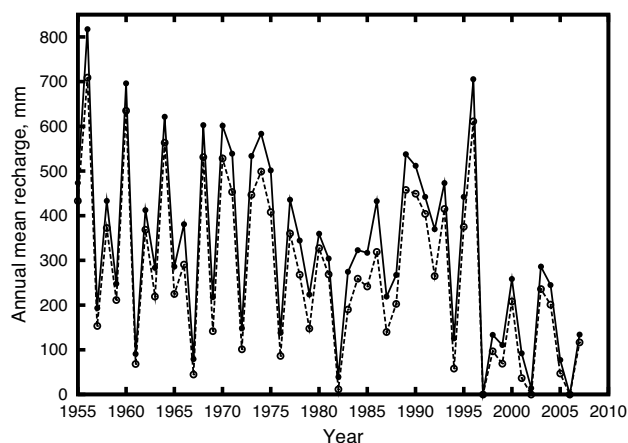
The E_p rate is a function of the amount of energy supplied to evaporate water. This includes both radiant and sensible heating. Over the period 1955–2007, the annual mean temperature in Melbourne has increased at a rate of $0.2\text{ }^{\circ}\text{C decade}^{-1}$, with increases in both maximum and minimum temperatures of 0.18 and $0.22\text{ }^{\circ}\text{C decade}^{-1}$ respectively for the annual means. In order to investigate whether the decrease in GWR, Fig. 8, has been affected by this warming trend, the model was run with the actual temperature data, and also with a synthetic data set where the temperature for each day was set to the long-term mean minimum and maximum value for that month, averaged over the period 1955–2007, thereby removing the warming trend.

Figure 13 shows the results for Warburton. It is clear that the 1997–2007 decrease in GWR does not primarily result from the warming trend in the temperature data. More complex interactions may still exist: for example, less cloud (and so potentially less rain) and increased sunshine may lead to warmer maximum temperatures and also increased E_p and reduced GWR. However the primary reason for the post-1997 decrease in streamflow is the change in the rainfall regime, as discussed earlier, not the increase in temperature.

Effect of rainfall frequency and rainfall rate

As noted earlier, the change in the nature of the rainfall post-1997 is a reduction in the actual number of days in the ‘higher rainfall per rain day’ category, rather than a change in the daily amount of these rainfalls when they do occur. It is of interest however to consider the effect of varying the amount of rainfall received per rain day, whilst keeping the total fall per month constant. A series of synthetic data sets was analysed: these used the mean minimum and maximum temperatures for each month, as described above (thereby de-trending the temperature data), and the monthly mean precipitation, divided so that it all fell on one day, two days, or so on up to six days, which were evenly distributed

Fig. 13 Annual GWR at Warburton, mm, using actual daily temperature (solid) and long-term monthly mean temperature (dashed). Soil water holding capacity is 300 mm.



throughout the month. The means were calculated using data for the period 1955–2007 and Fig. 14 shows the results for Warburton.

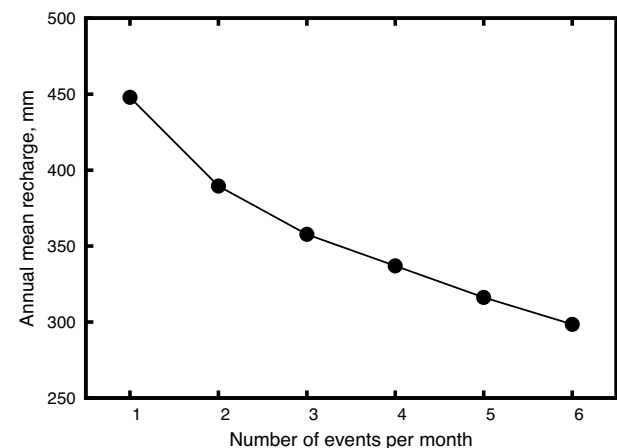
It is evident that the mean annual GWR is maximised when all of the rain falls on a single day each month, and gets progressively less as the number of rain days increases (and so rainfall per rain day decreases). This effect is driven by the reduced E_a rates of the cooler months, which allow soil moisture levels to build up. When a single large rainfall event then occurs, significant GWR results. From the daily observations it is evident that rainfall is often clustered into a series of rain days spread over consecutive days. Such clustering is therefore likely to be a more efficient producer of GWR than when the same amount of rainfall falls, separated by some days, during which time the soil can dry out again.

