Dear Dr. Andréassian:

We have completed revision of our manuscript entitled “The Budyko and complementary relationships in an idealized model of large-scale land–atmosphere coupling” (hessd-11-9435-2014). We wish to thank the two reviewers for their insightful and constructive comments. Based on their input, we believe the quality of our manuscript is improved. Below we provide point-by-point responses to each of the reviewer’s comments. We appreciate your consideration of our article for publication in HESS and look forward to your response.

Sincerely,

Benjamin R. Lintner

Reviewer #1

We thank Reviewer #1 for the careful and thorough reading of our manuscript, especially with respect to clarifying language and methodological details.

• P9440: did Morton (1983) provide the justification for the complementary relationship over long time scales? I thought the challenge is show why the relationship holds on daily time scale.
We have revised the language here to eliminate the line in question.

• P9443: It is not clear how the relationships shown in Figure 1a were obtained and I assume they are from a model? In that case, it would be helpful to provide a brief description of the model.
The model used to produce Figure 1 is just the prototype outlined in Section 2. However, given Reviewer #1’s comment, we have added some text for further clarity.

• P9444: If the good correspondence may be coincidental, what these results mean in terms of the prototype?
We have clarified and added to the discussion: the text now reads:

> The prototype complementary relationship shows a qualitative, and arguably quantitative, correspondence with both observational datasets. Of course, we should point out that the prototype was not explicitly tuned to represent the hydroclimate of these locations, and as such, any quantitative correspondence may be coincidental. Moreover, the scatter inherent in the observations would permit a range of plausible complementary relationships.

• “A slight decrease in E is observed for W> 0.7”. I am not sure what the authors are referring to here? Please explain.
We refer here to the slight downward bend in the gray curve in Figure 1a. We have added a parenthetical note to this effect following this sentence.

• “We also note that at low soil moisture values, precipitation essentially mimics the E response..”. I find this statement confusing. I thought under low soil moisture values, E would mimic precipitation. I can’t understand why this would explain a complementary relationship between precipitation and potential evaporation.

We simply intended to point out that because E must balance P at low soil moisture, and E is complementary to Ep, then P is complementary to Ep. Given that this may be obvious, we have decided to remove the sentence in question.

• P9445: "the prototype's complementary relationship", is this different from Bouchet's complementary relationship?
The language was meant to convey the complementary relationship as it occurs in our prototype, but we are using complementary relationship in the sense of Bouchet. We have revised the language to (hopefully) clarify.

• P9447: what is η? What is α? I find the description of the Budyko curve is confusing.
The authors frequently refer to the prototype and it would be helpful if this can be avoided.
The parameters used are summarized in Table 1: α is the Priestley-Taylor coefficient while η is the power law scaling exponent for runoff dependence on soil moisture. However, we should apologize for an error in this section in which we used $\alpha = 4$ instead of $\eta=4$.

• What is the implied slope of Ep vs. E? Is this the line shown in Figure 1b? What is the significance of the slope?
The Reviewer is correct about the line in 1b representing the slope of Ep vs. E. The significance here is that the value of the slope represents a convenient measure for the change in the complementary relationship.

• P9449: "In contrast to the complementary relationship, the Budyko curve is extremely robust, with no apparent change in the shape for these variations". The Budyko curve (Budyko, 1974) has no parameter and it is simply a function of the dryness index, but other similar equations can vary depending on the model parameter. The authors seem to suggest that the complementary relationship is not robust. Any evidences to support this claim?
Both Reviewers have rightfully identified some deficiencies in our discussion of the Budyko curve. As far as Reviewer 1's point, we used the term robust to describe the Budyko curve when in fact we meant insensitive to parameter change, following the reasoning noted by the reviewer. We have altered the text of this part to reflect a more proper characterization of the nature of the Budyko curve. (We still think it's important to point this out, which is why we haven't completely removed the discussion.)

• P9450: "Rather than present the complementary relationship..., we instead show the surface temperature and specific humidity profile as functions of soil moisture". Does this mean the surface temperature vs soil moisture relationship can be used to represent the complementary relationship? I find it difficult to follow the discussion. What is the fixed Ep case?
In these experiments, by prescribing either the evaporative efficiency β or the potential evapotranspiration *Ep*, we effectively disable the complementary relationship. That said, we can still compare the behavior of surface temperature and specific humidity in the baseline and intervention experiments.

• "This in turn feeds back onto precipitation... with increase water vapour and connective cloudiness". Is this result of the model (i.e. the prototype)? Or this is just a general statement?

This behavior is inherent in the physics of the prototype, which we expect should be generally representative of the leading-order large-scale thermodynamic response (especially in the tropics) to warming. Of course, some important aspects are not captured by the prototype, such as changes to convective inhibition, which may have a strong impact on the triggering/frequency of convection.

• P9542: Why would E increase under warming at low soil moisture and decrease for soil moisture above 0.5? What are the mechanisms for such changes in E?
At low soil moisture, the vertically-averaged atmospheric moisture balance is between precipitation and evapotranspiration. At higher soil moisture the balance involves a progressively larger moisture convergence term.

• P9454: The authors stated that they derived analytic expressions for the Budyko and complementary relationships based on an idealised prototype. Do the authors mean that they have derived Equation (11) and Equation (15)?
Upon reflection, we agree with Reviewer #1 (and Reviewer #2)'s comment, namely that we do not "derive" the Budyko relationship, given that Budyko is inherent from the way runoff and evapotranspiration are formulated in our prototype. We have hopefully adjusted the language to the satisfaction of both Reviewers.

Reviewer #2
We thank Reviewer #2 for a thought-provoking review. We also appreciate the encouragement and support in Reviewer #2's opening summary of our manuscript.

• Expression for Budyko relationship
In the abstract the authors claim to derive the Budyko relationship from an idealized prototype for Large-Scale Land–Atmosphere Coupling.
To my understanding, please correct me if I am wrong, the form of the Budyko curve
is already determined by prescribing runoff as function of soil moisture as done on P9442L9: Q = Pwη = P E η. Hence, although precipitation P and potential evaporation Ep are model outcomes, they do not change the form of the Budyko curve (Fig.8), because E seems to be fixed by the prescribed runoff power law. Thus only changes in the runoff function change the form of the Budyko curve (Fig. 3). Hence, as the model is set up, one might conclude (as the authors do) that land-surface interaction does not change the form of the Budyko curve.

While this may be true for scenarios where one can assume a steady function for runoff without changes in the storage discharge relation and without vegetation adaptation, it is certainly not true for changes in the frequency and intensity of precipitation (Milly, 1994; Porporato et al., 2004; Donohue et al., 2012).

Given the model setup and especially Section 3.2, it seems to me that the Budyko relationship is prescribed, rather then derived. Please comment on why the Budyko relationship is thought to be derived from the prototype.
As we noted above, Reviewer #1 had similar questions regarding the way in which we either framed our analysis of the Budyko curve or interpreted results.

• As a side note, it would be quite interesting to see, if changes in the parameterization of the runoff power law (which was done for Figure 3) would effect the CR?

• Decline of E at large soil moisture
The authors report on the behavior that E is declining at higher soil moisture values (P9444L3ff, Fig. 1 and Fig 2). Could it be that this behavior is due to the case that under high soil moisture conditions we typically find humid conditions which are energy limited? Hence $E \leq Ep$ for any case. Such a behavior can simply be demonstrated with the Budyko water-energy framework (using Budyko's curve or Eq. 15) e.g. forced with $P = 1$ and $Ep = 0...10$ and soil moisture $w = E/Ep$ this can be plotted as done in Fig.1a. At high soil moisture, E must converge to Ep, and will decline towards Ep(w).
We agree with Reviewer #2 that this behavior is consistent with energy limitation under humid conditions, and we thank Reviewer #2 for pointing out the connection of the Budyko framework. We have added a brief discussion of this at the end of Section 3b.

• Model configuration and replication of results
The authors should provide all details, necessary to replicate the results, possibly in an appendix or supplement. This is not only required for scientific reasons, but also for intended use of the model as diagnostic tool of more complex climate models. I could not really follow the configurations of the intervention experiments and the impact experiments. It was not always clear which variables are prescribed and which respond. Actually, a sketch of the prototype model would have been very helpful in understanding the model and how processes are linked to each other. I can however understand that the author want to direct any questions to the first publication of the model prototype.
Rather than explicitly add an appendix or supplement, we have opted to incorporate further clarifying language throughout the text where appropriate. In our view, given the equations we have provided in the text, the table of baseline parameter specifications (Table 1), as well as the values tabulated in the original set-up discussed in Lintner et al. (2013), it should be possible for one to replicate our results.

• Prototype transect
The authors should explain what they mean with prototype transect, e.g. used P9446L19-20. I.e. what is the spatial resolution of the model/prototype C3598 and how are different moisture states realized.
We apologize for not making our usage of the term "prototype transect" explicit. We have thus added the following text (at the end of section 1).

> Here we are interested in examining the behavior across a range of hydroclimatic states; in what follows, we use the term "prototype transect" to refer to this range. This may be viewed as representing either a spatial sampling of states across a climatological gradient in soil moisture at a fixed point in time or a temporal sampling (as under the seasonal evolution) at a fixed point in space.

