On the Hydrologic Adjustment of Climate-Model Projections: The Potential Pitfall of Potential Evapotranspiration

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ABSTRACT: Hydrologic models often are applied to adjust projections of hydroclimatic change that come from climate models. Such adjustment includes climate-bias correction, spatial refinement ("downscaling"), and consideration of the roles of hydrologic processes that were neglected in the climate model. Described herein is a quantitative analysis of the effects of hydrologic adjustment on the projections of runoff change associated with projected twenty-first-century climate change. In a case study including three climate models and 10 river basins in the contiguous United States, the authors find that relative (i.e., fractional or percentage) runoff change computed with hydrologic adjustment more often than not was less positive (or, equivalently, more negative) than what was projected by the climate models. The dominant contributor to this decrease in runoff was a ubiquitous change in runoff (median −11%) caused by the hydrologic model's apparent amplification of the climate-model-implied growth in potential evapotranspiration. Analysis suggests that the hydrologic model, on the basis of the empirical, temperature-based modified Jensen–Haise formula, calculates a change in potential evapotranspiration that is typically 3 times the change implied by the climate models, which explicitly track surface energy budgets. In comparison with the amplification of potential evapotranspiration, central tendencies of other contributions from hydrologic adjustment (spatial refinement, climate-bias

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adjustment, and process refinement) were relatively small. The authors' findings highlight the need for caution when projecting changes in potential evapotranspiration for use in hydrologic models or drought indices to evaluate climate-change impacts on water.

KEYWORDS: Hydrologic model; Climate change; Potential evapotranspiration

1. Introduction

Climate-change experiments with numerical climate models produce projections of changes in the water cycle. These projections include changes in fluxes (precipitation, evapotranspiration, and runoff) and storage (snowpack, soil water, and groundwater). The value of these projections for long-term water-resource planning and risk analysis is compromised by many factors, but here we focus on three problems: 1) the biases in the modeled historical climates, 2) the coarse resolution of the climate models, and 3) the imperfect representation of hydrologic processes in the climate models. The first problem is illustrated by the fact that a climate model could err locally, for example, by tens of percent in its computation of mean annual precipitation on a given watershed during the twentieth century; associated errors in hydrologic sensitivities might be considerable. The second problem implies that the models overlook finescale (e.g., <100 km) details in climate and hydrology, such as those associated with mountainous topography or a coastal environment. The third problem calls into question the ability of the models realistically to translate a given change in climate into an associated change in water flux or storage.

Most hydrologic modeling studies of climate-change impacts have adopted a common strategy to address the problems noted above. Under this strategy, problems 2 and 3 are treated by use of a hydrologic model that contains representations of the desired processes at the desired spatial scales. The hydrologic model is forced by the sum of historical climate (here also termed the "baseline" climate, for consistency with Hay et al. 2010) and climate-model-derived changes in climate (climate-model future versus climate-model baseline period), thereby treating problem 1. The climate change is prescribed on the basis of temporal differences (usually for air temperature) and/or ratios (usually for precipitation) of climate variables in the climate model.

The validity of the strategy outlined above relies on implicit assumptions that climate change is insensitive to 1) the baseline climate and 2) the hydrologic response. Together, these assumptions would justify the transfer of the climate-model climate changes (differences and/or ratios, available only at coarse scale) to the stand-alone hydrologic models (at finescale). These are fundamental assumptions, in need of scrutiny, but the assessment of their validity is beyond the modest scope of this paper.

Which set of variables should be used for the handoff of climate information from the climate model to a hydrologic model? Commonly, precipitation and near-surface air temperature are used, and the change in potential evapotranspiration rate for the hydrologic model is then computed from the change in temperature; potential evapotranspiration is a conceptual variable whose use simplifies the treatment of surface energy balances in hydrological models. The computation of change in potential evapotranspiration from change in temperature typically is

accomplished either on the basis of an empirical equation relating potential evapotranspiration to air temperature (e.g., the Thornthwaite method) or (either explicitly or implicitly) through the solution of a surface energy-balance equation (e.g., the Penman–Monteith equation). Even in the energy-balance approaches, the changes in some variables, such as shortwave and longwave radiation fluxes, wind speed, and atmospheric vapor pressure, commonly have been expressed in terms of temperature changes by means of empirical relations and/or physical arguments (Maurer et al. 2002). Empirical temperature-based relations for potential evapotranspiration are many and varied, as are empirical temperature-based relations for the inputs to the energy-balance methods. Some of these relations may be valid in a changing climate; others may not. Climate models directly compute a full surface energy balance, usually without recourse to an explicit definition of potential evapotranspiration, and do not rely on empirical relations among atmospheric variables, because all of the relevant variables are computed on the basis of dynamic interactions within the model.