Discussion

Different parts of the catchment system have various response times in response to precipitation, which itself is temporally and spatially variable. The purpose of this study was to investigate the climatological changes in rainfall, and the effects of this on local soil–water balance. Extrapolation of this to a catchment scale requires caution. However, the flow of water through the soil and into the subsoil will act to smooth out some of the temporal variation, whilst the large size of the catchments relative to that of the precipitating clouds will smooth out the short-term spatial variations in rainfall intensity. So, we expect our results to be indicative of the processes occurring within the catchments.

Chiew et al. (1995), describe the amplification that occurs in the percentage reduction in runoff relative to a given rainfall decrease. In agreement with their study, the modelled results we present show how the changed

Fig. 14 Annual GWR at Warburton, mm, calculated using mean daily maximum and minimum temperatures, and rainfall as described in the text, for the period 1955–2007. Rainfall is distributed between 1 to 6 rain days per month. Soil water holding capacity is 300 mm.

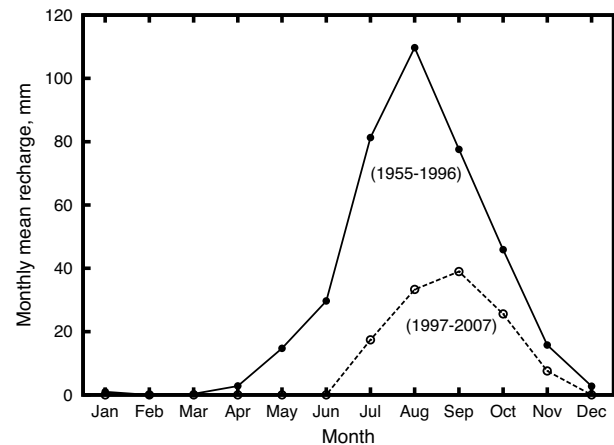


rainfall regime, interacting with the soils and forests of the Melbourne catchment, result in the proportionally greater decreases in streamflow in comparison to the decrease in rainfall, reported by Timbal and Jones (2008), Cai and Cowan (2008), Kiem and Verdon-Kidd (2010) and Potter et al. (2010). The frequency and amount of rainfall per rain day in the lowest ($0.1\text{--}4.9\text{ mm day}^{-1}$) range has remained unchanged. These rainfall amounts, however, are similar to the typical daily evapotranspiration rate, and so will not produce any significant streamflow.

The number of rain days, especially those with more intense rainfall per rain day, decreased during the 'Big Dry'. These days with heavier rainfall, which would have maintained the soil either closer to, or at, field capacity, were then less likely to, because of their reduced frequency of occurrence. As a consequence, the heaviest, though least frequent, high-rainfall rain days ($> 20\text{ mm day}^{-1}$), which were those most likely to lead to significant streamflow, more often had first to bring the soil to field capacity, before any of the precipitation could contribute to GWR. Subsequent evapotranspiration of this stored soil water, pending the next rainfall, then removed water from the system that could potentially have contributed to streamflow. It is this interaction between rainfall and soil water storage that explains why the decrease in streamflow has been proportionately greater than the decrease in rainfall. For a given vegetation cover, soil depth and rainfall regime, a threshold of rainfall per rain day and/or frequency exists for streamflow to occur. Simplistically, this can be envisaged as follows: for a soil at field capacity, with an E_a rate of 5 mm day^{-1} , a rainfall event, reaching the soil, of 20 mm every three days would yield 5 mm of GWR every three days. If the amount of rainfall on these rain days were to decrease to 15 mm every three days, or else the frequency change to a 20 mm rain day every four days, then GWR would become zero, and remain zero if conditions do not improve. A frequency and/or intensity threshold thus exists, below which rainfall will only succeed in adding to the soil moisture store, but will not be intense and/or frequent enough to bring the soil to field capacity and so produce streamflow. Such a reduced rainfall regime may still provide enough soil water (though no streamflow) to enable the forests to continue to grow, while the forests will experience longer or more frequent periods of drought. A short-term response to this outcome is increased leaf shedding, and so a reduction in the canopy interception efficiency, which allows more rainfall to reach the soil, while at the same time reducing the actual evapotranspiration from the tree. Given the extended nature of the 'Big Dry', it seems likely that the forests in the Melbourne catchments would have reduced their leaf area at the onset of the drought, with little subsequent change as the drought progressed and so little further change in catchment yield from this effect either.