• Large scale irrigation setup
Although the scenario is of great interest, especially when focusing on land-atmosphere interaction, the outcome is not what I expected. What I would expect is that irrigation increases E and thus also P. Instead Fig. 9 shows no relevant change in E and a consistent decrease of P,

probably by the 2 mm/d of added irrigation. As I cannot fully understand the configuration and results I ask for a more detailed explanation. Possibly a water or energy budget at some value of w would help to understand what is going on.
We believe the incomplete understanding here may stem from the depiction of the quantities in Fig. 9 as functions of soil moisture. We have thus decided to include an additional panel in Fig. 9, depicting Ep, E, and P as functions of horizontal (drying) moisture advection (see Figure 9a). What is evident in this view is that E and P do increase with inclusion of irrigation, consistent with Reviewer 2's expectation. (We note that we had stated this in words, but this is clearly a case where a figure is far more illustrative!) It is when we plot the results with respect to soil moisture that the "expected" relationship does not appear to hold. In fact, that was our motivation for showing the soil moisture relative values in the first place.

• Minor comments:
Normalized form of the slope of the saturation vapor pressure curve γ
Please provide an inline equation of γ along with the explanations on page 9442.
The inline equation $\gamma = \frac{dq*}{dT}$, where $q*$ denotes saturation specific humidity, is now shown inline upon first usage of γ in Section 2.

• Derivation of analytic complementary relationship I do not fully understand the function f(T) in equations 11-13. Please provide a full form in the appendix.
We have defined this function in-line in the text. The point in expression equation (13) as we have done is to emphasize a part that looks like Priestley-Taylor and a "correction" factor that depends explicitly on precipitation and with a temperature dependence inherent in f(T). Since we prescribe T itself, this is just a constant over the transect. We hope this motivation is clear!

• Statement about temperature dependence of (Priestley and Taylor, 1972) on P9446L12ff
The Priestley-Taylor equation has a distinct temperature dependence through the slope of the saturation vapor pressure curve. With that I do not understand the sentence on P9446L12ff.
We were somewhat careless with the description of equation (13). Please see the revised discussion.

• P9449L5 Typo section 5
Thanks for the catch—corrected!

• P9449L27ff the description of the experiment (ii) is not clear to me
We have included some additional text to clarify this experiment.

• P9450 the first two intervention experiments are hardly discussed. What does it mean for the feedback and which variables are altered, e.g. when E/Ep is kept fixed? P9459L27ff unclear explanation, please explain
As above, we have included additional text to clarify these experiments.

• Table 1: there is a typo for the Priestley-Taylor coefficient
Thanks for catching the typo-correct!

• Axis labeling two panel figures is hardly readable, please increase font size

To do

• Fig 2 axis labels are not explained and not consistent with text
We have added the following to the Fig. 2 caption:
The values along the abscissa, E_{MI}, correspond to the ratio of actual to pan evaporation in Kahler and Brutsaert (2006) and are identical to soil moisture in the prototype.

• Fig.4 unclear unit of y-axis
Corrected.

The Budyko and complementary relationships in an idealized model of large-scale land-atmosphere coupling

Benjamin R. Lintner[1]*, Pierre Gentine[2], Kirsten L. Findell[3], and Guido D. Salvucci[4]

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[1]*Department of Environmental Sciences, Rutgers, The State University of New Jersey, New Brunswick, NJ, USA*

[2]*Department of Earth and Environmental Engineering and Earth Institute, Columbia University, New York, NY, USA*

[3]*Geophysical Fluid Dynamics Laboratory, Princeton, NJ, USA*

[4]*Department of Earth and Environment, Boston University, Boston, MA, USA*

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*Corresponding author: Benjamin R. Lintner, Department of Environmental Sciences, Rutgers, The State University of New Jersey, 14 College Farm Road, New Brunswick, NJ 08901-8551, USA

Email: lintner@envsci.rutgers.edu

Telephone: (848) 932-5731

Abstract.

~~Expressions corresponding to~~ Two well-known relationships in hydrology and hydrometeorology, the Budyko and complementary relationships, are examined ~~derived~~ within an idealized prototype representing the physics of large-scale land-atmosphere coupling developed in prior work. These relationships are shown to hold on long (climatologic) time scales because of the tight coupling that exists between precipitation, atmospheric radiation, moisture convergence and advection. The slope of the CR is shown to be dependent the Clausius-Clapeyron relationship between saturation specific humidity and temperature, with important implications for the continental hydrologic cycle in a warming climate, e.g., one consequence of this dependence is that the CR may be expected to become more asymmetric with warming, as higher values of the slope imply a larger change in potential evaporation for a given change in evapotranspiration. In addition, the transparent physics of the prototype permits diagnosis of the sensitivity of the Budyko and complementary relationships to various atmospheric and land surface processes. Here, the impacts of anthropogenic influences, including large-scale irrigation and global warming, are assessed.

1) Introduction

Observations of the annual terrestrial surface water balance demonstrate a tight and relatively simple functional dependence of evapotranspiration on the atmospheric water supply (precipitation) and demand (potential evaporation) at the surface (Budyko 1961, Porporato et al. 2004, Roderick and Farquahr 2011, Williams et al. 2012, Zanardo et al. 2012). Such observations have stimulated development of simplified analytical formulations of the annual water balance (Budyko, 1961; Lettau, 1961; Eagleson 1978a, 1978b, 1978c; Fu 1981; Milly 1994; Porporato et al. 2004; Harman et al. 2011; Sivapalan et al. 2011). Budyko (1961, 1974) developed arguably the most well-known approach for characterizing catchment-scale hydrologic balances on long (decadal and greater) timescales. By hypothesizing limitations on land surface evapotranspiration imposed by the availability of water and energy, Budyko introduced a relationship of the form:

$$\frac{E}{P} = B\left(\frac{E_p}{P}\right) \quad (1)$$

where E, P, and E_p are evapotranspiration, potential evapotranspiration, and precipitation, respectively. The ratio E_p/P is commonly known as the dryness or aridity index (ϕ), and hereafter we denote the ratio E/P as $B(\phi)$, i.e., the Budyko curve. Empirical forms of $B(\phi)$ have been obtained by fitting to observed E, P, and E_p with E typically estimated as the residual of precipitation and basin-scale streamflow (Budyko 1961, 1974). $B(\phi)$ appears to be rather stable across different regions and hydroclimatic environments (Potter et al. 2005; Yang et al. 2007; Gentine et al. 2012).

The last two decades have seen renewed interest in the Budyko curve (e.g., Milly 1994; Koster and Suarez 1999; Milly and Dunne 2002; Porporato et al. 2004; Potter et al. 2005; Donohue et al. 2007; Yang et al. 2008; Gerrits et al. 2009; Gentine et al. 2012; Istanbulluoglu et al. 2012; Williams et al. 2012). Many studies have employed simplified models of the surface and groundwater moisture response to precipitation forced by E_p ~~potential evapotranspiration~~ to investigate the robustness of the Budyko curve in different catchments. Milly (1994) and Porporato et al. (2004), in particular, investigated the response of the annual water balance to changes in the characteristics of potential evaporation and precipitation intensity and frequency yielding new insights on the sensitivity of the annual water balance to changes in surface energy and water forcing, all other factors, e.g., vegetation characteristics, soil, topography, remaining constant. However, apart from a few studies investigating the catchment co-evolution and adaptation of vegetation to the water and energy forcing (Troch et al. 2009; Sivapalan et al. 2011; Gentine et al. 2012), a direct explanation of the stability and widespread applicability of the Budyko relationship across a range of conditions remains elusive.

In the Budyko framework, as in most hydrologic models, E_p ~~potential evaporation~~ is prescribed as a forcing, which is thought to be independent from the surface. Nonetheless, E and E_p ~~and~~ have been ~~shown~~ hypothesized to be inversely related to one another (Bouchet 1963; Morton 1983; Brutsaert and Stricker 1979; Hobbins et al. 2001) on daily to annual time scales. In fact, the coupling between E and E_p provides the basis for a foundational relationship in hydrometeorology: the complementary relationship, first introduced by Bouchet (1963) and Morton (1983). Mathematically, the complementary relationship (CR) can be expressed as:

$$E_p + bE = (1 + b)E_{wet} \; . \quad (2)$$

In (2), E_p is potential evaporation and E_{wet} is the energy-limited, wet surface equilibrium evapotranspiration. Bouchet (1963) assumed a value of 1 for the scale factor b while and Salvucci (2009), hereafter PS09, report values in the range 3-6 based on numerical simulations of an evaporation pan in a drying environment. Since E_p cannot directly be measured, pan evaporation has often been used in lieu of potential evaporation using a pan correction factor (Bosman 1987, Roderick and Farquahr 2004; van Heerwaarden et al. 2010). Kahler and Brutsaert (2006) assumed that use of pan measurements in equation (2) may account for $b > 1$ because a pan transmits heat, resulting in warmer water inside the pan relative to a larger free water body (e.g., lake) under similar ambient conditions.

