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Comparison of hydrological impacts of climate change simulated by six hydrological models in the Dongjiang Basin, South China

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Summary Large differences in future climatic scenarios found when different global circulation models (GCMs) are employed have been extensively discussed in the scientific literature. However, differences in hydrological responses to the climatic scenarios resulting from the use of different hydrological models have received much less attention. Therefore, comparing and quantifying such differences are of particular importance for the water resources management of a catchment, a region, a continent, or even the globe. This study investigates potential impacts of human-induced climate change on the water availability in the Dongjiang basin, South China, using six monthly water balance models, namely the Thornthwaite–Mather (TM), Vrije Universitet Brussel (VUB), Xinanjiang (XAJ), Guo (GM), WatBal (WM), and Schaake (SM) models. The study utilizes 29-year long records of monthly streamflow and climate in the Dongjiang basin. The capability of the six models in simulating the present climate water balance components is first evaluated and the results of the models in simulating the impact of the postulated climate change are then analyzed and compared. The results of analysis reveal that (1) all six conceptual models have similar capabilities in reproducing historical water balance components; (2) greater differences in the model results occur when the models are used to simulate the hydrolog-

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ical impact of the postulated climate changes; and (3) a model without a threshold in soil moisture simulation results in greater changes in model-predicted soil moisture with respect to alternative climates than the models with a threshold soil moisture. The study provides insights into the plausible changes in basin hydrology due to climate change, that is, it shows that there can be significant implications for the investigation of response strategies for water supply and flood control due to climate change.

Introduction

Climate change associated with global warming induced by the increase in carbon dioxide and other radiative trace gases in the atmosphere has been the focus of a multitude of scientific investigations for over the past two decades (e.g., Xu et al., 2005). These investigations are driven by the recognition that climate change has significant implications for the environment, ecosystems, water resources and virtually every aspect of human life. One of the most important and immediate effects of global warming would be the changes in local and regional water availability, since the climate system is interactive with the hydrologic cycle. Such effects may include the magnitude and timing of runoff, the frequency and intensity of floods and droughts, rainfall patterns, extreme weather events, and the quality and quantity of water availability; these changes, in turn, influence the water supply system, power generation, sediment transport and deposition, and ecosystem conservation. Some of these effects may not necessarily be negative, but they need to be evaluated as early as possible because of the great socio-economic importance of water and other natural resources.

Modelling the hydrologic impacts of global climate change involves two issues: climate change and the response of hydrologic systems. Climate change is a complex problem involving interactions and feedbacks between atmosphere, oceans, and land surface. In order to better understand this problem it has been customary to use climate models which are mathematical descriptions of large-scale physical processes governing the climate system. Currently general circulation models (GCMs) are considered to be the most comprehensive models for investigating the physical and dynamic processes of the earth surface-atmosphere system and they provide plausible patterns of global climate change. However, it is not yet possible to make reliable predictions of regional hydrologic changes directly from climate models due to the coarse resolution of GCMs and the simplification of hydrologic cycle in climate models (e.g., Arora, 2001). An investigation of climate-change effects on regional water resources, therefore, consists of three steps (e.g., Xu, 1999): (1) using climate models to simulate climatic effects of increasing atmospheric concentration of greenhouse gases, (2) using downscaling techniques to link climate models and catchment-scale hydrological models or to provide catchment scale climate scenarios as input to hydrological models, and (3) using hydrological models to simulate hydrological impacts of climate change. Errors occur at every step of the investigation (Xu et al., 2005). Large differences in global and regional climate change scenarios, as calculated by the use of different GCMs and downscaling techniques, have been widely discussed in the literature (e.g., Arnell, 1995; Arora, 2001). For a given study region, the range of diversity obtained when different hydrological models are used for a given climate scenario has not been widely investigated and reported in the literature. A few examples of studying the differences in hydrological impacts of climate change obtained by different hydrological models include Boorman and Sefton (1997) who investigated the range of diversity resulting from the use of two conceptual hydrological models. The study concluded that both models exhibited similar capabilities in reproducing historical flow indices; however, significant differences existed in the simulated flow indices of the changed climate even for the same climate scenario and for the same catchment. The study also suggested that more studies on comparing the impacts due to different hydrological models need to be carried out; this would help quantify the differences resulting from different hydrological simulation models and provide useful guidelines for water resources planners and managers. Panagoulia and Dimou (1997a,b) examined the differences in predictions of two hydrological models under both historical and alternative climate conditions. The models for comparison were a monthly water balance (MWB) model and the snow accumulation-ablation (SAA) and soil moisture accounting (SMA) models of the US National Weather Service (US NWS). It was shown that the models had small differences in monthly runoff values and greater interannual variability for the SAA—SMA models under historical climate conditions. When the alternative climates were used as input to the hydrological models, greater differences were obtained with different models. The MWB model showed a greater runoff increase in winter and a greater decrease in summer as compared with the SAA—SMA model. Whereas the SMA model soil moisture varied substantially, the MWB model soil moisture remained unaffected for any climate during winter. The soil moisture reduction predicted from the MWB model was greater than that predicted from the SMA model in late spring and summer.

