A simple framework for relating variations in runoff to variations in climatic conditions and catchment properties

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[1] We use the Budyko framework to calculate catchment-scale evapotranspiration (E) and runoff (Q) as a function of two climatic factors, precipitation (P) and evaporative demand (E_o = 0.75 times the pan evaporation rate), and a third parameter that encodes the catchment properties (n) and modifies how P is partitioned between E and Q. This simple theory accurately predicted the long-term evapotranspiration (E) and runoff (Q) for the Murray-Darling Basin (MDB) in southeast Australia. We extend the theory by developing a simple and novel analytical expression for the effects on E and Q of small perturbations in P, E_o, and n. The theory predicts that a 10% change in P, with all else constant, would result in a 26% change in Q in the MDB. Future climate scenarios (2070–2099) derived using Intergovernmental Panel on Climate Change AR4 climate model output highlight the diversity of projections for P (±30%) with a correspondingly large range in projections for Q (±80%) in the MDB. We conclude with a qualitative description about the impact of changes in catchment properties on water availability and focus on the interaction between vegetation change, increasing atmospheric [CO_2], and fire frequency. We conclude that the modern version of the Budyko framework is a useful tool for making simple and transparent estimates of changes in water availability.

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1. Introduction

[2] Assume the long-term annual average precipitation in a catchment changed by 20%. How would the runoff change? This is a deceptively simple question that has attracted the attention of hydrologists, agricultural scientists, geoscientists, water supply engineers and others for at least the last 50 years [*Nêmec and Schaake*, 1982; *Schaake and Liu*, 1989; *Dooge*, 1992]. Water for societal use (e.g., irrigation, urban or industrial uses, etc.) in many regions is harvested from runoff. Hence the answer to the question is also the starting point for evaluating the impacts. Given the importance, it is not surprising that the question has received extensive attention.

[3] A unifying approach to this question is being built using the Budyko framework [*Budyko*, 1948, 1974; *Milly*, 1993, 1994; *Dooge et al.*, 1999; *Koster and Suarez*, 1999; *Zhang et al.*, 2001; *Arora*, 2002; *Koster et al.*, 2006; *Yang et al.*, 2007; *Gerrits et al.*, 2009], where the partitioning of precipitation (P) between evapotranspiration (E) and runoff (Q) is treated as a functional balance between the supply of water from the atmosphere (precipitation, P) and the demand for water by the atmosphere (here called the evaporative demand, E_o). In parallel with the ongoing empirical work there is long-standing interest in theoretical aspects of the "Budyko curve."

[4] Two threads of theoretical interest are discussed here. One of these appears to originate from Russian scholars (who were perhaps colleagues of M. I. Budyko) [*Bagrov*, 1953; *Mezentsev*, 1955] and can trace a near-continuous lineage to the current day [*Turc*, 1954; *Pike*, 1964; *Choudhury*, 1999; *Milly and Dunne*, 2002]. This thread has used a "Budyko equation" of the generalized form [*Choudhury*, 1999]

$$E = \frac{P\ E_0}{\left(P^n + E_0^n\right)^{1/n}} \tag{1}$$

with a catchment-specific parameter, n (dimensionless), that modifies the partitioning of P between E and Q.

[5] Another thread starts with an independent mathematical derivation by the Chinese scholar Baopu Fu [*Fu*, 1981]. This work received little attention until an English translation of the main derivation became available [*Zhang et al.*, 2004]. It has since found widespread application [*Zhang et al.*, 2004; *Potter et al.*, 2005; *Yang et al.*, 2006, 2007]. This thread has used a "Budyko equation" of the generalized form

$$E = P + E_o - \left(P^\omega + E_o^\omega\right)^{1/\omega} \tag{2}$$

and also has a catchment-specific parameter, ω, that performs a similar role to that of n.

[6] Recently, these two equations have been reconciled by *Yang et al.* [2008], who showed that equation (1) can be derived using Fu's approach. Further analysis of data from numerous Chinese catchments has shown that the resulting predictions from the two "Budyko equations" are almost

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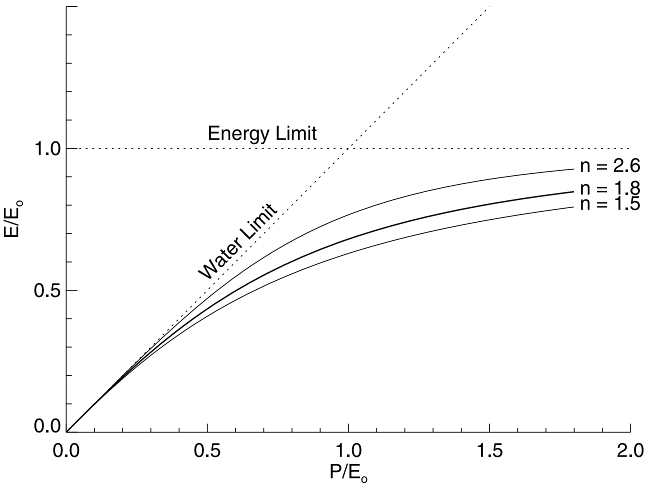

Figure 1. Relations between elements of the catchment water balance as per the Choudhury-Yang formulation of the Budyko framework (equation (1)) for the typical range in n.

identical and that the catchment parameters, ω and n are more or less linearly related as follows [*Sun*, 2007; *Yang et al.*, 2008]:

$$\omega \approx n + 0.72 \tag{3}$$

With that result we now have a way of assimilating the key results from both theoretical threads. In terms of equation (1), values of n are typically in the range 0.6 to 3.6, but most are in a smaller range from 1.5 to 2.6 [*Choudhury*, 1999; *Yang et al.*, 2007; *Yang et al.*, 2008; *Yang et al.*, 2009]. Note that higher values of n denote a higher estimate of E for a given P and E_o. This range (1.5 to 2.6) neatly brackets the original value (n = 2) assumed in the so-called Turc-Pike relation [*Turc*, 1954; *Pike*, 1964] as well as the recommended default value for n (= 1.8) [*Choudhury*, 1999].