Although several studies (Zhang et al. 2004; Ramirez et al. 2005; Szilagyi and Jozsa 2009) have discussed possible links between the CR and the Budyko curve, to date a theoretical framework encompassing either or both of these relationships is still absent. Milly (1994) underscores the lack of physical understanding for why the Budyko curve deviates from its water- and energy-limited asymptotes, while Ramirez et al. (2005) suggest that, apart from heuristic arguments or under restrictive conditions, no general proof of the CR is available. Moreover, questions remain about the applicability and validity of the Budyko and complementary relationships across different spatial and temporal scales. Here we note that observational data confirm the CR holds on daily to annual time scales (Bouchet 1963; Kahler and Brutsaert 2006) and across local to regional spatial scales (Granger 1989, Szilagyi 2001; Crago and Crowley 2005; PS09 ~~Pettijohn and Salvucci 2009, hereafter PS09~~; van Heerwaarden et al. 2010), which is usually understood in terms of the diurnal-scale interactions of the boundary layer with the surface, although questions have been raised about some of the assumptions inherent in these approaches. Indeed, as PS09 note, some explanations for the CR have relied on contradictory assumptions. For example, the derivation of Szilagyi (2001) assumes that as E decreases, the surface temperature of the evaporation pan remains constant while the overlying near surface specific humidity decreases, increasing the vapor deficit and thus the evaporation rate over the pan. By contrast, in Granger (1989) surface temperature is assumed to increase while specific humidity remains constant, thus also increasing the humidity gradient and pan evaporation rate. Lhomme and Guilioni (2006) have questioned the physical validity and applicability of these assumptions.

In the present study, we make use of a semi-analytic, idealized prototype of large-scale land-atmosphere coupling developed in prior work (Lintner et al. 2013) to derive the Budyko and complementary relationships. Our approach differs from prior analyses in that: (i) it is implicitly large-scale and relevant on climatic timescales; and (ii) convergence, advection, precipitation and atmospheric radiation are treated implicitly rather than as exogenous forcing. The latter renders the atmospheric and surface moisture interactive and tightly coupled vertically but also horizontally—through the (nonlocal) effects of moisture advection and convergence. Several studies have pointed out that such tight coupling between radiation, larger-scale circulation and local surface energy budget is key to understanding locally-observed land-atmosphere interactions (Betts et al. 1996; 2003; Betts and Viterbo 2005; Betts 2007a; 2014). The analytic simplicity of the idealized prototype facilitates straightforward diagnosis of factors influencing the large-scale coupling, as highlighted in Lintner et al. (2013).

A key motivation for this study is consideration of how the continental hydrologic cycle, and more precisely how the behavior reflected in the Budyko curve and CR, may respond to anthropogenic influences. Indeed, the projected response of terrestrial hydrologic cycle to various climate change mechanisms in models remains subject to large uncertainties (Sherwood and Fu 2014). Emergent behaviors such as the Budyko and complementary relationships may provide useful constraints on such uncertainties. For example, Brutsaert and Parlange (1998) have suggested that the CR may explain the apparent paradox between observed downward trends in pan evapotranspiration over the late 20th century and anticipated increases in evaporation resulting from a more intense hydrologic cycle in a warming atmosphere.

The paper is organized as follows. After providing a brief review of the prototype assumptions and governing equations (Section 2), we analyze the generalized Budyko and complementary relationships within our prototype (Section 3) and consider the physical parameters and processes impacting these relationships (Section 4). Here we are interested in examining the behavior across a range of hydroclimatic states; in what follows, we use the term "prototype transect" to refer to this range. This may be viewed as representing either a spatial sampling of states across a climatological gradient in soil moisture at a fixed point in time or a temporal sampling (as under the seasonal evolution) at a fixed point in space. In Section 5, we examine how anthropogenic influences such as global warming and large-scale irrigation affect these relationships.

2) Overview of the idealized land-atmosphere coupling prototype

In prior work (Lintner et al. 2013), we developed a semi-analytic prototype for land-atmosphere coupling. This prototype describes the coupling at spatial scales for which both local (evapotranspiration) and nonlocal processes (horizontal moisture advection and convergence) may be important to the water cycle budget. We consider steady-state conditions, corresponding to the climatological state of the hydrologic cycle. Although the steady-state assumption clearly limits the applicability of our model in the presence of important time-dependent processes operating in the climate system, we again note that the CR has been observed to hold across a range of timescales. Similarly Budyko curves have been estimated from yearly mean observations (Budyko 1974, Gentine et al. 2012).

The atmospheric component of this prototype is based on vertically-integrated tropospheric temperature and moisture equations from the Quasi-equilibrium Tropical Circulation Model (QTCM; Neelin and Zeng 2000; Zeng et al. 2000), an intermediate level complexity model for the tropical atmosphere:

$$-Ms\nabla_H \cdot \mathbf{v} + P + R_{net} + H = 0 \tag{3}$$

$$Mq\nabla_H \cdot \mathbf{v} - P + E - \mathbf{v_q} \cdot \nabla_H q = 0 \tag{4}$$

where ∇_H is the horizontal gradient operator; R_{net} is the net column (top of the atmosphere minus surface) radiative heating; Ms and Mq are the dry static stability and moisture stratification and $\nabla_H \cdot \mathbf{v}$ is signed positive for low-level convergence; and $\mathbf{v_q}$ is the vertically-averaged horizontal wind vector weighted by the moisture vertical structure assumed in QTCM1. (Note that baseline values for parameters such as Ms and Mq are given in Lintner et al. 2013 and references therein. Table 1 summarizes parameter values most relevant to the present study.) The term P in (3) and (4) represents the net convective (condensational) heating and drying, respectively; the negative sign in (4) indicates that precipitation is a net sink of vertically-averaged tropospheric moisture. For the temperature equation (3), we have neglected horizontal temperature gradients following the weak temperature gradient assumption (Sobel and Bretherton 2000; Sobel et al. 2001). Note that all terms appearing in (3) and (4) are implicitly scaled to units of mm day^{-1} by absorbing constants such as (specific) heat capacity, latent heat of fusion, and column mass per unit area ${}^{\Delta p}/_{g}$, where Δp is the tropospheric pressure depth.

A steady balanced surface energy flux constraint, in which the annual-mean ground surface heat flux is neglected, reads:

$$R_{surf} - E - H = 0 \tag{5}$$

where the net surface radiative heating, R_{surf}, is signed positive downward.

In Lintner et al. (2013), we consider tropospheric temperature (T) as prescribed and solve the system of equations (3)-(5) for q, $\nabla_H \cdot \mathbf{v}$, and surface temperature Ts for prescribed large-scale advection. A closed-form, self-consistent solution can be obtained by invoking the steady-state soil moisture budget:

$$P - E - Q_{runoff} = 0 \tag{6}$$

where Q_{runoff} is the net runoff. For analytic simplicity, we assume a simple bucket model, with an evaporative efficiency, $\beta = \frac{E}{E_p}$, for which we assume a simple linear relationship $\beta = w$ (Porporato et al. 2001, 2004), where w is the dimensionless soil moisture (actual soil moisture normalized by a holding capacity). Q_{runoff} is represented as the precipitation rate times a power law of soil moisture, $Q_{runoff} = Pw^{\eta}$ (Kirchner 2009). The baseline power law scaling exponent is $\eta = 4$, the value used in Lintner et al. (2013). The suitability of invoking a single moisture storage variable to represent both basin scale evaporative efficiency and runoff has recently been demonstrated at nine watersheds containing Ameriflux eddy covariance measurements of evaporation and gauged streamflow (Tuttle and Salvucci, 2012).

For analytic simplicity, we consider linearized radiative and surface turbulent fluxes of the form:

$$\begin{aligned} H &= H_0 + \varepsilon_H(Ts - a_{1s}T) \\ E &= \beta[Ep_0 + \varepsilon_H(\gamma Ts - b_{1s}q) \\ R_x &= R_{x0} + \varepsilon_{Ts}^{Rx}Ts + \varepsilon_T^{Rx}T + \varepsilon_q^{Rx}q + c_xP \end{aligned} \tag{7}$$

Quantities with subscript "0" denote the values about which the fluxes are linearized, with coefficients ε representing the linear sensitivity of fluxes to T, q, and Ts. The scale factors a_{1s} and b_{1s} relate vertically-averaged temperature and moisture to near surface values appropriate for computation of surface bulk turbulent fluxes. The coefficient $\gamma = \frac{dq*}{dT}$ is the slope of the saturation specific humidity (q^*) with respect to temperature, as defined via the Clausius-Clapeyron equation. Note that in our usage, q (or q^*) is implicitly scaled to units of temperature (K) via absorption of the psychrometric constant; thus γ is dimensionless. The radiative fluxes are calculated at the top-of-the-atmosphere and the surface ($x = toa$ and $x = surf$), and the coefficients c_x are cloud-radiative forcing sensitivities, with cloud-radiative fluxes assumed to be linearly proportional to the precipitation rate. Precipitation (convective heating in equation (3) or convective drying in equation (4)) is formulated in terms of a Betts and Miller (1986)-type relaxation scheme:

$$P = \max\left[\varepsilon_c\big(q - q_c(T)\big), 0\right]. \quad (8)$$

Here, $q_c(T)$ is a temperature-dependent moisture threshold and ε_c is the convective adjustment rate coefficient (inversely related to the timescale for convective adjustment τ_c).