Recently, the U.S. Geological Survey (USGS; Hay et al. 2010) computed sensitivities of 14 river basins in the United States to scenarios of future climate change. The study followed the hydrologic-adjustment strategy described above and used the Precipitation Runoff Modeling System (PRMS) to quantify hydrologic response. Potential evapotranspiration was specified according to a modified Jensen–Haise formulation (Jensen et al. 1970); that formulation is one of the multiple options for describing potential evapotranspiration within PRMS. The multiplicity of basins across a range of climates allows us to use the USGS study as a case study within which we can examine issues related to hydrologic adjustment of climate-change impacts.

The objectives of this paper are, within the context of the larger USGS study, 1) to estimate the extent to which stand-alone hydrologic-model water-balance changes associated with climate change differ from the water-balance changes in the climate models that are used to estimate the climate change; 2) to identify the relative importance of distinct contributors to those differences; 3) in particular, to assess the consistency of potential evapotranspiration changes between the climate model and the modified Jensen–Haise formulation; and 4) by example, to demonstrate a methodology that can be applied to similar assessments of other potential evapotranspiration formulations used in other hydrologic-adjustment studies.

2. Analytic strategy

The partitioning of precipitation p into evapotranspiration e and runoff r is controlled largely by the relative magnitudes of precipitation and potential evapotranspiration e_p and by physical characteristics of a given river basin (Milly 1994). A parsimonious expression of these controls is the generalized Turc–Pike relation (Choudhury 1999),

$$e = p - r = p\varphi_v(e_p/p), \quad \text{where} \quad \varphi_v(x) \equiv [1 + x^{-v}]^{-1/v}, \tag{1}$$

or simply

$$r = r(p, e_p, v), \tag{2}$$

in which the parameter v characterizes the tendency of the river basin to conserve water for evapotranspiration. Our strategy is to use this relation as a crude "model of all models" to approximate the behaviors of PRMS and the land representations in the climate models and, thereby, to facilitate the elucidation of their differing behaviors.

The generalized Turc–Pike equation captures the transition of evapotranspiration from energy limitation to water limitation over the range of climates from humid to arid (Figure 1). For any value of v, (1) dictates that mean actual evapotranspiration approaches an energy ($e \rightarrow e_p$) or water ($e \rightarrow p$) limitation asymptotically under very wet or dry climatic conditions, respectively. The parameter v controls the overall departure of basin behavior from the asymptotes and is partially indicative (but in a very nonlinear way) of the water-conserving tendency of the basin. A very deep swimming pool would have a large value of v; in contrast, an impermeable parking lot would have a small value. Additionally, positive temporal correlation of precipitation and potential evapotranspiration tend to increase the value of v and, thus, evapotranspiration.

We consider a climate change from some baseline state (p, e_p) to some other, say, projected future state ($f_p p, f_e e_p$), where the f coefficients are factors by which the climate variables change from baseline to future state. Throughout this analysis and generally consistent with the methodology of the USGS study, we focus on quasi-equilibrium climate change, but the analysis would be only slightly affected by consideration of transient climate change. Under the assumption that the basin characteristics do not change (v is a constant), we have a change in runoff δr given by

$$\delta r = r(f_p p, f_e e_p, v) - r(p, e_p, v). \tag{3}$$

So far, we have used (1) as a simple model of actual water balances. Now, we shall adopt it as a model of other water-balance models: that is, of PRMS (superscript H for hydrologic model) and of the water-balance descriptions embedded in climate models (superscript C for climate model). Thus,

$$\delta^H r = r(f_p^H p^H, f_e^H e_p^H, v^H) - r^H; \quad r^H \equiv r(p^H, e_p^H, v^H) \quad \text{and} \tag{4}$$

$$\delta^C r = r(f_p^C p^C, f_e^C e_p^C, v^C) - r^C; \quad r^C \equiv r(p^C, e_p^C, v^C). \tag{5}$$