[7] In this paper, we use equation (1) to develop a general analytical framework to estimate the change in runoff in a given catchment as a function of changes in two climatic variables (P, E_o) and in the catchment properties (n). We apply the theory using data from the Murray-Darling Basin (hereafter MDB) in southeast Australia. Following that, we show how climate change scenarios can be incorporated within the framework and finish with a qualitative discussion about how changes over time in the catchment properties impact on changes in Q in the MDB.

2. Theory: The Budyko Framework

2.1. Steady State Water Balance

[8] The formulation used here assumes steady state conditions and therefore require a time scale whereby changes in catchment storage are small relative to the magnitude of fluxes (P, E, Q) [*Donohue et al.*, 2007]. In practice this requires averages over at least 1 year and here we assume climatic time scales usually based on 30 year averages. With that assumption in mind the functional form of the "Budyko equation" used here (equation (1)) is shown in Figure 1 for the typical range in n. Note that irrespective of the value of n, equation (1) predicts that under arid conditions ($P \ll E_o$), we have $E \to P$. That asymptotic behavior captures the limit on E imposed by water supply. Alternatively, under humid conditions ($P \gg E_o$), the asymptotic energy limit, $E \to E_o$, applies. For given P and E_o, an increase in n, means that E will increase with a complementary decrease in Q.

2.2. Change in the Steady State Water Balance

[9] We use the formulation (equation (1)) to calculate the changes in E due to changes in climate (P, E_o) and catchment properties (n). To first order, the change in E is

$$dE = \frac{\partial E}{\partial P} dP + \frac{\partial E}{\partial E_o} dE_o + \frac{\partial E}{\partial n} dn \tag{4}$$

with the respective partial differentials given by

$$\frac{\partial E}{\partial P} = \frac{E}{P}\left(\frac{E_o^n}{P^n + E_o^n}\right) \tag{5a}$$

$$\frac{\partial E}{\partial E_o} = \frac{E}{E_o}\left(\frac{P^n}{P^n + E_o^n}\right) \tag{5b}$$

$$\frac{\partial E}{\partial n} = \frac{E}{n}\left(\frac{\ln\left(P^n + E_o^n\right)}{n} - \frac{\left(P^n \ln P + E_o^n \ln E_o\right)}{P^n + E_o^n}\right) \tag{5c}$$

These equations form a basis for understanding how changes in climate and catchment properties impact on E.

[10] We make the additional assumption that the change over time in the catchment of interest is from one steady state to another steady state. By steady state, we mean that any transient changes in storage can be ignored [e.g., see *Li et al.*, 2007, Figure 3]. With that assumption, the change in Q is given by

$$dQ = dP - dE \tag{6}$$

Combining equations (4) and (6), we have the following expression for the change in Q

$$dQ = \left(1 - \frac{\partial E}{\partial P}\right) dP - \frac{\partial E}{\partial E_o} dE_o - \frac{\partial E}{\partial n} dn \tag{7}$$

and the relative change in Q is

$$\frac{dQ}{Q} = \left[\frac{P}{Q}\left(1 - \frac{\partial E}{\partial P}\right)\right] \frac{dP}{P} - \left[\frac{E_o}{Q}\frac{\partial E}{\partial E_o}\right] \frac{dE_o}{E_o} - \left[\frac{n}{Q}\frac{\partial E}{\partial n}\right] \frac{dn}{n} \tag{8}$$

The terms in square brackets can be called the sensitivity coefficients and are analytical expressions formally equivalent to the "elasticity" concept [*Schaake and Liu*, 1989]. Previously, the sensitivity coefficients have been estimated empirically [e.g., *Sankarasubramanian et al.*, 2001; *Chiew*, 2006]. In contrast, the above theory derives analytical expressions for the sensitivity coefficients that are functions of the existing climate (P, E_o) and catchment properties (n). (See section 6.1 for a full discussion about the interpretation of the catchment properties parameter, n.)

3. Climate Data

[11] To apply the theory formulated here, we use publicly available long-term hydroclimatic data for the Murray-

Table 1. Estimates of Precipitation, Runoff, and Pan Evaporation for the Murray-Darling Basin

Climate Variable	Period	Data Source
Precipitation (mm yr^{-1})		
457	1895–2006	*CSIRO* [2008], *Chiew et al.* [2008] (scenario A)
440	1997–2006	*CSIRO* [2008], *Chiew et al.* [2008] (scenario B)
Runoff (mm yr^{-1})		
27.3	1895–2006	*CSIRO* [2008], *Chiew et al.* [2008] (scenario A)
21.7	1997–2006	*CSIRO* [2008], *Chiew et al.* [2008] (scenario B)
Pan Evaporation (mm yr^{-1})		
2121	1975–2006	Bureau of Meteorology[a]
2149	1997–2006	Bureau of Meteorology[a]

[a]From http://www.bom.gov.au (accessed July 2010).

Darling Basin (area of 1,060,000 km^2) in southeast Australia. The precipitation and runoff estimates are sourced from the Commonwealth Scientific and Industrial Research Organisation (CSIRO) Sustainable Yields Project [*CSIRO*, 2008; *Chiew et al.*, 2008; *Jones et al.*, 2009]. The runoff estimate is based on modeling by CSIRO that used historic climate data and current (2006) levels of water resource development. The long-term catchment-wide integrals span both the instrumental record (1895–2006) and drier conditions over the last decade of that period (1997–2006) (Table 1).