3) Overview of the baseline relationships

3a) Complementary relationship

Figure 1a illustrates the functional relationships between soil moisture and E, P, and E_p of the prototype. The general response of E and E_p with increasing soil moisture, namely E_p decreasing and E (generally) increasing, is consistent with the CR (2). E_p relates to w through the deficit between ambient and saturation specific humidity at the surface (Lintner et al. 2013): the deficit decreases with increasing soil moisture since q increases and Ts decreases. Figure 2 depicts the results of the prototype against E and E_p observations from Little Washita River Basin near Chickasha, Oklahoma (see Kahler and Brutsaert 2006 for a full description of the dataset). Here

E_p was obtained directly from pan evaporation measurements with a correction factor and E was measured using the Bowen ratio (EBBR) technique. E and E_p are presented in dimensionless units by dividing by the Priestly and Taylor (1972) evaporation, i.e., $E_{wet} = \alpha \frac{\gamma}{1+\gamma} R_{surf}$, where α is a correction factor with an estimated value of ~1.26. The CR from the prototype shows a qualitative, and arguably even quantitative, correspondence with both observational datasets. Of course, we should point out that the prototype was not explicitly ~~configured~~ tuned to represent the hydroclimate of these locations, and as such, any quantitative correspondence may be coincidental. Moreover, the scatter inherent in the observations would permit a range of plausible complementary relationships.

A slight decrease in E is observed for $w \gtrsim 0.7$ (see the gray curve in Figure 1a). To our knowledge such a decrease has not been previously investigated, even though it appears in *in situ* measurements (Kahler and Brutsaert 2006; PS09), as evident in Figure 2. The decrease ~~in~~ of both E and E_p arises from the monotonic decrease in E_p ($\frac{\partial E_p}{\partial w} < 0$) at large soil moisture, as pointed out by Lintner et al. (2013). Indeed, in the prototype, increasing soil moisture is a consequence of increasing precipitation, with the latter progressively balanced by higher moisture convergence (c.f., Figure 3 of Lintner et al. 2013). Increasing moisture convergence is associated with increasing (low-level) humidity, which reduces the surface vapor pressure deficit, ~~which ultimately reduces~~ thereby reducing ~~the potential evaporation~~ E_p. Overall the moisture balance at large soil moisture values implies a greater role for nonlocal processes. On the other hand, at very low soil moisture values, mean P is mostly balanced by E, resulting in a tight link among E, E_p, and w. This explains the success of local coupled land-boundary layer models (Bouchet 1963, van Heerwarden et al. 2010) in representing the drier regime of the CR. ~~We also note that at low soil moisture values, the precipitation essentially mimics the E response~~

~~and thus a complementary relationship also exists between precipitation and potential evaporation.~~

In prior studies of the CR, a quantity E_{wet} is introduced to denote the point of convergence of E and E_p under unlimited soil moisture (saturated surface) conditions i.e. at high soil moisture values (see equation 2). Conventionally, E_{wet} is assumed to represent equilibrium E from a saturated surface when advection is minimal and is usually computed empirically following Priestly and Taylor (1972). While the prototype E_p does indeed converge toward E as soil moisture increases, there is in fact no unique value of E_{wet} because of the decline E at high soil moisture.

As PS09 further note, plotting potential evapotranspiration and evapotranspiration against soil moisture (or a similar variable) may mask the linear nature of the CR. Thus, Figure 1b depicts E_p directly as a function of E. In this case, the approximate linearity implied by equation (1) is found to hold over a large range of E and E_p values. A linear regressive best fit to the E_p vs. E relationship for $E < 6$ mm day^{-1} yields a slope, i.e., the parameter b in equation (1), of magnitude ~3.8. This value of b is quantitatively consistent with the estimates of Kahler and Brutsaert (2006), Szilagyi (2007), and PS09, and unlike the original treatment of Bouchet (with $b = 1$), implies a strongly asymmetric CR. The implied value of b is close to γ, the dimensionless slope of the saturation specific humidity curve, which is 3.5 in the baseline configuration; in fact, as we show in the parameter sensitivity analyses in Section 4a), b varies predominantly with γ, which is consistent with the theoretical arguments presented in Granger (1989).

We can, in fact, derive an analytic expression for the CR for our land-atmosphere coupling prototype. We begin by subtracting γH from E_p and invoke the zero surface flux constraint to yield:

$$E_p - \gamma H = E_p + \gamma E - \gamma R_s \tag{9}$$

Then, using the bulk formulae expressions for E_p and H, the left-hand side of (9) can be expanded as

$$E_p - \gamma H = \epsilon_H(a_{1s}\gamma T - b_{1s}q) \tag{10}$$

Since precipitation rate $P = \epsilon_c\big(q - q_c(T)\big)$, q can be eliminated in favor of P, which upon rearranging the terms gives:

$$E_p + \gamma E = \gamma R_s - \frac{b_{1s}\epsilon_H}{\epsilon_c}P + f(T) \tag{11}$$

In (11), $f(T) = \epsilon_H(1+\gamma)^{-1}[a_{1s}\gamma T - b_{1s}q_c(T)]$ is just a function of the (prescribed) tropospheric temperature, and is thus constant ~~over the transect between dry and humid conditions~~ over the prototype transect.

Comparing (11) to (2), we find:

$$b = \gamma \tag{12}$$

and

$$E_{wet} = (1+\gamma)^{-1}[\gamma R_s + f(T) - \frac{b_{1s}\epsilon_H}{\epsilon_c}P]. \tag{13}$$

As discussed above, in prior work, E_{wet} was defined using the relationship of Priestley and Taylor (1972). The ~~first term of~~ term $(1+\gamma)^{-1}\gamma R_s$ in equation (13) corresponds to the Priestley-Taylor formulation but with a coefficient of 1 in lieu of α. It is worth noting ~~of~~ that Kahler and Brutsaert's *in situ* observations imply a Priestley-Taylor coefficient in the range of 0.89 to 1.13, which is in line with the value of 1 for the coefficient of E_{wet} suggested by (13). The ~~second~~

~~term in~~ remaining terms in E_{wet} are not explicitly represented in the Priestley-Taylor relationship E_{wet}, although this dependence may be implicitly captured through variation of the Priestley-Taylor coefficient. However, we note that the dependence of equation (13) on precipitation implies a negative feedback of P on the E_{wet}, since higher tropospheric moisture is associated with higher precipitation, thereby decreasing vapor pressure deficit at the surface and suppressing E_{wet}. This is consistent with the decrease to E at high soil moisture evident in Figure 1a.

It is obvious that E_{wet} as defined by (13) is not constant across the prototype transect, as it depends on surface radiative heating and precipitation. (In a more general model, variations in tropospheric temperature across the transect would also impact the value of E_{wet}.) Again, we point out that the Priestley-Taylor relationship only shows an explicit dependence on radiation (and the Clausius-Clapeyron slope). In addition to the negative feedback of precipitation on E_{wet} through the vapor deficit, R_{surf} itself also decreases as P increases, owing to the negative cloud-radiative forcing associated with deep convective clouds (see Lintner et al. 2013). Related to the non-constancy of E_{wet}, we also note that value of b differs slightly from the value inferred from directly fitting to the linear portion of the E_p vs. E curve in Figure 1b.

3b) Budyko curve

Within our prototype, the steady-state soil moisture equation (6) can be recast as:

$$B(\phi) = \left[1 - Q_{runoff}/P\right] \tag{14}$$

For the simple case of a land surface bucket model with a runoff power law scaling exponent of $\eta = 2$, and noting that soil moisture can be expressed as $w = \beta(w) = \frac{E}{E_p} = \frac{B(\phi)}{\phi}$, equation (14) reduces to a quadratic equation in $B(\phi)$, with an analytic solution in terms of ϕ expressed as:

$$B(\phi) = \frac{\phi^2}{2}\left(\sqrt{1 + \frac{4}{\phi^2}} - 1\right) \quad (15)$$

Figure 3 illustrates the Budyko curves for the baseline configuration of the prototype ($\eta = 4$) and the analytic solution for $\eta = 2$, with Budyko's well-known empirical formulation $B(\phi) = \sqrt{\phi \tanh(\phi^{-1})(1 - e^{-\phi})}$ for comparison. Also depicted are the energy- and water-limited asymptotes. For the baseline configuration, $B(\phi)$ at intermediate values of aridity index lies above the empirical Budyko fit, while the $\eta = 2$ curve lies below. Variation in the shape of $B(\phi)$ with increasing values of the power law scaling exponent η is consistent with decreasing runoff for a given value of precipitation, which in turn necessitates shifting the surface water balance to favor E over Q_{runoff}. We point out that equation (15) possesses limiting behavior consistent with empirically-derived estimates in prior studies (Budyko 1961, Fu 1981): thus, $B(\phi) \to 0$ as $\phi \to 0$ with a linear asymptote $B(\phi) \sim \phi$, while and $B(\phi) \to 1$ as $\phi \to \infty$ with an asymptote $B(\phi) \sim 1 - \phi^{-1}$. The limiting behavior of $B(\phi)$ as $\phi \to 0$ further implies that $E \to E_p$, which in turn necessitates $E_p \to 0$ in this limit. In other words, the decline in E at high soil moisture noted in Section 3a is also consistent with the Budyko curve.