The difference in runoff changes computed by the hydrologic model and the climate model can be computed directly from (4) and (5). For practical purposes, hypothetical climate-induced runoff differences often are expressed as relative changes (change divided by baseline value). In this study, we are interested in how hydrologic adjustment transforms the climate-model estimate of relative change in runoff to an adjusted estimate of relative change. The use of relative change can
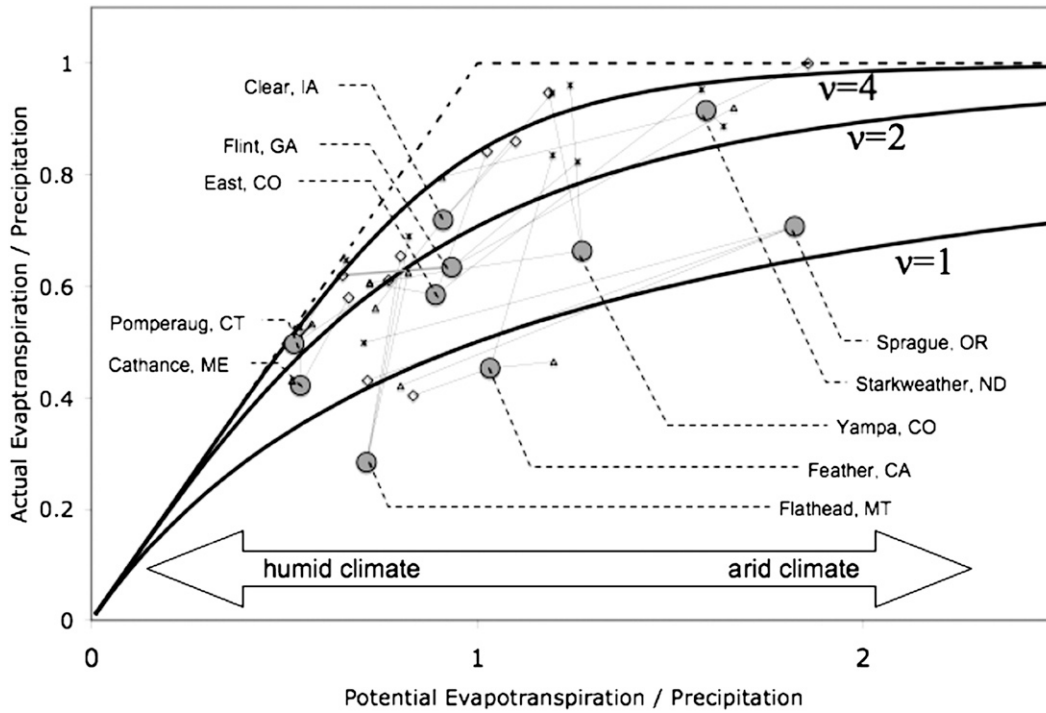
Figure 1. Ratio of actual evapotranspiration to precipitation as a function of ratio of potential evapotranspiration to precipitation; all fluxes are long-term means. Curves represent the generalized Turc–Pike relation, with parameter values (v) indicated; dashed line segments ending at the point (1,1) represent the asymptotes of the Turc–Pike relation. Large, filled circles indicate baseline data from PRMS (which was calibrated to observations); other symbols indicate baseline data from three climate models (CSIRO = diamond; INM = triangle; MIROC = asterisk). Same-basin data are connected by thin solid lines.

create confusion because the baseline runoff (which enters the denominator) differs between climate models and the hydrologic model. To minimize such confusion, we find it informative to separate the difference in relative change into a part related to the change in absolute runoff and a part related to the difference in denominators,

$$\frac{\delta^H r}{r^H} - \frac{\delta^C r}{r^C} = \frac{\delta^H r - \delta^C r}{r^H} + \left(\frac{1}{r^H} - \frac{1}{r^C}\right)\delta^C r. \tag{6}$$

Additionally, we can decompose the absolute change ($\delta^H r - \delta^C r$) into parts associated with differences in various input variables in (4) and (5). The decomposition can be done many ways. We choose to use a series of perturbations away from climate-model values of the input variables, leading to

$$\frac{\delta^H r}{r^H} - \frac{\delta^C r}{r^C} = [r(f_p^C p^H, f_e^C e_p^H, v^C) - r(f_p^C p^C, f_e^C e_p^C, v^C)$$
$$- r(p^H, e_p^H, v^C) + r(p^C, e_p^C, v^C)]/r^H$$
$$+ [r(f_p^H p^C, f_e^C e_p^C, v^C) - r(f_p^C p^C, f_e^C e_p^C, v^C)]/r^H$$
$$+ [r(f_p^C p^C, f_e^H e_p^C, v^C) - r(f_p^C p^C, f_e^C e_p^C, v^C)]/r^H$$
$$+ [r(f_p^C p^C, f_e^C e_p^C, v^H) - r(f_p^C p^C, f_e^C e_p^C, v^C)$$
$$- r(p^C, e_p^C, v^H) + r(p^C, e_p^C, v^C)]/r^H$$
$$+ \left(\frac{1}{r^H} - \frac{1}{r^C}\right)\delta^C r + \varepsilon, \tag{7}$$

where ε is a residual necessitated by nonlinearities. The first four of the six terms on the right side of (7) arise from differences between the climate models and the hydrologic model with respect to the baseline climate (p, e_p), the multiplier for change in precipitation (f_p), the multiplier for change in potential evapotranspiration (f_e), and the behavior of the basin (v), respectively.