[12] The evaporative demand, E_o has traditionally been set to be equivalent to the net radiation [e.g., *Choudhury*, 1999]. On that basis, the Budyko equation partitions the precipitation between E and Q and simultaneously partitions the net radiation between evapotranspiration and the sensible heat flux. The practical difficulty is that net radiation is not a strictly climatological phenomenon because it also depends on surface albedo (i.e., reflected short-wave radiation) and surface temperature (i.e., emitted long-wave radiation). Consequently, while net radiation estimates are available from research in some regions [e.g., *Donohue et al.*, 2010], those estimates are not as yet part of routine monitoring networks. The one measure of evaporative demand that is routinely observed in many regions is pan evaporation [*Stanhill*, 2002; *Rose*, 2004; *Roderick et al.*, 2009a, 2009b] and we use that measure here. Routine recording of class A pan evaporation by the Bureau of Meteorology began in the early 1970s and in the absence of an alternative we use the spatially integrated estimate of pan evaporation for the MDB [*Jovanovic et al.*, 2008] for 1975–2006 as an estimate for the entire instrumental record. As will be evident later, the resulting conclusions are not particularly sensitive to that assumption since E_o is much larger than P in the MDB. We also use the same source to estimate pan evaporation for the 1997–2006 period (Table 1). We note that the numerical value of the catchment properties parameter, n, will depend on the measure for E_o that is adopted.

[13] Evaporative demand (E_o) is estimated using the pan evaporation measurements as follows:

$$E_o = k\ E_{pan} \tag{9}$$

The pan coefficient (k) is traditionally set to be 0.70 [*Stanhill*, 1976]. Here k is set to be 0.75 to account for the 7% reduction in pan evaporation due to the bird guards used on Australian pans [*van Dijk*, 1985]. In principle, E_o can change because of a change in E_{pan} [e.g., *Roderick et al.*, 2009a, 2009b] or because of a change in k [*Shuttleworth et al.*, 2009]. In the absence of better knowledge, we assume here that k remains constant over time.

4. Applying the Theory: The Murray-Darling Basin

4.1. Steady State Water Balance: MDB, 1895–2006

[14] Here we use the "Budyko equation" (equation (1)) to estimate E and Q for the 1895–2006 period and compare the estimate with observations. The calculations are summarized in Table 2.

[15] The resulting estimate of E (432.1 mm yr^{-1}) over the period 1895–2006 is within 1% of the observed value (429.7 mm yr^{-1}) when using a default value (n = 1.8) [*Choudhury*, 1999] for the catchment properties parameter. The subsequent estimate for Q was 24.9 mm yr^{-1} (assuming steady state) and within 10% of the observed value

Table 2. Comparison of the Long-Term (1895–2006) Annual Average Runoff Calculated Using the Budyko Equation (Equation (1)) With Observations[a]

Variable	Value	Comments
Data (1895–2006, MDB)		
P (mm yr^{-1})	457	Table 1
E_{pan} (mm yr^{-1})	2121	Table 1 (assume 1895–2006 = 1975–2006 average)
n	1.8	default value [*Choudhury*, 1999]
Calculations		
E_o (mm yr^{-1})	1591	equation (9), k = 0.75
E (mm yr^{-1})	432.1	equation (1)
$P - E$ (mm yr^{-1})	24.9	
Observed runoff		
Q (mm yr^{-1})	27.3	Table 1

[a]Note that for exact agreement with the observed runoff, n has the value 1.745.

Table 3. Comparison Between Theoretical Predictions and Observations for the 1997–2006 Annual Average Runoff for the Murray-Darling Basin[a]

Variable	Value	Comments
Data (1997–2006, MDB)		
P (mm yr^{-1})	440	Table 1
E_o (mm yr^{-1})	1612	Table 1 and equation (9), $E_o = 0.75 \times 2149$ mm yr^{-1}
n	1.745	Table 2 caption, "tuned value"
Calculations		
dn	0	assume no change in catchment properties
dP (mm yr^{-1})	−17	$= 440 - 457$
dE_o (mm yr^{-1})	+21	$= 1612 - 1591$
dE (mm yr^{-1})	−13.8	equation (11), $= 0.845\ (-17) + (0.028)\ (21) = -14.4 + 0.6$
dQ (mm yr^{-1})	−3.2	equation (6), $dQ = dP - dE$
Q (mm yr^{-1})	24.1	Q (1997–2006) = Q (1895–2006) + dQ = 27.3 − 3.2 = 24.1
Observed Runoff (1997–2006)		
Q (mm yr^{-1})	21.7	Table 1

[a]MDB, Murray-Darling Basin.

(27.3 mm yr^{-1}) (Table 2). To get exact agreement with observations requires n to be 1.745. We use that "tuned" value in subsequent calculations.

4.2. Calculating the Sensitivity Coefficients for the MDB

[16] The change in Q as a function of change in climate (P, E_o) and catchment properties (n) can be calculated (equations (6)–(8)) once the numerical values of the sensitivity coefficients (equation (5)) are known. Using the long-term (1895–2006) data for the MDB (see Table 2; P = 457 mm yr^{-1}, E_o = 1591 mm yr^{-1}, Q = 27.3 mm yr^{-1}, $E = P - Q$ = 429.7 mm yr^{-1}, n = 1.745) the calculated sensitivity coefficients for the MDB are

$$\frac{\partial E}{\partial P} = 0.845; \quad \frac{\partial E}{\partial E_o} = 0.028; \quad \frac{\partial E}{\partial n} = 46.4 \tag{10}$$

Substituting those values into equation (4), the perturbation in E due to changes in climate (dP, dE_o) and catchment properties (dn) in the MDB is

$$dE = 0.845\ dP + 0.028\ dE_o + 46.4\ dn \tag{11}$$

The associated change in Q (equation (7)) is

$$\begin{aligned} dQ &= (1 - 0.845)dP - 0.028\ dE_o - 46.4\ dn \\ &= 0.155\ dP - 0.028\ dE_o - 46.4\ dn \end{aligned} \tag{12}$$

and the relative change in Q (equation (8)) is

$$\frac{dQ}{Q} = 2.6\ \frac{dP}{P} - 1.6\ \frac{dE_o}{E_o} - 3.0\ \frac{dn}{n} \tag{13}$$

Equation (11) predicts that in the MDB, E will be much more sensitive to a change in P (= 0.845 dP) than to a comparable change in E_o (= 0.028 dE_o). This result is typical of arid environments ($P \ll E_o$) where in the long term, E is largely controlled by water availability (P) [*Roderick et al.*, 2009b]. Q (equation (12)) is also predicted to be more sensitive to a change in P (= 0.155 dP) than to comparable changes in E_o (= 0.028 dE_o) although the contrast between the two competing controls is much reduced for Q compared to E.