4) Parameter and process sensitivity

4a) Parameter sensitivity

We now explore how the CR depends on parameter values assumed in the prototype. In particular, we focus on how the implied slope of E_p vs. E (e.g., the slope of the dashed black line in Figure 1b) depends on a subset of four prototype parameters: the dimensionless slope of the saturation specific humidity curve, γ; the surface drag coefficient, which is embedded in the turbulent flux scaling coefficient ε_H; the surface cloud longwave forcing, c_{surf}; and the convective adjustment timescale, τ_c. Our consideration of γ ~~as a parameter~~ is motivated by its control on the Bowen ratio or, similarly, the evaporative fraction (Gentine et al. 2011), and thus effectively the relative variation of q compared to Ts. We note that with global warming the saturation specific humidity is expected to increase and thereby modify precipitation and the hydrologic cycle (Held and Soden 2006). The surface drag coefficient depends on surface roughness, which could account for some of the heterogeneity in the observed slope of E_p vs. E. The remaining two parameters, c_{surf} and τ_c, are associated with two of the more uncertain aspects of current generation climate models, namely the effect of clouds on radiation and the parameterization of deep convection, respectively. Although the representation of clouds and convective processes are grossly simplified in the prototype, we can view the parameter sensitivity to c_{surf} and τ_c as a guide for anticipating how uncertainty in analogues to these parameters contained in more complex climate models may be expected to influence the CR evident in these models.

Figure 4 illustrates the percentage variation of the CR slope relative to its baseline value (~3.8) as functions of percentage variations in each of the four sensitivity parameters. Each of the latter is varied uniformly over a range of ±50% of the baseline values indicated in Table 1. It is immediately clear that the slope of the CR varies in a 1:1 manner with γ. On the other hand, for the remaining three parameters, the percent change in the CR slope is typically an order of magnitude smaller. We point out the nonlinearity associated with changing the surface drag coefficient, as decreasing surface drag produces a proportionately larger reduction in the slope of the CR compared to increasing surface drag by the same amount.

Since γ is just the slope of the Clausius-Clapeyron relationship, it has a quasi-exponential dependence on temperature and thus may be expected to vary sharply across the range of observed terrestrial temperature conditions. One consequence of this dependence is that the CR may be expected to become more asymmetric with warming, as higher values of the slope imply a larger change in E_p for a given change in E. In turn, this may have implications for the strength of the coupling between the land surface and atmosphere in a warming climate. For example, recent work by Dirmeyer et al. (2012, 2013, 2014) points to increases in metrics of land-atmosphere coupling strength in a warming climate. In section 4 5, we further explore the response of our prototype to a global warming scenario.

In contrast to the complementary relationship, the Budyko curve exhibits no apparent change in shape for these parameter variations. This shape invariance is unsurprising given that the Budyko curve is only a function of the aridity index ϕ, as can be seen directly in the analytic solution for the analytic solution for $\eta = 2$ (equation 15) or by substituting the expression for runoff in equation (14). Thus, while E, E_p, and P vary in response to changing prototype parameters, their ratios are constrained to lie along a fixed Budyko curve. Yang et al. (2009) suggest such shape invariance is characteristic of the Budyko curve response to what they broadly term "climate conditions," as they note climate forcing at a particular location simply moves the system from one point along its characteristic Budyko curve to another. By contrast, Yang et al. (2009) show how different locations fall onto distinct Budyko curves as a result of land surface or landscape properties such as soil, vegetation cover, rooting depth, etc.

4b) Process intervention experiments

Apart from considering the sensitivity of the CR relationship (or Budyko curves) to parameter values, we can also assess how the prototype solutions respond to altering a particular process or term in the governing equations. For the first such intervention-type experiment, we alter the evapotranspiration (E-intervention experiments) by prescribing as constant either i) β or ii) E_p. E-intervention experiment i) is analogous to the methodology adopted in the Global Land Atmosphere Coupling Experiment (GLACE)-type studies for comparing simulations with and without interactive soil moisture (Koster et al. 2004, 2006, Seneviratne et al. 2006). E-intervention experiment ii) is similar to the approach Lintner et al. (2013) used to sever the feedback of near surface climate onto E_p, which here is mediated principally through suppression of the dependence of potential evapotranspiration on "atmospheric drying power", since the variation of radiative forcing across the prototype transect in its baseline configuration is weak (see discussion below).

Rather than present the complementary relationship for the E-intervention experiments (since the CR necessarily breaks down in either case), we instead show Ts and q as functions of w (Figures 5a and 5b, respectively). For E-intervention experiment i) with β prescribed, the variation in Ts across soil moisture conditions (gray curve) is considerably reduced: while the difference in the baseline Ts (black curve) between the driest and wettest conditions is roughly 5K, it is under 0.5K with β prescribed. Similarly, the range of variation in specific humidity (here scaled to its surface value) across soil moisture states is attenuated relative to the baseline, although it is less pronounced than for surface temperature. Qualitatively opposite behavior is seen under E-intervention experiment ii) with E_p prescribed, as the variations of both Ts and q across the range w are increased relative to their baseline values. Note that at low soil moisture, the behavior of the baseline case more closely resembles E_p prescribed experiments, while at high soil moisture, it is more similar to the β prescribed experiment.

We can further assess how the CR changes with intervention in either surface sensible heat H or surface radiative heating R_s, by prescribing either of these fluxes to a constant value. Under

R_s-intervention (Figure 6a), there is little net change in the CR relative to the baseline over most of the range of soil moisture; however, at high w, both E_p and E are slightly increased. The increase in E_p arises through a slight elevation of surface radiation heating above the baseline state, since the negative effect of cloud shortwave forcing, i.e., more convective clouds leading to less surface shortwave heating, is absent. This in turn feeds back onto precipitation, which is slightly enhanced. We further mention a competing effect, namely increased surface longwave forcing (and hence warming) with increased water vapor and convective cloudiness. For the parameter values chosen, this effect loses out to the shortwave forcing. On the other hand, uncertainty in these parameter values, particularly the cloud forcing, could alter the balance of these two effects.

Under H-intervention (Figure 6b), the CR is dramatically altered, as E_p drops off more rapidly with increasing soil moisture, while E rises faster at low soil moisture and then flattens off. The quantitative details of the change in shape of E_p and E depend on the value of sensible heat flux prescribed. Plotting E_p vs E (not shown) yields a best fit linear regressive slope of ~28, consistent with the very asymmetric nature of the CR when the variation in H across the transect is suppressed.

5) Impacts of global warming and large-scale irrigation

5a) Global warming

How the hydrologic cycle responds to global warming is clearly of great significance to projecting climate change impacts (Allen and Ingram 2002, Milly et al. 2005). At present, our understanding of the hydrologic cycle response to warming is guided by some theoretical constraints, e.g., the Clausius-Clapeyron relationship promotes enhanced tropospheric moistening, although model projections show considerable spread in the regional signatures of hydrologic cycle change (Held and Soden 2006; Neelin et al. 2013). Assessing global warming impacts on the terrestrial hydrologic cycle is complicated by changes in land use such as deforestation and agricultural conversion and coupling to vegetation (Lee et al. 2011).

Over the latter half of the 20th century, several studies have reported widespread decreases in pan evaporation (Lawrimore and Peterson 2000; Hobbins et al. 2004; Roderick and Farquhar 2004; Shen et al. 2009), which can be related to E_p. Several hypotheses have been proposed to explain the decreasing trend in pan evaporation, including increasing precipitation reducing the vapor pressure deficit of the lower atmosphere, global dimming reducing shortwave radiative heating at the surface, and stilling of surface winds reducing the exchange coefficient. Van Heerwaarden et al. (2010) conducted an extensive set of sensitivity tests for each of these effects on the CR using a conceptual model of the diurnal terrestrial boundary layer. They concluded that "except over wet soils, the actual evapotranspiration is more sensitive to changes in soil moisture than to changes in short wave radiation so that global evaporation should have increased. Nevertheless, Wild et al. (2004) speculate that in the latter half of the 20th century, increased moisture transport from the oceans enhanced precipitation over land, but suppressed the evaporation—opposite to [their] expectations".

Figure 7 depicts the effect on prototype hydroclimate of imposing a 2K warming of the prescribed column-mean temperature. In this figure, differences between the 2K warming configuration and the baseline are plotted against baseline values of w; also shown is the difference in soil moisture between the 2K warming and baseline scenarios. Across the range of baseline soil moisture conditions, the imposed warming decreases E_p (black), since q itself increases. While E (gray) increases with warming at w, it decreases for $w > 0.5$, which is consistent with the results of van Heerwaarden et al. (2010). In addition, P (blue) increases under warming over the entire range of precipitation values (similar to Wild et al. 2004), albeit with a local minimum at intermediate soil moisture values.

The opposing changes of E_p and E with warming at low soil moisture are consistent with expectations from the CR, as an increase in one corresponds to a decrease in the other. Of course, increasing the temperature increases the value of γ, which means the slope of the CR increases between the baseline and 2K warming scenarios. On the other hand, since E_{wet} decreases with tropospheric warming, both E_p and E follow the behavior of E_{wet} and therefore decrease.

While the values of hydroclimatic variables may change substantially between the baseline and 2K warming scenarios, the Budyko curve is unaltered, as discussed in Section 4a. However, as Figure 8 illustrates, forcing conditions, i.e., the value drying advection, identified with specific points along the Budyko curve in the baseline scenario are shifted to lower values of the aridity index ϕ in the 2K warming scenario, since E_p decreases while P increases. Based on these results, we note the potential utility of the Budyko curve in providing qualitative or even quantitative constraints on how terrestrial hydroclimate variables will respond to warming.