A complication of the analytic framework presented above is that, as noted already, climate models generally do not employ the concept of potential evapotranspiration. However, hydrologic science provides guidance for estimating e_p from the physical variables that are represented in climate models. To enable application of (7) in this analysis, we evaluate climate-model e_p by use of the formula of Priestley and Taylor (Priestley and Taylor 1972),

$$e_p = \alpha[\Delta/(\Delta + \gamma)](R_n - G)/L, \tag{8}$$

where α is an empirical constant; Δ is the slope of the saturation vapor-pressure curve (a function of air temperature); γ is the psychrometric constant (proportional to atmospheric pressure); R_n is the surface net radiation; G is the heat flux into the ground, which is negligible at the multiyear time scale of our application; and L is the latent heat of vaporization of water. The Priestley–Taylor formula was chosen because it is a transparent coupling of theory and empiricism. For $\alpha = 1$, (8) expresses the equilibrium evaporation rate, a hypothetical rate over a moist surface, at steady state, at large fetch (Slatyer and McIlroy 1961). On the basis of observations, Priestley and Taylor settled on a value of $\alpha = 1.26$ for nonwater-stressed surfaces (e.g., a forest with ample soil water), and subsequent studies generally have supported a central estimate near this value (Brutsaert 1982, 219–221). Undoubtedly, the value is a complex function of atmospheric processes, which are not modeled perfectly in climate models. To allow for the dependence of α on differing approaches to modeling atmospheric processes, we use climate-model outputs to estimate a value of α for each climate model, as described in section 3.2.

3. Data and computations

3.1. Data from PRMS hydrologic modeling

Hay et al. (Hay et al. 2010) ran PRMS for 14 river basins with a historical (baseline) 12-yr (based on water years starting with October) daily climate time

Table 1. River-basin-identifying information and drainage area.

Short name	River	Stream gauge location	ID of USGS stream gauge	Drainage area (km^2)
Cathance, ME	Cathance Stream	Edmunds, ME	01021230	85
Clear, IA	Clear Creek	near Coralville, IA	05454300	254
East, CO	East River	Almont, CO	09112500	748
Feather, CA	Feather River	(constructed from subbasin gages)		9324
Flathead, MT	South Fork of the Flathead River	near Columbia Falls, MT	12362500	4307
Flint, GA	Flint River	Montezuma, GA	02349500	7511
Pomperaug, CT	Pomperaug River	Southbury, CT	01204000	194
Sprague, OR	Sprague River	near Chiloquin, OR	11501000	4053
Starkweather, ND	Starkweather Coulee	near Webster, ND	05056239	543
Yampa, CO	Yampa River	at Steamboat Springs, CO	09239500	1471

series and with numerous perturbed climate time series. Each perturbed time series was built from the baseline time series by addition of a temperature difference and multiplication by a precipitation factor. The differences and factors used in these adjustments were obtained from the climate-model outputs. Each such adjustment, computed for each month of the year, corresponded to a moving 12-yr future time window applied to the output of a single climate model that was run with a particular climate-forcing scenario from the Special Report on Emissions Scenarios (SRES; Nakicenovic et al. 2000). Differences and factors were computed relative to the historical climate computed with the same model. The adjustments were derived from data archived at the Lawrence Livermore National Laboratory (United States) for the Coupled Model Intercomparison Project (CMIP). We obtained all basin- and time-averaged (over 11 years, after a year for spinup) values of precipitation, runoff (discharge), and potential evapotranspiration that were available from the PRMS study at the time of our analysis. Data available at the time of our analysis limited us to 10 of the 14 basins. The 10 basins, with drainage areas from 85 to 9324 km^2, are distributed across the contiguous United States (Table 1) and span a wide range of climatic moisture availability (Figure 1); the basins range from snow dominated to snow free. Only three of the five climate models were used because we could not obtain climate-model runoff data from one climate model and we could not reconcile climate-model outputs of climate variables with PRMS outputs for another climate model. The three climate models used were the Institute of Numerical Mathematics Coupled Model, version 3.0 [INM-CM3.0 (INM); Russia]; Model for Interdisciplinary Research on Climate 3.2, medium-resolution version [MIROC3.2(medres) (MIROC); Japan]; and the Commonwealth Scientific and Industrial Research Organisation Mark version 3.0 [CSIRO-Mk3.0 (CSIRO); Australia]. For all three models, we used PRMS runoff-change results only for the SRES A1B scenario.