[17] The final theoretical result for the MDB (equation (13)) predicts that a 10% increase in P would increase Q by around 26% while a 10% increase in E_o would decrease Q by 16%. For comparison purposes, the CSIRO Sustainable Yields Project [*CSIRO*, 2008] used a hydrologic model calibrated over the MDB [*Chiew et al.*, 2009] to predict that a 13% reduction in P (by the year 2030) would decrease Q by 33% (implying a sensitivity of ~2.5) while an 8% increase in P (by 2030) would increase Q by 16% (implying a sensitivity of ~2.0) [*CSIRO*, 2008]. Our analytical results are broadly consistent with that range. We apply the analytical theory further in section 4.3.

4.3. Response to Climate Perturbation: MDB, 1997–2006

[18] We test the theoretical expressions (equations (11) and (12)) by comparing the predicted perturbation in Q for the 1997–2006 period with observations (Table 3). The predicted value for Q (24.1 mm yr^{-1}) is close to the observed (21.7 mm yr^{-1}) and the small difference (2.4 mm yr^{-1}) is probably within the bounds of measurement error and/or overall uncertainty of the technique. Nevertheless, recall that the calculations assume a change from one steady state to another steady state. An alternate interpretation is that the difference implies a storage change, equivalent to +24 mm over the 10 year period. While plausible, there are no data to evaluate the magnitude of the indicated storage term. Gravity satellite and field hydrological data suggest a decline in storage in the MDB over the period 2001–2006 of ~100 mm [*Leblanc et al.*, 2009]. Hence, for the calculations to be correct, there would need to have been an increase in storage of a comparable amount at the start of the period (1997–2001). Again, that is certainly plausible but currently unknown. An alternative interpretation is that there was no change in storage but that the catchment properties (n) changed slightly. In future, as longer-term gravity satellite measurements become routinely available, it should be possible to distinguish between changes in storage and changes in catchment properties when comparing the observed and predicted Q. We cannot do that here because of the short length (it began in 2002) of the gravity satellite record.

[19] The encouraging results demonstrate that it is feasible to construct a quantitative theoretical framework that uses readily available climate data to examine the overall sensitivity of the catchment water balance to a change in the climate. We use that knowledge in subsequent sections to examine the general sensitivity of the MDB catchment

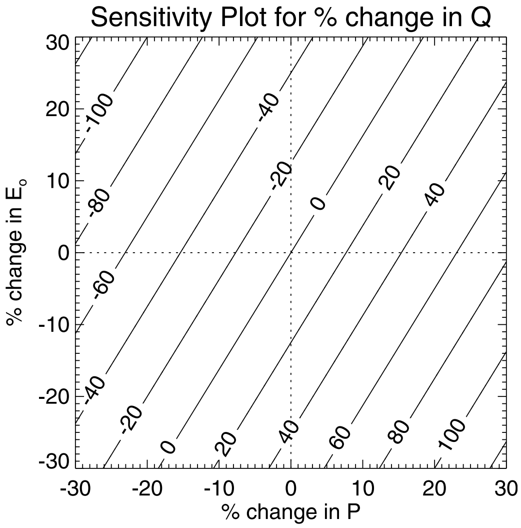

Figure 2. Sensitivity of the runoff (Q) to changes in climate in the Murray-Darling Basin (MDB) (equation (13)). Contours show the percent change in Q as a function of percent changes in precipitation (P) and evaporative demand (E_o). Dotted lines show the lines of no change with the contour interval set at 20% change in runoff. The calculations assume steady state conditions with no change in catchment properties (dn = 0).

runoff to changes in climate (section 5) and catchment properties (section 6).

5. Change in Runoff due to Change in Climate

5.1. Sensitivity of Runoff to Changes in Climate in the MDB

[20] We use the previously developed theory to calculate the expected change in runoff as a function of changes in precipitation and evaporative demand for the MDB (Figure 2). Note that in terms of the theory presented here, Figure 2 represents some climate change possibilities (i.e., the changes in Q due to changes in P and E_o) in a single diagram.

5.2. Incorporating Climate Change Scenarios

5.2.1. The Ideal Approach

[21] Scenarios of future climate can be incorporated as an overlay on the base sensitivity figure (Figure 2). In the ideal situation we would start with the simulated change in P and E_o extracted from an individual run of a climate model with E_o calculated from the model output using the PenPan model to estimate pan evaporation [*Rotstayn et al.*, 2006; *Roderick et al.*, 2007]. Then the change in P and E_o from each climate model run would be plotted as a single realization on Figure 2. The collection of such points from all climate model runs would define an "ensemble" of hydroclimatic projections, and would show, at a single glance, the range of the various projections of runoff. This approach recognizes that in water-limited environments, the change in P is often, but not always, negatively correlated with the change in E_o [*Roderick et al.*, 2009b]. Unfortunately, the current situation is not ideal. Suitable projections for P are readily available for the globe [*Lim and Roderick*, 2009] but equivalent projections are not yet available for E_o. Consequently, we have constructed an interim scenario to demonstrate the basic concept.