The concept of using regional hydrologic models for assessing the impact of climate change has several attractive features (Gleick, 1986; Schulze, 1997). First, models tested for different climatic/physiographic conditions as well as models structured for use at various spatial scales and dominant process representations are readily available. This affords flexibility in identifying and choosing the most appropriate approach to evaluate any specific region. Second, hydrological models can be tailored to fit the characteristics of available data. The GCM-derived climate perturbations (at different levels of downscaling) can be used as model input. A variety of responses to climate change scenarios can hence be modelled. Third, regional-scale hydrological models are considerably easier to manipulate than general circulation models. Fourth, such regional models can be used to evaluate the sensitivity of specific watersheds to both hypothetical changes in climate and to changes predicted by large-scale GCMs. Finally, the models that can incorporate both detailed regional hydrologic characteristics and output from large-scale GCMs will be well situated to take advantage of continuing improvements in the resolution, regional geography, and hydrology of global climate models.

It would not be unfair to say that all kinds of models find their usefulness in different applications. Physically-based distributed-parameter models are complex in terms of structure and input requirements and can be expected to provide adequate results for a wide range of applications. On the other hand, simpler models which have a smaller range of applications can yield adequate results at greatly reduced cost, provided that the objective function is suitable. The distinction between simple and physically-based distributed-parameter models is not only one of lesser or greater sophistication, but is also intimately linked with the purpose for which such models are to be used. Thus, choosing a suitable model is equivalent to distinguishing the situation between when simple models *can* be used and when complex model *must* be used. The choice of a model for a particular study depends, therefore, on many factors (Gleick, 1986), amongst which the purpose of study and model and data availability have been the dominant ones (Ng and Marsalek, 1992; Xu, 1999). For example, for assessing water resources management on a regional scale, *monthly rainfall-runoff (water balance) models* were found useful for identifying hydrologic consequences of changes in temperature, precipitation, and other climate variables (e.g., Gleick, 1986; Schaake and Liu, 1989; Mimikou et al., 1991; Arnell, 1992; Xu and Halldin, 1997; Xu and Singh, 1998). For detailed assessments of surface flow and other water balance components, *conceptual lumped-parameter models* are used. One of the more frequently used models in this group is the Sacramento Soil Moisture Accounting Model (Burnash et al., 1973). Many researchers have used the same model or similar models for studying the impact of climate change (e.g., Nemec and Schaake, 1982; Gleick, 1987; Lettenmaier and Gan, 1990; Schaake, 1990; Nash and Gleick, 1991; Cooley, 1990; Panagoulia, 1992; Leavesley, 1994; Panagoulia and Dimou, 1997a,b). In Nordic countries the HBV model is widely used as a tool to assess the climate change effects (e.g., Vehviläinen and Lohvansuu, 1991). For simulation of spatial patterns of hydrological response within a basin, *process-based distributed-parameter models* are needed (Beven, 1989; Thomsen, 1990; Running and Nemani, 1991; Bathurst and O'Connell, 1992). For estimating changes in the average annual runoff for different climate change scenarios simple empirical and regression models have been used, as for example, the models by Revelle and Waggoner (1983) in the United States, and Arnell and Reynard (1989) in the UK.