5b) Large-scale irrigation

Over many parts of the world, irrigation has been adopted to support agricultural production. Over India, for example, irrigation is now sufficiently extensive that large-scale alterations of hydroclimate may be occurring (Cook et al. 2010, Guimberteau et al. 2011). With respect to potential irrigation-induced changes in hydroclimate, Ozdogan et al. (2006) employed a mesoscale climate model and field data to demonstrate that large-scale irrigation in southeastern Turkey has impacted evaporation and potential evaporation in a complementary manner. They found a variety of interactions responsible for the trends, including increased atmospheric stability, decreased vapor pressure deficit, and, interestingly, a strong decrease in wind speed. Han et al. (2014) point out that while trends in E_p have often been invoked to estimate possible trends in E, how irrigation may impact E_p has typically been neglected in assessment and interpretation of E_p trends. It is thus worth briefly investigating how large-scale irrigation modulates the Budyko and complementary relationships within the framework of the prototype analyzed here. To do this, we consider the addition of an irrigation source, I, to the soil moisture balance equation (6), which then becomes $P + I = E + Q_{runoff}$.

To see the impact of irrigation on the prototype hydroclimate, Figure 9 depicts E, E_p, and P ~~as functions of soil moisture~~ in the baseline and $I = 2$ mm day^{-1} configurations. Here we have plotted these quantities with respect to both horizontal dry advection (Figure 9a) and soil moisture (Figure 9b). Given the direct moistening of the atmosphere by irrigation, the transition between nonprecipitating and precipitating conditions in the presence of irrigation occurs at a significantly larger value of drying advection in the large-scale irrigation scenario, as shown in Figure 9a. Across the range of moisture advection values, both E and P are enhanced in the presence of large-scale irrigation, as expected, and E_p decreases, in line with complementarity. Additionally, at low to intermediate precipitation, E exceeds P. Thus, whatever soil moisture is not locally recycled as precipitation would instead be transported downwind, as suggested by some studies on observed irrigation (e.g., DeAngelis et al. 2010). Viewed with respect to soil moisture (Figure 9b), the inclusion of large-scale irrigation is seen to induce a slight increase in E for a given value of w. On the other hand, since the value of drying advection is larger at a given w in the presence of irrigation, E_p itself is larger. Directly relating E_p to E indicates effectively no change in the slope of the CR, though the intercept is increased when irrigation is applied (not shown). Precipitation at a given value of w is lowered in the irrigated scenario.

As a consequence of the changes in E and P, the Budyko curve for irrigated conditions (Figure 10, dotted line) is shifted above its baseline: in fact, the irrigated Budyko curve extends above 1 for $\phi > 1$, as water limitation is effectively alleviated with the imposed irrigation water source. When E is replaced by the residual $E^* = E - I$, the resulting Budyko-like curve (stars) drops below the baseline.

Of course, we should point out that by imposing irrigation in the prototype with tropospheric temperature prescribed, we are neglecting a potentially important cooling of the lowermost atmosphere, not to mention that our prototype does not account for changes in convective initiation or triggering that may occur, e.g., through changes to atmospheric stability. Moreover, we do not take into account vegetation control on E since we only represent soil moisture dependence of evapotranspiration through a bucket model. Thus, the irrigation impacts described here merely reflect the direct effect of added moisture to the atmosphere.

6. Summary and conclusions

In this study, we use an idealized prototype incorporating the key physics of large-scale land-atmosphere coupling to derive analytic expressions for the well-known Budyko and complementary relationships. Our approach differs from previous analytic approaches in that precipitation and moisture convergence are treated implicitly rather than applied as an external forcing. The analytic solutions permit straightforward diagnosis of the sensitivity of the Budyko and complementary relationships to atmospheric and land surface parameters. In particular, the slope of the CR is shown to be mostly dependent on the temperature with important implications for the continental hydrologic cycle with a warming climate. One consequence of this dependence is that the CR may be expected to become more asymmetric with warming, as higher values of the slope imply a larger change in potential evaporation for a given change in evapotranspiration. On the other hand the Budyko curve is very stable to many parameterization of the model parameters or global temperature. It is thus expected that the Budyko curve should remain relatively stable under a warming climate. Other causes of anthropogenic changes such as large-scale irrigation are however shown to strongly impact the Budyko curve with little impact on the CR.

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Table and Figure Captions

Table 1: Parameter definitions and values in the baseline land-atmosphere coupling prototype.

Figure 1: Complementary relationship in the baseline configuration. a) Potential evapotranspiration (E_p; black), evapotranspiration (E; gray), and precipitation (P; blue) as functions of soil moisture (W). b) E_p vs. E (black) and the 1:1 line (gray). Also shown is the best fit linear regression of the E_p to E relationship (squares).

Figure 2: Baseline complementary relationship compared to Kahler and Brutsaert's observational data from the Little Washita River basin in Oklahoma, USA (c.f., Figure 5 in Kahler and Brutsaert, 2006). Symbols shown correspond to two different normalizations of the observations and gray lines to best fits through these points. The values along the abscissa, E_{MI}, correspond to the ratio of actual to pan evaporation in Kahler and Brutsaert (2006) and are identical to soil moisture in the prototype. Prototype E_p and E (red and purple curves, respective) have been normalized with respect to the value of E_{wet} corresponding to the maximum value of E along the transect.

Figure 3: Prototype Budyko curves for the baseline prototype, i.e., $\eta = 4$ in the formulation of runoff (thick black), for equation (15) for $\eta = 2$ (gray), and Budyko's empirical formula (squares). The dashed lines are the energy- and water-limited asymptotes.

Figure 4: Parameter sensitivity of the complementary relationship slope. Results shown are for varying: the slope of the saturation specific humidity with respect to temperature (no symbols), surface drag coefficient (circles), surface cloud radiative forcing (stars), and convective adjustment timescale (squares).

Figure 5: Comparison of prototype (a) surface temperature *Ts* and (b) surface air humidity *qa* for the baseline (black), fixed β (gray), and fixed Ep (blue) configurations of the prototype. Note that *qa* is converted to temperature units of K.

Figure 6: Impact on the complentary relationship from prescribing either (a) net surface radiative heating or (b) sensible heat flux. The curves depicted correspond to E_p (black), E (gray), and P (blue) for the baseline configuration (solid) and prescribed radiative or sensible heat fluxes (dashed).

Figure 7: Differences in E_p (black), E (gray), and P (blue) for a +2K warming relative to the baseline configuration as functions of soil moisture in the baseline configuration. Also shown is the difference in soil moisture (dashed black), which has been rescaled by a factor of 10.

Figure 8: Shift in selected points along the Budyko curve for the baseline (black symbols) and +2K warming (red) configuration. Pairs of like shaped symbols correspond to the same level of imposed drying advection forcing. The values shown in the inset are the percentage changes for each of the baseline and +2K warming pairs.

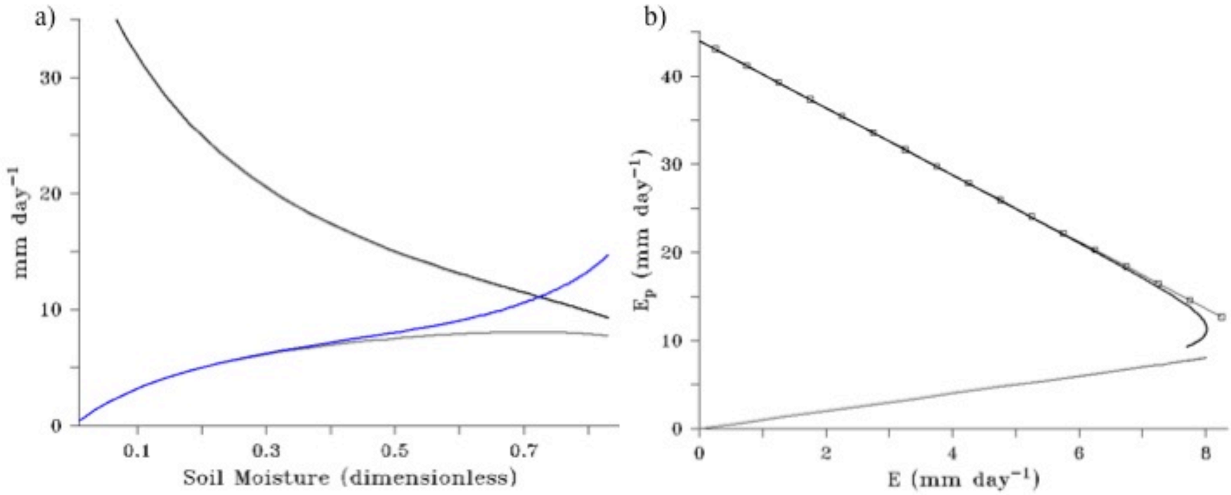
Figure 9: Large-scale irrigation impacts on prototype hydroclimate. a) E_p (black), E (gray), and P (blue) for the baseline (solid curves) and $I = 2$ mm day^{-1} large-scale irrigation scenario (dotted curves), plotted with respect to horizontal moisture advection, scaled to units of mm day^{-1}. Here,

horizontal advection values correspond to drying advection, which are associated with decreasing precipitation. b) As in a), but with soil moisture as the abscissa.