5.2.2. Interim Climate Change Scenario for the MDB

[22] For the change in P we followed the simplest and most transparent approach. We used climate model projections based on the Intergovernmental Panel on Climate Change (IPCC) A1B emissions scenario and took the difference (2070–2099 less 1970–1999) in P for each of the 39 available model runs [*Lim and Roderick*, 2009] extracted for the MDB [*Sun et al.*, 2011]. That difference was converted to a percentage change using the observed precipitation for 1970–1999 (= 517.4 mm yr^{-1}). We could have used alternative approaches, e.g., by using bias-corrected output, but the resulting scenarios are more or less unchanged [*Sun et al.*, 2011]. The final precipitation scenario based on that approach is summarized in Table 4. Note that the scenario is reasonably symmetric, i.e., a roughly equal number of model runs show increases (22 out of 39) as show decreases (17 out of 39) in precipitation and the average across all model runs (+1%) shows basically no change.

[23] To construct the scenario for evaporative demand (Table 4), we are fortunate in that the first study to examine pan evaporation projections using global climate model output was for Australia [*Johnson and Sharma*, 2010]. That study used the PenPan model [*Rotstayn et al.*, 2006; *Roderick et al.*, 2007] to calculate pan evaporation using the output from four climate models for the A2 emissions scenario and output from five models for the B1 scenario. Importantly, they also compared pan evaporation estimates (from five climate models) with Australian observations over the period, 1975–1999. Observations over that period show a general decline in evaporative demand over much of Australia [*Roderick and Farquhar*, 2004; *Rayner*, 2007; *Roderick et al.*, 2007; *Jovanovic et al.*, 2008] that was not, in general, simulated correctly by the climate models. However, there was a single noteworthy exception, the CSIRO (Mk3.5) climate model. Calculations using output from a single run of that model did reproduce the observed decline in evaporative demand, both in terms of continental averages as well as the broad spatial patterns. Further, observations show that wind speed decline has been the major contributor to declining evaporative demand over much of Australia in that period [*Roderick and Farquhar*, 2006; *Rayner*, 2007; *Roderick et al.*, 2007; *McVicar et al.*, 2008] and the CSIRO model also correctly simulated that as well. Given the apparent success of that model, one could

Table 4. Interim Climate Change Scenarios for the MDB Based on Global Climate Model Output for the A1B Scenario[a]

	Percent Change in P	Percent Change in E_o
Minimum	−31	+1
10th percentile	−15	
Mean	+1	+7
Median	+2	
90th percentile	+15	
Maximum	+30	+10

[a]For 2070–2099 (A1B scenario) relative to 1970–1999. The change in precipitation (P) is derived from the 39 Intergovernmental Panel on Climate Change AR4 climate model runs reported by *Sun et al.* [2011]. The change in evaporative demand (E_o) is an interim scenario derived from an analysis of the output from four climate models [*Johnson and Sharma*, 2010, Figure 7]. See the text for a more detailed description of the interim E_o scenario.

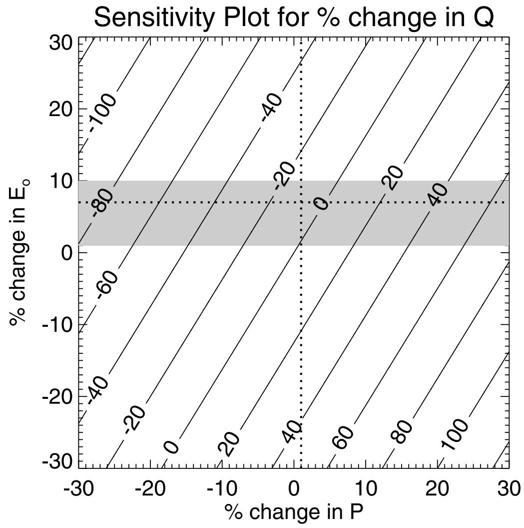

Figure 3. Interim climate change scenario for changes over the next 100 years (P, E_o in 2070–2099 relative to 1970–1999) (shaded) overlaid on the MDB runoff sensitivity plot. Dotted lines denote the mean change in the two climate variables (Table 4).

make an argument for using it alone to develop the E_o scenario. However, we take a conservative approach and use the results from all models.

[24] The E_o scenario should use the same A1B emission scenario used for P (A1B). However, that was not available so we used previous results for the A2 emissions scenario that were based on the output of four climate models [*Johnson and Sharma*, 2010, Figure 7]. Future research will be needed to compile estimates of E_o using the 20 major climate models and the A1B emissions scenario to complement the scenario for P.

[25] The sensitivity of Q to changes in climate (Figure 2) has been combined with the interim climate change scenario (Table 4) in Figure 3. The large range in possible future hydroclimatic conditions (shaded area in Figure 3) is a consequence of the large range in the local scale hydroclimatic projections made by different models for the MDB [*Sun et al.*, 2011]. In general, all climate models predict more or less the same globally averaged increases in P as the earth warms [*Held and Soden*, 2006; *Lim and Roderick*, 2009]. However, at regional scales, the predicted changes in P are highly variable between different models, and for some models, between different runs of the same model [*Rotstayn et al.*, 2007; *Johnson and Sharma*, 2009; *Lim and Roderick*, 2009]. With that in mind, we note that the "average change," i.e., a 1% increase in precipitation coupled with a 7% increase in evaporative demand, would, according to the theory, lead to a 9% decline in runoff. However, we emphasize that this must be considered an interim projection that will be updated once full scenarios for E_o become available.

6. Change in Runoff due to Change in Catchment Properties

[26] The predictions for the impact of changes in climate on Q (Figures 2 and 3) assume a transition from one steady state to another with no change in catchment properties. In this section we describe how changes in catchment properties impact on E and Q under an assumed constant climate. However, we are unable at this stage to provide quantitative estimates for the impact of changes in various catchment properties on runoff and instead focus on qualitative changes. This section is organized as follows: we first explain the theoretical meaning of the catchment properties parameter, and following that, we provide qualitative estimates (i.e., increase or decrease in n) for some major factors relevant to the MDB and summarize all key results.