3.2. Data from CMIP

For analysis of water balance in the climate models, we obtained, directly from the CMIP database, the historical and SRES A1B time series of the near-surface air temperature, precipitation, and runoff for the climate-model grid cell containing

each river basin. We also obtained CMIP data on (upward and downward, short-wave and longwave) surface radiation and used them to compute annual surface net radiation time series for each basin. CMIP data for surface atmospheric pressure from the SRES A1B scenario were not available for all models, so we obtained and used only the historical CMIP pressure data for the evaluation of γ in (8); on the basis of an examination of pressure changes from one model and on the basis of simple order-of-magnitude estimates for the other models, we determined that the effect of change in pressure on γ is minuscule. All CMIP data (and quantities derived therefrom) were time averaged to correspond to the same time periods used for the hydrologic modeling.

The climate-model e_p was estimated by use of (8), applied with monthly values of net radiation, temperature, and surface atmospheric pressure, and then time averaged. A single value of α was estimated independently for each climate model. We set α to $(1 + 0.1n)$, choosing the smallest integer n that would result in e_p greater than or equal to climate-model actual evapotranspiration (estimated as climate-model p-r) for all basins during the baseline period.

4. Results

4.1. Values of α and ν

Estimates obtained for the Priestley–Taylor α, following the method described in section 3.2, were 1.3 for CSIRO, 1.1 for INM, and 1.2 for MIROC. These values are reasonably consistent with the nominal value of 1.26 from observations. On the basis of the estimated values of α, climate-model data are included in Figure 1, which reveals large biases in the water balances of the climate models.

Estimated values of ν varied substantially across basins, but systematic differences across models also were apparent. For PRMS, the inferred values spanned the range 0.6–2.8 (median 1.45). For CSIRO, the range was 0.9–11.3 (median 2.95); for INM, it was 0.8–4.0 (median 2.1); and, for MIROC, it was 1.4–6.0 (median 3.15). Systematically higher values for all climate models indicate a climate-model tendency to produce less runoff and more evapotranspiration than the PRMS models for a given climate (Figure 1). Because the PRMS models had been calibrated for each basin, this result is suggestive of a climate-model bias toward production of insufficient runoff and excessive evapotranspiration.

4.2. Evaluation of generalized Turc–Pike representation

For the decomposition by means of (7), we focused on the final time period (2088–99) under the A1B scenario. We first compared estimates of $(\delta^H r)/r^H - (\delta^C r)/r^C$ predicted by the generalized Turc–Pike formula via (2), (4), and (5) to the actual values determined from PRMS and the hydrologic models (Figure 2). This comparison is important to establish the validity of the generalized Turc–Pike formula as the basis for decomposition. Specifically, we ask whether the generalized Turc–Pike equation, with a value of ν fitted using only baseline data, can predict the PRMS sensitivity to climate change. The first black bar in each set in Figure 2 is the difference (PRMS minus climate model) in the relative change of runoff associated with
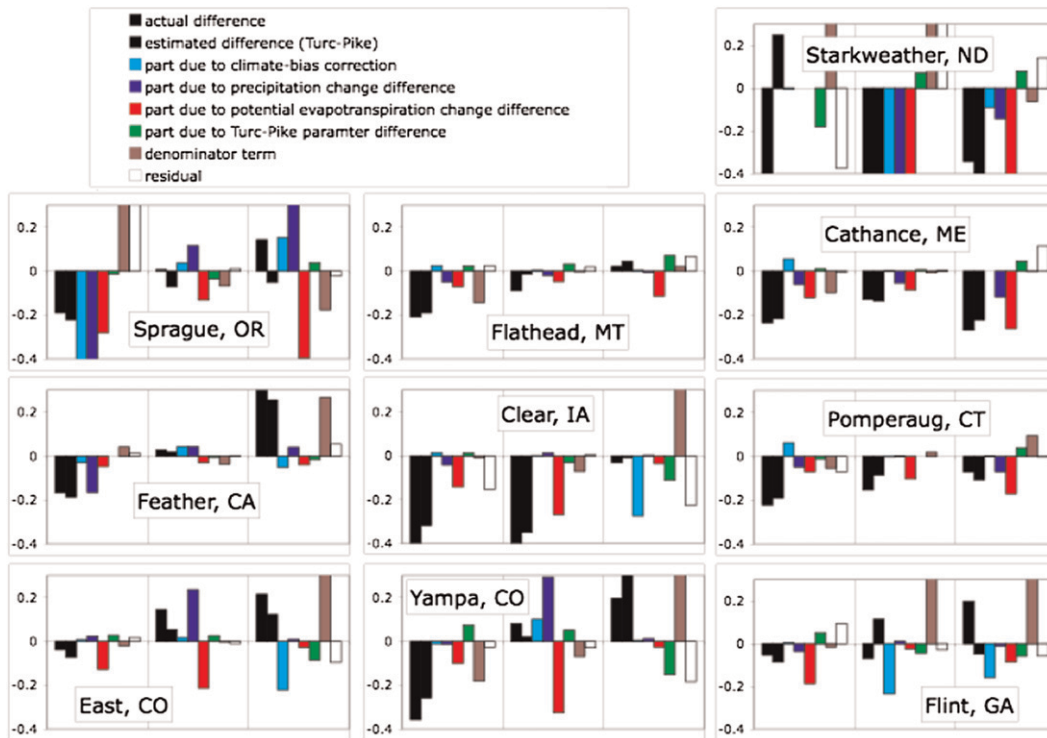
Figure 2. Bar graphs showing differences, and decompositions thereof, in estimated relative change in runoff $(\delta^H r)/r^H - (\delta^C r)/r^C$ for the time period 2088–99 under the A1B scenario, expressed relative to the baseline period 1988–99. Each panel represents one river basin, and the three sections of each panel represent climate changes from the three climate models (from left to right: CSIRO, INM, and MIROC). The first black bar in each section represents the difference $(\delta^H r)/r^H - (\delta^C r)/r^C$ obtained directly from the models. The second black bar represents the prediction of that difference obtained from the generalized Turc–Pike equation; that difference, in turn, is decomposed into the six remaining contributions: from removal of baseline climate bias (light blue); from difference in relative change of precipitation (dark blue); from difference in potential evapotranspiration (red); from difference in υ (green); from denominator effect (brown); and residual (unfilled).