Figures 9 and 10 indicate that the reduction in precipitation from 1997 onward was relatively evenly distributed over the year, however the change in GWR was not (Fig. 15). Whereas

Fig. 15 Calculated mean monthly GWR at Warburton for the period 1955–1996 (solid) and 1997–2007 (dashed), for an available soil water holding capacity of 300 mm .



in the period 1955–1996 some GWR began to occur around April, in the dry period of 1997–2007 it did not occur until July. This also follows from the interaction of the reduced frequency of higher rainfall per rain day, with the soil water storage. As the soils coming out of summer and into autumn in the 1997–2007 period have little available soil water and the amount of rainfall due to the heavier events (which would start to bring the soil toward field capacity as the weather cools and E_a decreases) is less, this means that it will take longer for these soils to yield significant streamflow. Even over winter, the reduced frequency of more intense rainfall rain days allowed the soil, at least in the lower parts of the catchment modelled here, to drop below field capacity, and July and August GWR showed the greatest absolute deficit.

Roderick et al. (2007) and Roderick and Farquhar (2004) report a general decrease in pan evaporation since the 1970s in the region. If pan evaporation has not increased, then this implies that the observed decrease in streamflow must be due to the observed decrease in rainfall, which is consistent with our results. The greater percentage decrease in streamflow relative to the decrease in rainfall is due to the interaction of the soil–forest system described above. Unfortunately there are no evaporation pans within close proximity to our sites to which we can directly compare our data. However, using data from the high-quality monthly pan evaporation data set of Jovanovic et al. (2008) for the four nearest sites (Tottington, East Sale, Wurdiboluc Reservoir and Lake Eildon; see Table 1 for more detail), we compared the mean annual pan evaporation for the period 1970–1996 and 1997–2008 (Table 2). This indicated a higher rate of pan evaporation over the latter period, consistent with our modelled E_p data in Fig. 8. If this were the case, then it would serve to exacerbate the effect we describe, with reduced rainfall and higher evaporation rates from the soil leading to a greater reduction in GWR. We note however that in their regional maps, Jovanovic et al. (2008) show little long-term trend in pan evaporation over the region. Further it is highly likely that there are some local effects at these sites. The Lake

Table 2. Annual pan evaporation 1970–1996 and 1997–2008, mm.

Station	1970–1996	1997–2008	difference, mm	per cent increase
Tottington	1403	1500	+97	7
East Sale	1334	1387	+53	4
Wurdiboluc Reservoir	1378	1390	+12	1
Lake Eildon	919	1028	+121	12

Eildon site, which shows the greatest change, is located some 150 m from the extensive spillway system of the lake, with quite a few trees in the vicinity. During the ‘Big Dry’, the spillway dried up, and the trees may have lost some leaf cover, both of which may increase pan evaporation due to reduced vapour pressure and increased wind respectively. Similarly, Wurdiboluc Reservoir is adjacent to a significant water body, as is East Sale (which is also close to the sea). Tottington is situated well away from any surface water body.