6.1. The Meaning of the Catchment Properties Parameter in This Theoretical Framework

[27] The parameter that encodes catchment properties, denoted n (see equation (1) and Figure 1) modifies the partitioning of P between E and Q. In the formulation used here, it encodes *all* factors that change the partitioning of P between E and Q under a constant climate. Hence, the numerical value of n encodes catchment properties long known to influence runoff generation such as topography (e.g., slope, aspect, etc.), soil type/depth, geologic substrate, vegetation, etc. [*Yang et al.*, 2007, 2008]. Nevertheless, for a given catchment, many of the properties, especially the geologic and topographic properties, will remain nearly constant over decadal to century time scales. In contrast, vegetation can readily change over decades to centuries. For example, according to *Zhang et al.* [2001], a change in vegetation cover from grasses to trees would increase n, and in a constant climate, E would increase while Q would decrease. Hence, the hydrologic impact of changes in land cover [*Brown et al.*, 2005; *Jackson et al.*, 2005] as well as changes in vegetation water use, with, for example, increasing atmospheric [CO_2] [*Farquhar*, 1997; *Gedney et al.*, 2006] are also incorporated into the framework by changes in the value of n.

[28] However, the change over time in n will also depend on factors like changes in precipitation intensity or changes in the spatial distribution and/or seasonal timing of P and E_o (see section 6.2). We highlight those latter changes here to emphasize the point that changes in the catchment parameter may be subtle, and not strictly separated from, for example, intra-annual climatic effects.

[29] In sections 6.2–6.5 we provide qualitative estimates (i.e., increase or decrease in n) for some of the major factors that can change the catchment properties over decadal to century time scales. In section 6.6, we summarize the overall impact on MDB hydrologic function of the major factors considered here.

6.2. Changes in Spatial and Temporal Distribution of Rainfall and Evaporative Demand in the MDB

[30] To give a concrete example, the hydroclimatic regime of the MDB is characterized by an aridity gradient running from the arid northwest (low P, high E_o) to the humid southeast (high P, low E_o) [*CSIRO*, 2008]. Further, in accord with the theory (partial differentials in equation (4)) we assume no change in the climate (i.e., dP = 0, dE_o = 0) when integrated over the entire basin. Now assume that P increased in the arid northwest but that this increase was offset by an equal and opposite decrease in the humid southeast. With all else equal, the increase in P in the arid parts would lead to a small increase in Q, while the decrease in P in the humid parts would lead to a larger (in an absolute

Table 5. Summary of Major Factors Leading to a Change in Catchment Properties in the MDB and the Associated Changes on Catchment Properties, Evapotranspiration, and Runoff Assuming a Fixed Basin-Wide Average Climate[a]

Factors Leading to Changes in Catchment Properties	*n*	*E*	*Q*
Change in spatial distribution of P[b]			
Increased P in north and complementary decrease in south	+	+	−
Reverse: decreased P in north and increase in south	−	−	+
Change in temporal distribution of P[b]			
Increased P in summer and complementary decrease in winter	+	+	−
Reverse: decreased P in summer and increase in winter	−	−	+
Change in spatial distribution of E_o[b]			
Increased E_o in north and complementary decrease in south	−	−	+
Reverse: decreased E_o in north and increase in south	+	+	−
Change in temporal distribution of E_o[b]			
Increased E_o in summer and complementary decrease in winter	−	−	+
Reverse: decreased E_o in summer and increase in winter	+	+	−
Change in fire frequency[c]			
Increased fuel loads leading to increased bushfires in southeast forests	+	+	−
Reverse: decreased fuel loads leading to decreased bushfires in southeast forests	−	−	+
Impact of elevated [CO_2] on plant water use[d]			
Decreased stomatal opening	−	−	+
Increased leaf area	+	+	−

[a]Here n, catchment properties; E, evapotranspiration; Q, runoff. A plus indicates an increase, and a minus indicates a decrease.
[b]See section 6.2.
[c]See section 6.4.
[d]See section 6.5.

sense) decrease in Q. Hence in this hypothetical example, when integrated over the basin, E would increase and Q would decline (i.e., $dn > 0$), despite the fact that the overall catchment-wide average climate did not change. Analogous scenarios can be constructed for changes in the spatial distribution of E_o (see Table 5).

[31] The same underlying reasoning can be applied to seasonal changes. For example, in a constant catchment-wide climate, assume P increased in summer with an equal and opposite decrease in winter. Because E_o is generally higher in summer than winter, such a change would lead to enhanced E and reduced Q (i.e., $dn > 0$). All possible combinations of these space/time changes are summarized in Table 5.

6.3. Impact of Changes in Land Use

[32] The widespread replacement of perennial trees/shrubs with (mostly) annual crops and grasses for agriculture in the MDB over the last 100 years is believed to have led to rising water tables because of reduced E by the (mostly annual) agricultural species [*Nulsen et al.*, 1986; *Cartwright and Simmonds*, 2008]. The reverse situation should also hold: tree planting programs, whether for commercial forestry, carbon sequestration or other purposes may increase E and decrease Q (i.e., $dn > 0$) [*Zhang et al.*, 2001; *Brown et al.*, 2005; *Jackson et al.*, 2005]. Hence, changes in land use that result in a change in land cover from grasses (e.g., agricultural crops) to trees or vice versa will, no doubt, be important to the water balance in local catchments. Whether the changes become significant at the scale of the MDB would depend on the area involved. Current estimates for the future expansion of tree plantations (of order 10,000 to 100,000 ha) [*CSIRO*, 2008] are too small to have a significant MDB-wide impact on Q.

[33] The above analysis of land use/land cover change is subject to an important caveat. The impact of changes in land cover on E and Q assumed that P and E_o remain constant. Studies with climate models have reported that land clearing can lead to changes in the local climate [*Pitman et al.*, 2004; *McAlpine et al.*, 2007]. This represents a possible feedback between the catchment properties parameter (n) and the climate (P, E_o). In the framework used here, the feedback to climate would be included in the climate change projections (Table 4). However, none of the climate models used in the IPCC AR4 explicitly account for this type of change. The treatment of land surface feedbacks to the climate is an area of active and ongoing research.