climate change, $(\delta^H r)/r^H - (\delta^C r)/r^C$, computed directly from the respective model outputs. For the pooled 30 cases (10 basins, each with climate changes from three climate models), the values of the quantity (median -7.9%, range from -65% to $+67\%$) indicate that the effect of hydrologic adjustment is substantial. The second black bar in each set is the prediction of $(\delta^H r)/r^H - (\delta^C r)/r^C$ that is made on the basis only of the generalized Turc–Pike equation and information about model baseline climate and runoff. The median is -7.8% and the range is from -67% to $+31\%$. The overall correspondence between the direct computations and the generalized Turc–Pike predictions indicates that the generalized Turc–Pike formula captures the essence of runoff sensitivities of both PRMS and the climate models. The worst-case comparison was found for the Starkweather Coulee (North Dakota) with the CSIRO

model; this failure likely was somehow related to a very low baseline runoff (0.03 mm yr^{-1}) in the climate model. Despite some notable discrepancies, we judged that the generalized Turc–Pike formula nevertheless was sufficient to give reasonable estimates of the relative importance of the various factors in hydrologic adjustment.

4.3. Decomposition of differences in runoff change

By use of (7), the second black bar in each set in Figure 2 is decomposed exactly into the six bars to its right. The first four of the six are the runoff sensitivities to inputs, the fifth is the "denominator term" arising from differences in baseline runoff, and the sixth is the residual. The magnitudes of the residuals indicate the extent to which the linear perturbation terms fail to capture the difference in total sensitivity. In many cases, this term (the unfilled bars in Figure 2) was small. Because this is a nonlinear term, its largest magnitudes were found in situations where some other terms also were relatively large. In general, though, we judged that the residual was sufficiently small that we could consider the perturbation terms as at least qualitatively representative of the influence of the associated effects.

The first perturbation term, indicated by the lighter blue bars in Figure 2, is the component of hydrologic adjustment associated with removal of bias in the baseline climate. For the 30 cases, the median value of this term is less than 0.2%, and the values range from -113% to $+15\%$. In half the cases, the absolute value of the term is smaller than 2.5%. Not surprisingly, the largest magnitudes of this term were associated with large biases in the control climate of the climate model.

The second perturbation term, indicated by the darker blue bars in Figure 2, is the component of hydrologic adjustment associated with differences in relative change in precipitation applied in PRMS (Hay et al. 2010) versus those we extracted directly from the climate models. The median value of this term is less than 1% in magnitude, and the range is from -71% to $+35\%$. The absolute value is less than 4.2% in half the cases. The differences in precipitation change arise because different "downscaling" techniques were used between this study (simple nearest-neighbor interpolation of climate-model output) and the PRMS basin studies (method varying across basins; L. E. Hay 2009, personal communication).