The studies reported in the literature and discussed above have used one or a limited number of hydrological models (2 models in some cases) to simulate the impact of postulated climate changes; these studies represent the results of the selected models only. It is therefore desirable to compare the differences in hydrological impacts of alternative climates resulting from the use of more hydrological models. In this study, six monthly water balance models (i.e., Thornthwaite–Mather model (Alley, 1984), VUB model (Vandewiele et al., 1992), the monthly Xinanjiang model (Hao and Su, 2000), Guo model (Guo, 1992), WatBal model (Kaczmarek, 1993; Yates, 1996) and Schaake model (Schaake, 1990)) are used and their capabilities in reproducing historical water balance components and in predicting hydrological impacts of alternative climates are compared. The choice of these six models is based on the following considerations: First, climate change scenarios at a monthly time scale are easily available and more reliable. Second, water resources management and planning for a large basin or region are generally on a monthly time scale. Third, large scale observed hydrological data on a monthly scale are more easily available for calibrating and validating hydrological models, especially in developing countries.

Thus the main objective of this study is to quantify how large the difference one can expect when using different hydrological models to simulate the impact of climate change as compared to the model capabilities in simulating historical water balance components. The study is performed in two steps: First, the performance of models in reproducing historical water balance components is evaluated; and second, the differences in the simulated hydrological consequences of changed climate by various models are evaluated and compared.

Study area and data

The study area is the Dongjiang (East River) catchment, a tributary of the Pearl River in southern China. The Dongjiang catchment is located in the Guangdong and Jiangxi provinces (see Fig. 1). Originating in the Xunwu county of Jiangxi province, the river flows from north-east to south-west and discharges into the Zhujiang (the Pearl River) estuary with an average gradient of 0.39‰. The Dongjiang water is being transferred out of the basin to Hong Kong since the mid 1960s. The proportion of the Dongjiang water in Hong Kong's annual water supply has steadily increased from only 8.3% in 1960 to about 70% or even slightly over 80% in recent years. However, the water resources of the Dongjiang basin are already heavily committed. Rapid economic development and growth of population in this region have caused serious concerns over the adequacy of the quantity and quality of water withdrawn from the Dongjiang River in the future. Any significant change in the magnitude or timing of runoff or soil moisture in the Dongjiang basin induced by changes in climate variables would thus have important implications for the great economic success and prosperity of the Pearl River Delta region and Hong Kong. Therefore, a modelling study is urgently needed that can evaluate the potential impacts of future climate change on the water availability in this basin.

The Dongjiang basin has a sub-tropical climate with a mean annual temperature of about 21 °C and only occasional incidents of winter daily air temperature dropping below 0 °C in the mountainous areas of the upper basin. The average annual rainfall for the period of 1960–1988 is 1747 mm, and the average annual runoff is 935 mm, or roughly 54% of the annual rainfall. Precipitation is generated mainly by two types of storms: frontal type and typhoon-type rainfalls. There are large seasonal changes in rainfall and runoff in the