6.4. Impact of Fire

[34] Like land use change, fire is another pervasive disturbance in many regions. Previous research has documented substantial declines in Q after extensive wildfires in Mountain Ash catchments in Victoria [*Kuczera*, 1987] and similar results have been found in some other Eucalypt forests [*Cornish and Vertessy*, 2001]. That pattern is for the maximum decline in Q, and hence a maximum increase in E (both expressed as fractions of P) to be reached a few decades after a major fire. Following that, E and Q are predicted to gradually return to the prefire levels over time scales of ~100–200 years that more or less follow the pattern of ecological succession in those forests. Although there are many potential possibilities, the underlying biological-ecological-physical basis for those observed patterns remains unknown. Further, whether the pattern applies more widely to other Eucalypt forests (or indeed other forest types) also remains unknown. These are areas where research is urgently needed.

[35] The postbushfire runoff reduction phenomenon raises a more general question: what would be the impact of changed fire regimes on the water yield from forested catchments in southeast Australia or similar environments elsewhere? Of particular interest are wildfires in high water yielding catchments ($P > E_o$) located in the southeast parts of the MDB. This is a difficult question because fires are a complex phenomenon that also involve human behavior (e.g., fire suppression). In the most general sense, extensive

wildfires require fuel, suitable weather conditions and an ignition source. Of these, fuel availability is probably the most important in determining long-term trends. (Suitable weather for fires regularly occurs during summer and ignition sources, e.g., people and lightning, are readily available.) All else being equal, fuel availability is unlikely to decrease in future given that elevated [CO_2] often enhances photosynthetic rates [*Drake et al.*, 1997] (but see 6.5 for more discussion). If fuel availability remained (near) constant, then wildfire frequency may remain unchanged but any increase in fuel availability would likely increase fire frequency. Again, if the postbushfire runoff reduction phenomenon [*Kuczera*, 1987; *Cornish and Vertessy*, 2001] was found to be general, then the largest reduction in Q would occur if the landscape was transformed into a mosaic of regenerating forest patches with ages (i.e., time since burnt) of ~20–30 years. We are currently unable to be more quantitative and highlight the possible impact of changes in fire frequency in the summary (Table 5).

6.5. Impact of Elevated [CO_2] on Plant Water Use

[36] The relationship between plant water use and photosynthesis depends primarily on the level of atmospheric [CO_2] for a given leaf-to-air humidity difference [*Wong et al.*, 1979]. It is usually assumed that the relative humidity of the air will remain relatively constant in the future [*Held and Soden*, 2000, 2006]. Indeed, the majority of the future warming predicted by climate models in response to increasing greenhouse gases comes about via water vapor feedback that actually requires a near-constant relative humidity [*Held and Soden*, 2000]. Here we assume that the relative humidity of the near-surface air remains approximately constant as generally found in observations in the MDB [*Roderick et al.*, 2007] and worldwide [*Dai*, 2006]. On that assumption, higher levels of atmospheric [CO_2] imply that less transpiration is very likely for the same amount of photosynthesis. In energy-limited regions ($P > E_o$) we anticipate about the same amount of photosynthesis with less transpiration, while in water-limited regions ($P < E_o$), we anticipate the same amount of transpiration and hence more photosynthesis [*Farquhar*, 1997].

[37] Why the difference between energy- and water-limited regions? To give an example, assume a water-limited environment ($P < E_o$) like the northwest parts of the MDB. Further assume that under a constant climate ($dP = 0$, $dE_o = 0$), the transpiration per unit leaf area declines as atmospheric [CO_2] increases. In arid environments, water is an extremely limiting resource for plant growth and it is likely that the vegetation would still manage to extract and transpire most of the available water over a year or more. Hence with the climate assumed to be constant, the transpiration per unit of ground area would more or less remain constant while the transpiration per unit leaf area would decrease as atmospheric [CO_2] increases. From that combination, it follows that the leaf area per unit ground area will increase. For dry regions within the MDB, *Berry and Roderick* [2002] used similar logic and concluded that elevated [CO_2] would likely increase the evergreen cover (e.g., woody plants with long-lived leaves) at the expense of annual herbaceous vegetation. Recently, satellite data have shown relatively large increases in perennial (i.e., woody) vegetation cover across many parts of Australia since 1981 [*Donohue et al.*, 2009] in general accord with that prediction. In summary, for constant P and E_o, the above logic predicts near-constant E and Q ($dn \sim 0$). In such water-limited regions, even if there were small changes in E, the absolute changes in Q are expected to be small.

[38] What about the impact of changes in atmospheric [CO_2] on E and Q in the energy-limited ($P > E_o$) environments that generally have high water yields? In wetter environments, like the forested regions in the southeast of the MDB, the longer-term landscape scale [CO_2] response is more difficult to estimate. Again, for a given amount of photosynthesis per unit leaf area, the transpiration per unit leaf area should decrease as the atmospheric concentration of CO_2 increases, but in those humid environments, water is generally not the limiting factor for photosynthesis. Rather, it is light intensity and nutrients and the change in leaf area of the canopies is not as easy to predict. For example, low nutrient availability in alpine and other environments [*Costin*, 1954; *Roderick et al.*, 2000] can limit the photosynthetic response of vegetation to elevated [CO_2] [*Hungate et al.*, 2003; *Luo et al.*, 2004; *de Graaff et al.*, 2006]. In nutrient poor areas, where one might expect a limited response of photosynthetic rate to increasing [CO_2], we would also expect a reduction in transpiration rate per unit ground area with increasing [CO_2]. Hence, under those assumptions we expect, under a constant climate, for E to decrease and Q to increase ($dn < 0$) as atmospheric CO_2 increases, as appears to be happening in the humid environments of northern Europe [*Gedney et al.*, 2006].