The third perturbation term, indicated by the red bars in Figure 2, is the component of hydrologic adjustment associated with differences in relative change in potential evapotranspiration applied in PRMS versus those implicit in the climate model. The median value of this term is -11%. All values are negative, and the range is from -166% to a value less than 1% in magnitude.

The fourth perturbation term, indicated by the green bars in Figure 2, is the component of hydrologic adjustment associated with differences in basin response to a given climate forcing, embodied here in the generalized Turc–Pike parameter v. The median value of this term is $+0.8\%$ with a range from -18% to 8%. Half the values are less than about 3.7% in magnitude.

The fifth term, indicated by the brown bars in Figure 2, represents the denominator effect of the change in baseline runoff from climate model to PRMS. The median value is -0.3%, and the range is from -18% to $+96\%$. The distribution of this quantity is markedly bimodal with most values scattered around 0 but a substantial number (7) of the values in the 40%–100% range. These large values are
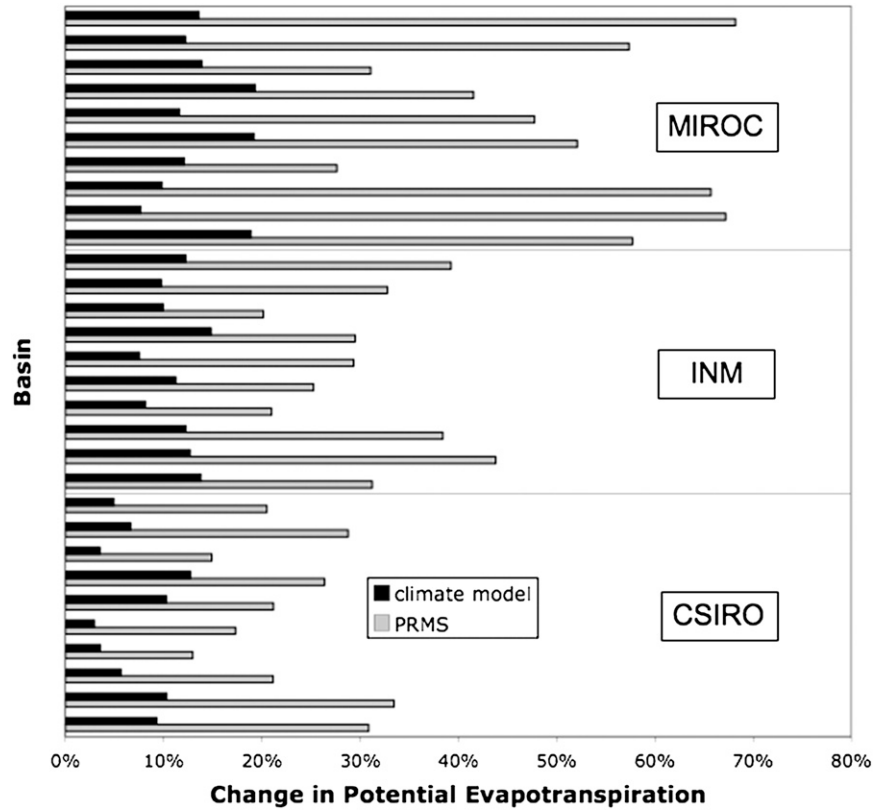
Figure 3. Changes in potential evapotranspiration for climate models and for PRMS for the time period 2088–99 under the A1B scenario, expressed relative to the baseline period 1988–99. Three sections of the chart correspond to the three climate models from which climate-change information is obtained. Within each section, basins are ordered alphabetically from the bottom of the chart upward by short name (Table 1).

generated in cases where the climate model estimates a baseline runoff much less than the baseline runoff in PRMS. That the large-magnitude values of this term should all be positive is consistent with the result that v tends to be larger in the climate models than in PRMS. The excessive values of v lead to systematic overestimation of evapotranspiration and underestimation of runoff in a given climate (Figure 1), explaining why PRMS baseline runoff can be substantially larger than climate-model runoff in some cases.

4.4. Comparison of changes in potential evapotranspiration

We have seen that the contributor to $(\delta^H r)/r^H - (\delta^C r)/r^C$ with by far the largest median value is that associated with differences in changes of potential evapo-transpiration. Because e_p plays the dominant role in the differences, we compare the changes in e_p (Figure 3). Climate models project increases in e_p that we estimated

via (8) to range from 3% to 19% with a median value of 11%, whereas the PRMS changes range from 13% to 68% with a median value of 31%. The ratio of the relative change in e_p in PRMS to the relative change in e_p in the climate models ranges from 2.0 to 8.8 with a median value of 3.3.