Figure 10: Comparison of Budyko curves for the baseline (solid) and I = 2 mm day^{-1} (dotted) configurations. Also shown is a Budyko-like curve in which E is replaced by $E^* = E - I$ (stars).

Table 1: Parameter definitions and values in the baseline land-atmosphere coupling prototype

Parameter	***Definition***	***Value***
a_{1s}	Weighting factor for surface temperature	0.30
α	Priestley-Taylor coefficient	1.26
b_{1s}	Weighting factor for surface moisture	1.15
b	Complementary relationship scale factor	--
c_{surf}	Surface cloud longwave forcing coefficient	0.18
γ	Dimensionless slope of Clausius-Clapeyron relationship	3.5
ε_H	Linearized surface turbulent flux scaling coefficient	42 mm day^{-1} K^{-1}
η	Runoff power law scaling exponent	4
τ_c	Convective adjustment timescale	2 hours



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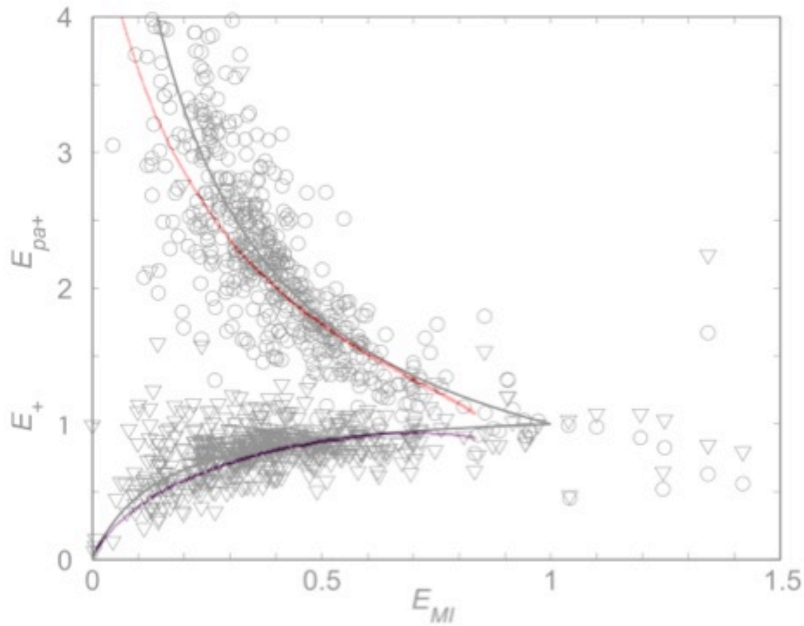

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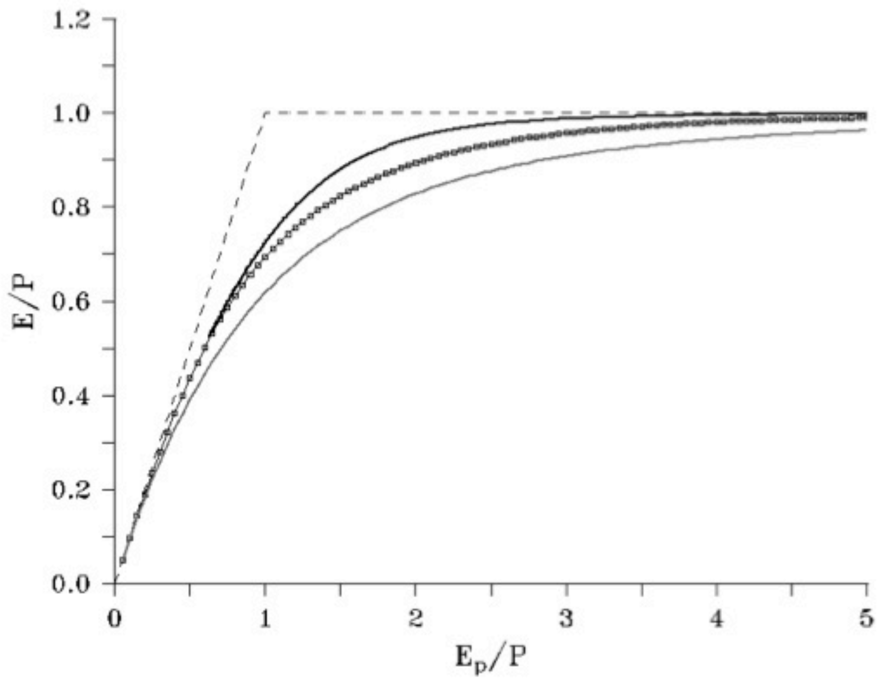

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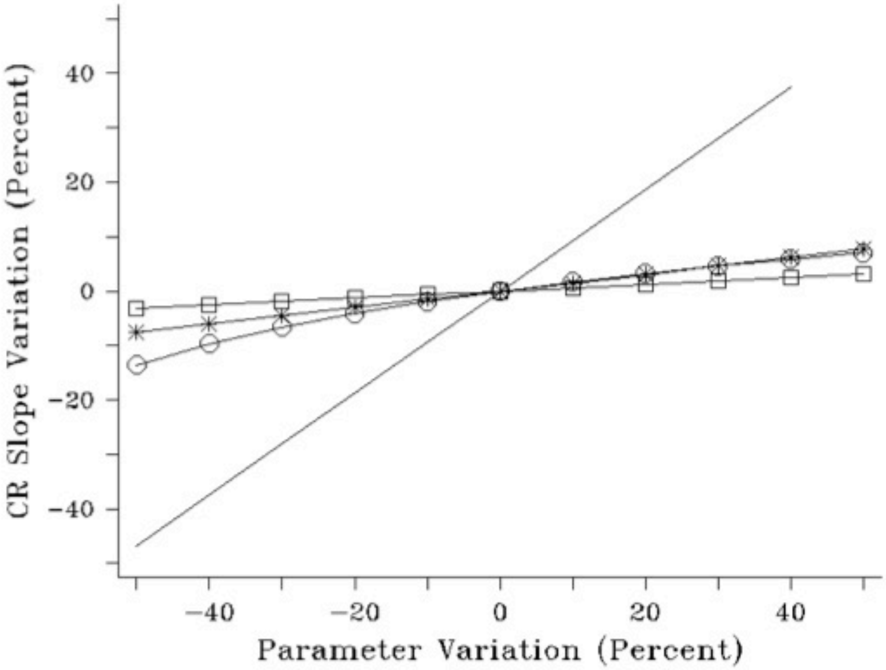

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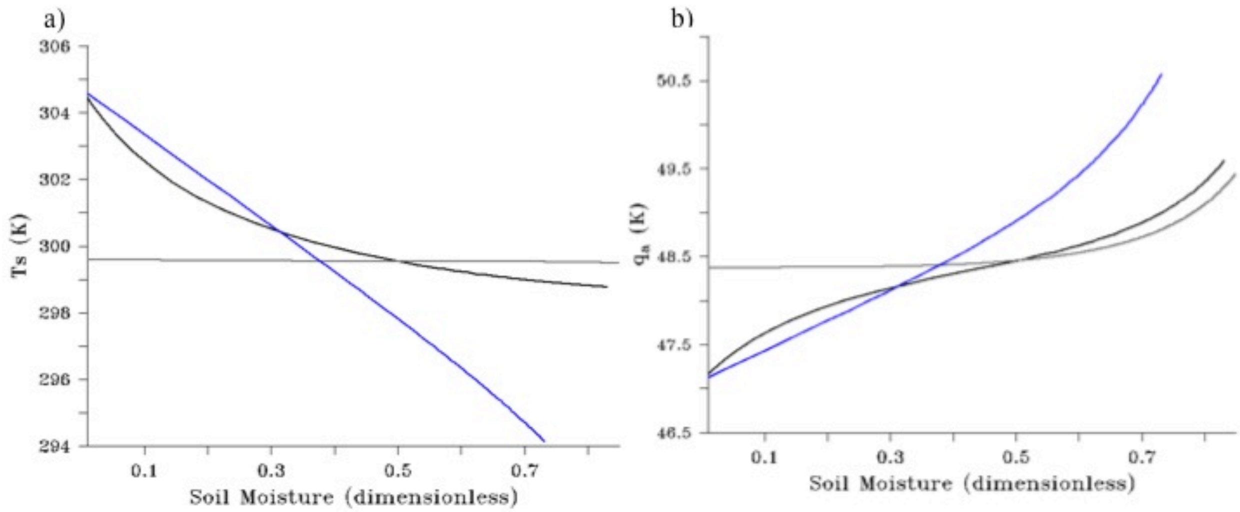