6.6. Summary of Impacts in the MDB

[39] Hydrologic impacts of the various changes in catchment parameters discussed above are summarized in Table 5.

[40] In terms of a synthesis, it is useful to put the hydrologic impacts of changes in atmospheric [CO_2] and fire frequency within the context of major land uses within the MDB. The three major land use categories are dryland crops (~10% of MDB), dryland pastures (~67% of MDB) and native vegetation (~20% of MDB) [*CSIRO*, 2008]. In the dryland cropping regions, the transpiration response to elevated [CO_2] will depend to a large extent on the agronomic practices (e.g., plant nutrition, crop varieties, economics) and is difficult to forecast. At any rate, in an absolute sense, Q is small and fire infrequent in dryland regions. The Native vegetation category is largely public lands (e.g., national parks and state forests) along the wetter southeast rim of the basin. Most of the basin Q originates in this region and we anticipate that any basin-wide changes in Q will be dominated by changes occurring in this land use category. With all else equal, we expect elevated [CO_2] to decrease E and increase Q ($dn < 0$) while we anticipate that any increase in fire frequency would increase E and decrease Q ($dn > 0$). Which of these two competing processes would dominate the final integrated response is currently unknown.

7. Discussion

[41] The Budyko equation used here expresses the evapotranspiration (E) as a function of two climate variables, precipitation (P) and evaporative demand (E_o) and a third parameter that represents catchment properties (n). For practical reasons, we estimated the evaporative demand using pan evaporation data. We found that the “tuned value”

of the catchment properties parameter, n (= 1.745) needed for exact agreement with observations for the Murray-Darling Basin (MDB) was well within the range previously found for that variable.

[42] We use the theory to develop a simple and novel expression for the effects on E and Q of small perturbations in P, E_o, and n. This theoretical approach is first used to construct sensitivity surfaces for the change in runoff as a function of changes in climate assuming no change in catchment properties. We also describe, qualitatively, why the catchment properties parameter will change over time. Some elements of this approach have been used previously although not in the same form as presented here. For example, many studies, often using slightly different variants of the Budyko equation, have assessed the sensitivity of Q to changes in the same two climate variables [e.g., *Dooge et al.*, 1999; *Milly and Dunne*, 2002; *Zhang et al.*, 2008]. Further, the inclusion of changes in catchment properties over time appears to have been first considered by *Yang et al.* [2006] but in that study they assumed that the catchment property term would remain constant over time and did not calculate the associated sensitivity coefficient. In that respect, *Sun* [2007] appears to have been the first to present a complete analytical treatment that was based on a perturbation of Fu's equation (equation (2)), but that work also did not consider how the catchment property parameter might change over time.

[43] The key result of the perturbation approach used here is the following prediction for the impact of changes in climate (P, E_o) in the MDB: a 10% increase (decrease) in P should increase (decrease) Q by 26%, while a 10% increase (decrease) in E_o should decrease (increase) Q by 16%. Previous empirical studies [*Chiew*, 2006; *Potter et al.*, 2008] and detailed simulation modeling [*Chiew et al.*, 2009] have given broadly similar results [*CSIRO*, 2008]. That agreement offers encouragement that the overall analytical approach developed here can be readily applied elsewhere.

[44] A scenario of future climate (2070–2099) for the MDB was developed by combining the estimates for P and E_o from models run for the IPCC Fourth Assessment Report. The range in P across the 39 available model runs is large, with 22 showing increases and 17 showing decreases under the A1B emissions scenario, compared to the modeled present day [*Sun et al.*, 2011]. Taking an average of the predicted change across all model runs gave an increase of 1% (5 mm yr^{-1}) but we emphasize the prediction bounds encompass a large range (about ±30% change in P) that reflect real uncertainty about future hydrologic conditions at regional scales [*Johnson and Sharma*, 2009; *Lim and Roderick*, 2009; *Chiew et al.*, 2011]. The scenario for evaporative demand was not ideal because we only had access to the output from four climate models based on the IPCC A2 emissions scenario [*Johnson and Sharma*, 2010] instead of the A1B scenario used for P. Hence, the E_o scenario is best described as an interim attempt that will be refined in future. We suggest the approach (see section 5.2.1) where the future values of P and E_o are considered to be a sole realization from a given climate model run. This will account for the physical fact, that, in many arid and semiarid regions, an increase in P is often accompanied by a decrease in E_o (and vice versa) [*Brutsaert and Parlange*, 1998; *Hobbins et al.*, 2004; *Shuttleworth et al.*, 2009] although we emphasize that this correlation is not universal [*Roderick et al.*, 2009b].

[45] It is often implicitly assumed that all of the major changes in catchment runoff are due to changes in climate. That is highly topical in the MDB given the drop in runoff observed in the southern parts over the last decade [*Potter et al.*, 2010]. While climatic factors have obvious and well known importance, they are not the only ones that can lead to changes in runoff and water availability. For example, vegetation changes over decadal to century time scales can have marked hydrologic changes [*Brown et al.*, 2005; *Jackson et al.*, 2005]. The Kuczera curves showing a drop in Q, perhaps as high as 50% or more, some 20–30 years after bushfires in Mountain Ash catchments in Victoria [*Kuczera*, 1987] are an extreme example of the role of vegetation change. In that respect we note that extensive wildfires occurred throughout the southern MDB in January 2003. Perhaps this has also, in part, contributed to the drop in runoff in the southern MDB. If true, then the Kuczera curves suggest it might well continue for another decade or two.

[46] The current study used a very simple form of the Budyko equation that included two climatic variables (P, E_o) and one parameter (n). While simple, the resulting analytical models are easy to understand. Nevertheless, it may prove useful to add another term to the model that can better account for phase differences between P and E_o via catchment storage [*Milly*, 1994; *Rodriguez-Iturbe and Porporato*, 2004]. Even without that extension, the original Budyko framework, in its modern formulation, has turned out to be very useful tool for making simple and transparent estimates of changes in water availability.

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