One cause from which a significant change in forest structure leading to a change in catchment yield may occur is forest fire. Fire is a key part of the ecology of these forests, as germination of new *E. regnans* seedlings requires high light levels, which only occur following the destruction of the mature forest by fire (Vertessy et al. 2001). In contrast, the mixed-Eucalyptus forests of the lower elevation and drier parts of the catchment are more resistant to fire, with regrowth occurring from the pre-existing trees. Extensive fires become more likely with the build-up of leaf litter and the dry conditions prevailing in the forests, as evidenced by the large fires in the region in February 2009. While these fires did not affect the main catchments for the Melbourne water supply, further fires may do so, and pose a significant risk to water supply. The effect of fire on streamflow yield has been the subject of a number of studies following the 13 January 1939 wildfires, which burnt through large portions of the forest in the Melbourne catchments. Langford (1976) reported a subsequent reduction of some 24 per cent in the stream flow. This was followed by Kuczera (1987) who, using additional data found that catchment stream flows decreased by some 40 per cent (300–400 mm) to a minimum level in the 15–20 years after the fire. A slow increase in stream flow then occurred, with a recovery to pre-fire stream flows expected to take 150–200 years (Kuczera 1987, Vertessy et al. 2001).

This decline and recovery of streamflow occurs because of the replacement of old-growth forest (around 65 stems per hectare) by numerous seedlings (2×10^6 stems per hectare) of regenerating *E. regnans*, which compete for resources, ultimately leading to some 1000 rapidly growing saplings per hectare about 20–25 years after the fire (Ashton 1976, Wood et al. 2008), at which time evapotranspiration rates peak and streamflows are lowest. As the forest continues to age, the density of trees per hectare gradually reduces and when the forest is 100–200 years old gaps in the canopy

form. These factors result in reduced evapotranspiration, and so increased stream flow from the catchment (Dunn and Connor 1993).

These earlier studies neglected evapotranspiration from the understorey vegetation, as it was assumed to be negligible. This assumption was revisited by Wood et al. (2008) who found that evapotranspiration from the understorey vegetation increased from near zero in a 24-year-old *E. regnans* forest, to some ten per cent of the total in an 80-year-old forest and up to 48 per cent of the total summer E_a in old (296-year) forest. As a consequence, the recovery to pre-fire conditions may occur more rapidly than envisaged by Kuczera (1987), being mostly attained by the time the forest is mature.

The immediate effect of a fire would be to reduce the canopy interception of rainfall, and the extent of the surface litter layer. Our model predicts that both of these will lead to increased GWR and an expected increase in streamflow immediately following the fire. While Kuczera (1987) did not find an increase in streamflow following the 1939 fires, it has been seen in other clear felling or thinning trials in catchments within the region (Jayasuriya et al. 1993, Vertessy et al. 2001, Wood et al. 2008). However, as shown by Wood et al. (2008), this short-term increase is followed by a reduction in streamflow of some ten per cent peaking about 25 years after the fire, and lasting for decades. The dense forest canopy that results, and so the higher canopy interception (30 per cent) of such a rejuvenating forest, combined with a tendency for less heavy rainfalls per rain day, such as occurred during the ‘Big Dry’ would all serve to significantly reduce GWR. Any significant fire in the catchments would thus exacerbate the effects of the reduced rainfall described here, with long-term consequences for Melbourne’s water supply.

Conclusions

The 20 per cent decrease in rainfall since 1997 into the Melbourne catchments has primarily occurred through a decrease in the number of rain days, particularly those of higher rainfall per rain day. There has been little change in the actual amounts of the daily rainfalls, when they occur. While temperatures have increased, our model indicates that this is not the primary reason for the decrease in GWR.

This decrease in the frequency of rain days, and so the increase in the time interval between them, allows evapotranspiration to deplete the soil-moisture levels by a larger amount. As a consequence, when the next rainfall occurs, proportionately more of the precipitation is used to bring the soil to field capacity before any recharge to the groundwater and the streamflow can then occur. These two factors appear to be the explanation for the observed 30 per cent or more decrease in streamflows from the catchments. Further, even though the frequency of those rain days with higher total rainfall has declined in all months of the year, a larger decline is observed in autumn streamflows.

This occurs in the model because the soils are well below field capacity coming out of the summer season (when streamflow is low even at the best of times). The reduced frequency of heavier rainfalls per rain day therefore means that it will take longer for these soils to reach field capacity, and for significant groundwater recharge to occur. Therefore, autumn streamflows in particular will remain at summer lows for longer. By winter the soils will have received enough water for GWR to occur, and coming into spring some GWR will still be produced, the soils still being closer to field capacity, and spring generally being a period of higher rainfall (Fig. 2).