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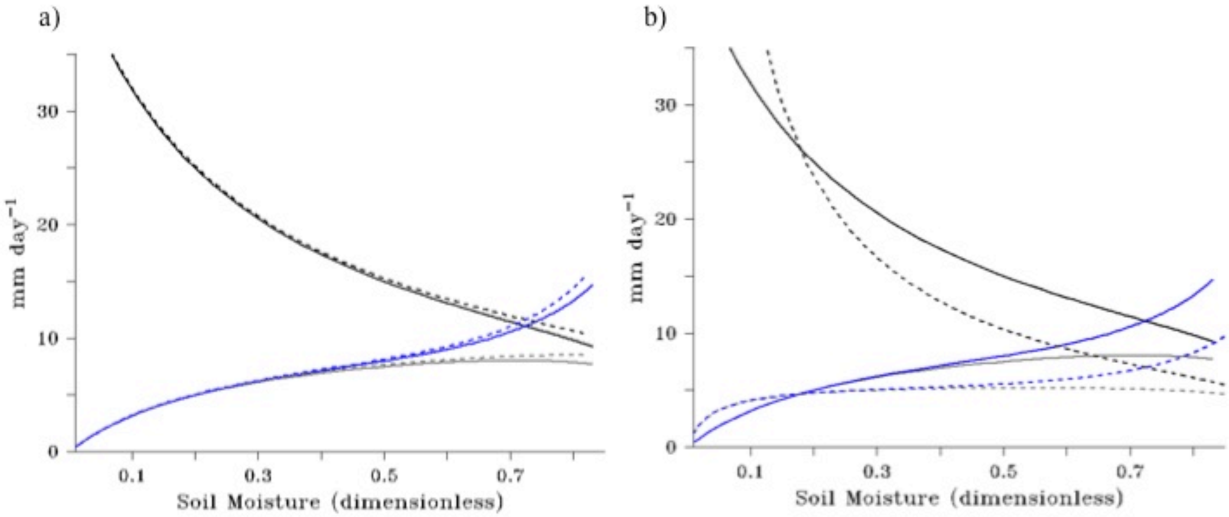

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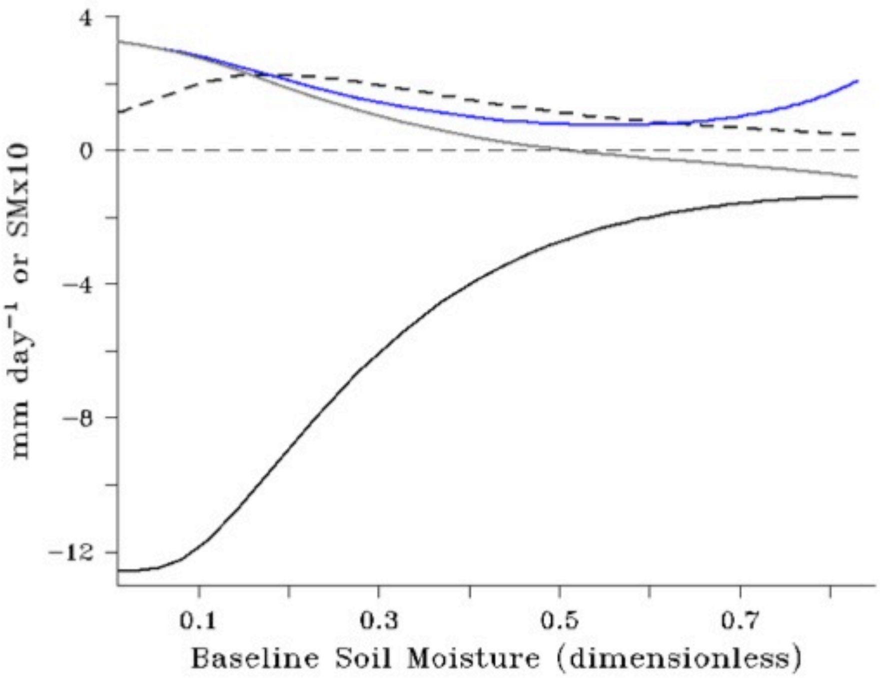

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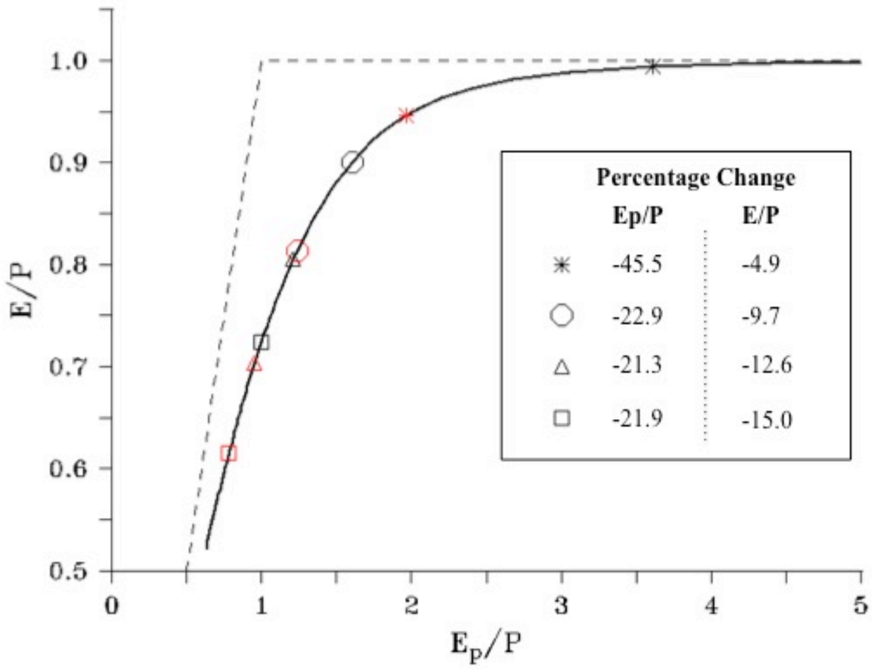

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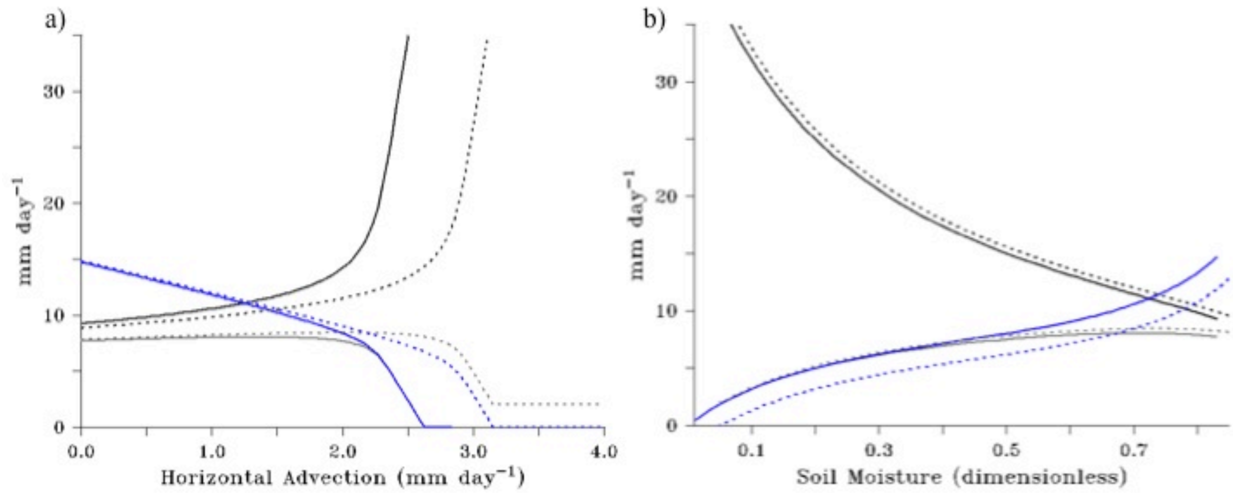

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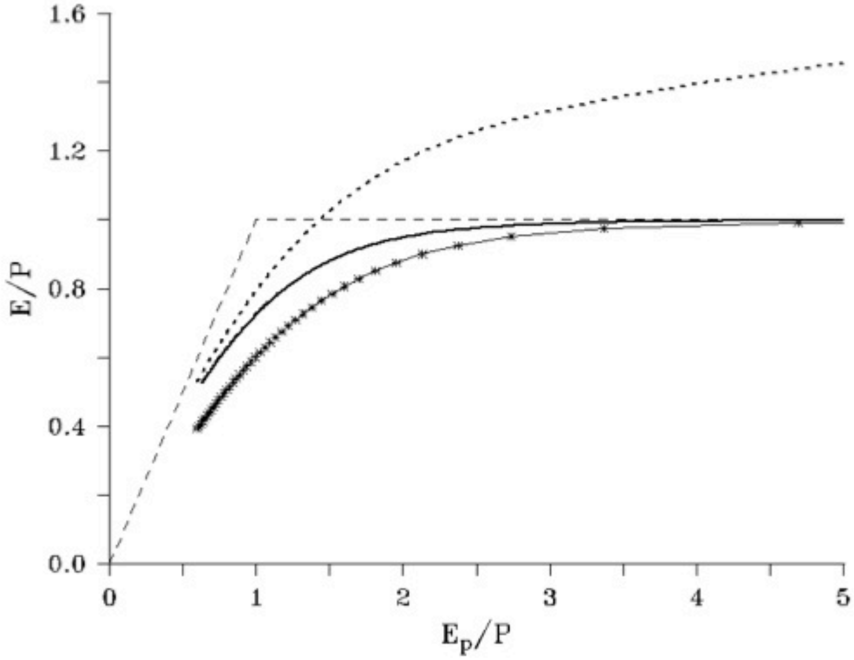

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