5. Discussion

Our analysis reveals an issue that might affect many of the studies in which projected climate change is used to drive hydrologic models or simpler drought indices to generate projections of hydrologic change. Many hydrologic models, as well as simpler drought or moisture indices (Dai et al. 2004), do not consider energy balance in the computation of potential evapotranspiration but rather employ empirical formulas, most frequently using air temperature as a key input. The specific formula used in the modeling study analyzed here differs from that used in many other studies, but the replacement of the energy-balance constraint with an empirical temperature function is commonplace in the field.

In this study, the threefold difference in magnitude of e_p change between climate-model estimates and those obtained from the modified Jensen–Haise (Jensen et al. 1970) formula is cause for concern because it reveals an apparent bias in some analyses of hydrologic change (and a potential for bias in others) that might be used to inform adaptation to climate change. A complete analysis of the discrepancy is beyond the scope of this paper, but pertinent issues are noted here. First, despite the physical basis of (8), which we used to estimate potential evapotranspiration from the climate models, the quantity α is an empirical parameter, which could change with changing climate. However, atmospheric processes place a strong upper bound on α, such that (8) cannot allow e_p to exceed R_n/L appreciably in long-term averages. In addition, because the quantity $\Delta/(\Delta + \gamma)$ increases with temperature, the upper limit on α decreases with temperature. Second, the Jensen–Haise formulation was developed to quantify evapotranspiration from a reference crop on the basis of an empirical temperature-dependent correlation with only the most commonly measured component of surface radiation budget (downwelling short-wave radiation). For this reason, Brutsaert (Brutsaert 1982, 222–223) suggests that equations of the Jensen–Haise type are specific to the location and surface type for which they are developed and require calibration for local conditions; arguably, climate change might also require recalibration. Third, the Jensen–Haise formula was developed to characterize water loss at small time and space scales in the context of irrigation engineering. The formula can be expected to describe best the nonequilibrium situation where hot, dry air flows over a cool, moist surface rather than the globally more normal situation where the overlying air mass is in a state relatively close to equilibrium with the underlying surface. Sensitivity to air temperature under strong advection conditions might be much greater than under the quasi-equilibrium conditions underlying (8) and expected to prevail at the large spatial scales considered by climate models. Fourth, the Jensen–Haise formula places no (total) energy-availability limit on evapotranspiration. Fifth, and related to the fourth point, hydrologic adjustment apparently severs the feedback from land to atmosphere. If the Jensen–Haise formula were used in a climate model (a practice not being advocated here), the relatively large increase in e_p might lead,

through this feedback, to amplified surface cooling and thence a moderation of the increase in surface temperature and e_p.

This analysis used a small sample size of only three climate models. Nevertheless, in view of the consistency of results (e.g., Figures 2, 3) across climate models, the conclusions reached here do not appear to depend on the climate model but rather on the formulation of potential evapotranspiration. However, as noted in the introduction, formulations for potential evapotranspiration vary widely, so the findings of this paper do not invalidate the use of other formulations. On the other hand, nearly all hydrologic-adjustment analyses, at some stage, depend on empirical correlations with air temperature and a need for caution seems to be indicated.

In light of this study, we recommend that the projections of potential evapotranspiration change obtained from hydrologic-adjustment analyses be checked for consistency against surface energy-balance changes in the climate models. This study illustrates one way that such checking can be done.

In the past, hydrologically relevant climate-model outputs other than precipitation and near-surface air temperature were not readily accessible outside the climate-modeling centers; this is no longer the case. We therefore recommend also that consideration be given henceforth to the direct use of radiation outputs from climate models when hydrologic adjustment is performed. Because radiative fluxes constitute the major source term in the surface energy balance, they seem to us to be more physically relevant than air temperature as a measure of energy availability for potential evapotranspiration.

6. Summary

In a case study, we analyzed the adjustment, by means of a hydrologic model, of climate-model runoff-change projections. We found that relative (i.e., fractional or percentage) runoff change computed with hydrologic adjustment more often than not was less positive (more negative) than what was projected by the climate models. We partitioned this total effect of hydrologic adjustment into contributions associated with differences between the climate model and the hydrologic model with respect to three factors: 1) absence or presence of bias adjustment for the baseline climate, 2) applied relative changes in precipitation and potential evapotranspiration, and 3) river-basin characteristics. For the case study, we found that the largest of these contributions (a median 11% decrease in runoff) arose from differences in the applied relative change of potential evapotranspiration. The large changes of potential evapotranspiration in the hydrologic model were not explicable by the combination of increases in surface net radiation and temperature-controlled turbulent-flux partitioning from a moist surface.

Analysis of specific model outputs in this paper does not imply endorsement of the models by the U.S. government.

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