Of the various components of the modelled catchment, some are relatively stable for a given site (such as swc) while others are more variable. Of these, the most significant variable component is the canopy interception factor. This becomes more important when the rainfall regime has fewer heavier rainfalls per rain day, and so a greater relative proportion of low rainfall per rain day (and probably more intermittent) events. Understanding and adequately modelling this will require sub-hourly data.

As precipitation increases with elevation, the lower parts of the catchment (which also represent a greater proportion of the total catchment area) will experience drier conditions and reduced GWR to streamflow sooner in response to a shift to reduced precipitation. This may have consequences for the biology of these lower elevation ecosystems and is likely to make these forests more fire prone, with any major fire posing a serious threat to Melbourne's water supply.

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Appendix

The soil water balance model

A 'bucket model' soil water balance was used. The (plant available) soil water capacity (swc), i.e. the amount of water the soil can hold within the rooting zone, is defined as the difference between field capacity and permanent wilting point. The actual (plant available) soil water (asw) therefore varies between zero (at permanent wilting point) and the swc (at field capacity).

The available soil water (asw) content was assumed to be zero at the start of each model run (January 1955), which approximates what might be expected at that time of the year.

For each day, the asw was calculated:

$$asw = asw_{-1} + (rainfall - Ci + S) - Ea \quad \dots \text{Eqn 1}$$

where asw_{-1} was the available soil water the day before, E_a is the actual evapotranspiration, defined below, S is the contribution due to stemflow, defined in the text, and C_i is the canopy interception following Zhou et al. (2002), calculated as:

$$Ci = 0.65 R^{0.55} \quad \dots \text{Eqn 2}$$

where R is the rainfall in mm. If asw were less than or equal to zero, then a 'dry soil day' was counted, and the soil remained with $asw = 0$. When rainfall exceeded evaporation to the extent that on a given day the asw became greater than field capacity (i.e. $asw > swc$) then the extra water above swc was counted as groundwater recharge (i.e. runoff and/or percolation to the sub-root zone), and asw was set equal to the swc so that:

$$runoff + percolation = asw - swc \quad \dots \text{Eqn 3}$$

The water lost as runoff and/or percolation was summed over the period from 1 January to 31 December, at which point the counters for GWR were reset to zero, but the asw was not. Thus no 'artificial rainfall' was added, and a wet/dry December could potentially affect the following January.

Calculation of evapotranspiration

Ideally potential evapotranspiration (E_p) would be determined based on the Penman–Monteith method, as recommended by the United Nations Food and Agriculture Organisation (Allen et al., 1998).

However, where not all of the data required for the full Penman–Monteith calculation are available (notably missing wet/dry bulb temperature) Allen et al. (1998) recommend that:

$$E_p = \frac{Q_a}{2.45} 0.0023(T_m + 17.8) \sqrt{(T_{max} - T_{min})} \quad \dots \text{Eqn 4}$$

be used, where Q_a is the daily solar energy flux at the top of the atmosphere (Equation 5 below) and T_{min} , T_{max} and T_m are the minimum, maximum and mean air temperatures respectively (in °C). The $(T_{max} - T_{min})$ term in Equation 4 acts as a surrogate measure of cloud cover, being smaller (larger) on overcast (clear) days. Q_a is converted from units of MJ m⁻² d⁻¹ to millimetres of water able to be evaporated per day by dividing it by the latent heat of vaporisation of water (2.45 MJ kg⁻¹)

Solar Radiation

Incoming solar radiation ultimately provides all of the energy to evaporate soil and plant water. Following Iqbal (1983) and Hartmann (1994), the solar flux Q_a (W m⁻²) incident at the top of the atmosphere (TOA) for each day was calculated as:

$$Q_a = \frac{Q}{\pi} \left(\frac{R_0}{r} \right)^2 (\omega \sin \phi \sin \delta + \cos \phi \cos \delta \sin \omega) \quad \dots \text{Eqn 5}$$

where Q is the solar constant (1367 W m^{-2}), R_0 the mean earth–sun distance ($149.59789 \times 10^9 \text{ m}$), φ is the latitude (negative in the southern hemisphere) and r is the actual distance from the sun on a given day. This distance is calculated as:

$$r = \frac{R_0}{\sqrt{e}} \quad \dots \text{Eqn 6}$$

with e being a correction factor for the earth's orbital eccentricity. It is calculated as:

$$e = 1.00011 + 0.034221\cos\Gamma + 0.00128\sin\Gamma + 0.0007198\cos 2\Gamma + 0.000077\sin 2\Gamma \quad \dots \text{Eqn 7}$$

in which Γ is the day angle (in radians):

$$\Gamma = 2\pi \frac{(d_n - 1)}{365} \quad \dots \text{Eqn 8}$$

with d_n (day number) ranging from 1 to 365 and ω being the hour angle at sunrise/sunset:

$$\omega = \arccos(-\tan\varphi \tan\delta) \quad \dots \text{Eqn 9}$$

with δ being the declination:

$$\delta = (0.006918 - 0.399912\cos\Gamma + 0.070257\sin\Gamma - 0.006758\cos 2\Gamma + 0.000907\sin 2\Gamma - 0.002697\cos 3\Gamma + 0.00148\sin 3\Gamma)(180/\pi) \quad \dots \text{Eqn 10}$$

The E_p calculated above assumes that the plant is able to extract water from the soil freely, however this is not the case as the soil becomes increasingly dry, in which case the actually observed evapotranspiration (E_a) is less than the E_p .

E_a is calculated following the method of Eagleman (1971), which uses the cubic equation:

$$E_a = 0.732 - 0.050(E_p) + \left[4.97(E_p) - 0.661(E_p)^2 \right] M_R - \left[8.57(E_p) - 1.56(E_p)^2 \right] M_R^2 + \left[4.35(E_p) - 0.880(E_p)^2 \right] M_R^3 \quad \dots \text{Eqn 11}$$

to represent the increasing difficulty with which the plant roots are able to extract soil water for a given soil moisture content (here, $M_R = \text{asw/swc}$) and E_p rate.

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Larsen and Nicholls (2012, hereafter LN2012), cited Wood et al. (2008, subsequently W2008), in several instances and used some of the data from W2008 in the validation of our calculation of evapotranspiration. W2008 was initially accessed by us in 2009 and used in the development of the model used in LN2012. Subsequent to the publication of LN2012 it was drawn to our attention that Benyon et al. (2010, hereafter B2010) identified errors in W2008, which was subsequently retracted in late 2010. In light of this we have reviewed LN2012 to see whether this retraction of W2008 affects our conclusions.

While we cited W2008 in our 'Introduction', which provided a general background to our modeling, it was in concert with other published work, and related to well-established features of the forest and hydrological system. We also cited an assessment by W2008 that earlier studies had suggested that the understorey evapotranspiration is 'negligible'. B2010 cited two prior studies that estimated understorey evapotranspiration as 26–54 per cent of total evapotranspiration.

B2010 were critical of the sap flow measurements in W2008, but we did not use these data. We used the monthly whole forest evapotranspiration data (in Figure 6A of W2008) as measured by the eddy covariance tower (re-expressed in mm d⁻¹ in our Figure 7) in part of the validation of our model. These eddy covariance data, which lead to an estimate of annual total evapotranspiration of around 700 mm (rather than the incorrect estimate of 840 mm noted in LN2012), appear plausible given the annual rainfall of around 720 mm for the year studied by W2008. The model used in LN2012 was also validated by comparison with the global empirical relationship between rainfall and evapotranspiration in Zhang et al. (2001).

Finally, we cited the results and conclusions of W2008 in a

brief, peripheral discussion of the possible consequences of a fire in the forest catchments for Melbourne's water supply. In view of the retraction of W2008, we cannot support the inclusion of the W2008 conclusions in our discussion, and thus affirm the understanding of Kuczera (1987) and Vertessy et al. (2001), including that the highest ground water yields will be obtained under old-growth forests.

We do not believe that the criticisms of W2008 by B2010, and the subsequent retraction of W2008, negate the conclusions in LN2012, but we encourage readers to refer to B2010 in this context.

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Rae Moran and Michael Manton alerted us after publication of LN2012 that W2008 had been retracted.

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