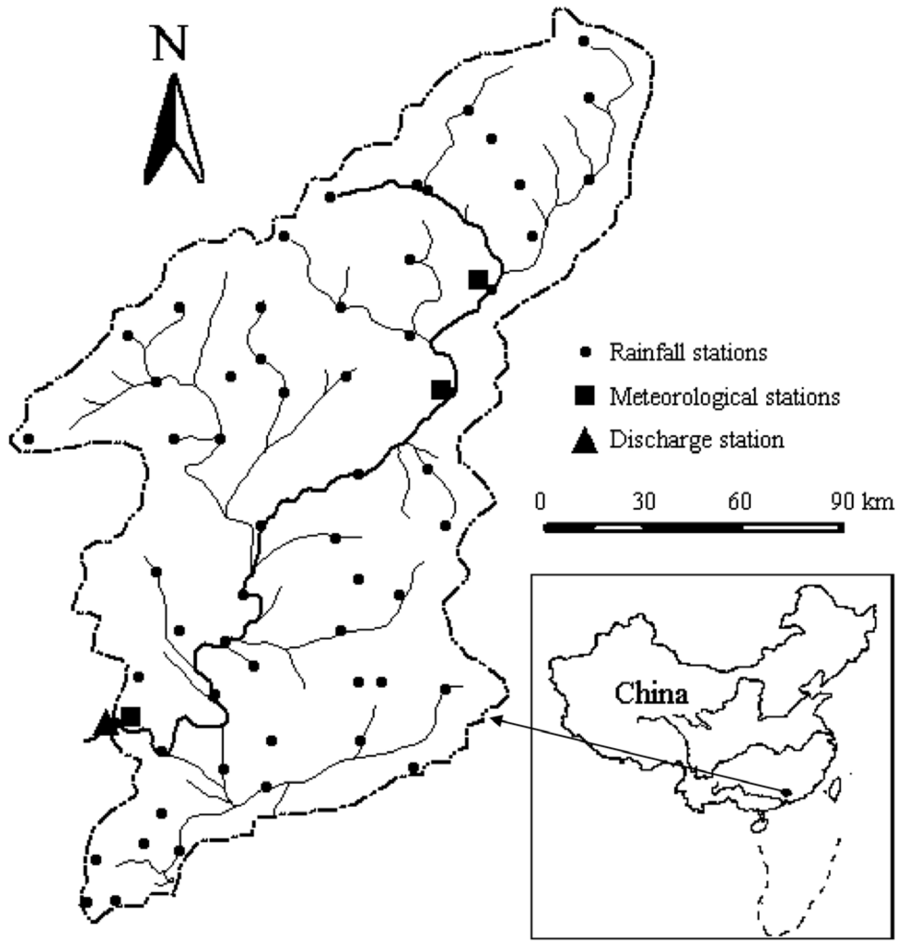

Figure 1 Location of the Dongjiang basin and the hydro-meteorological stations.

catchment: about 80% of the annual rainfall and runoff occur in the wet season from April to September, and about 20% occurs during the dry period of October to March. The geology of the catchment is complex. Precambrian, Silurian, and Quaternary geological formations are encountered at the surface with granites, sandstone, shale, limestone, and alluvium. The landscape is characterized by hills and plains, comprising 78.1% and 14.4% of the basin area, respectively. Forest covers upper elevations and intensive cultivation dominates hills and plains.

The data for model calibration and validation include monthly air temperature, pan evaporation, rainfall and streamflow. All the data were obtained from the published Yearly Hydrological Books of China for the period from 1960 to 1988. In this study, 46 major stations, almost evenly distributed over the whole basin with 29-year continuous records (1960–1988), were selected for model calibration and validation. Many methods are available for estimating mean areal rainfall over an area (e.g., a catchment) based on the results of meteorological observations (Naoum and Tsanis, 2004), which include Spline (Regularized & Tension), Inverse Distance Weighting (IDW), Trend Surface, Kriging, and Thiessen Polygons. The density of the gauge stations and their distribution are important for the accuracy of all the methods. In this study, the model input of areal rainfall was calculated from the records of the 46 stations using the Thiessen polygon method. This is mainly because (1) the Thiessen polygons are probably the most common approach for modeling the spatial distribution of rainfall and the Thiessen method is known to provide good results when used for relatively dense networks (Naoum and Tsanis, 2004); and (2) the rainfall stations are rather dense and evenly distributed (see Fig. 1) in the catchment that makes the differences resulting from using different interpolation and area averaging methods small. Pan (Φ80 cm) evaporation data of 3 meteorological stations were averaged to estimate areal potential evaporation, i.e., $PE = kE_{pan}$, where coefficient k is calibrated in the study and an average value of $k = 0.64$ was obtained. The use of a constant coefficient is based on two considerations. First, the study by Xu and Vandewiele (1994) has shown that this type of monthly water balance model is much less sensitive to the evaporation input than to precipitation. Second, the main objective of the study is to compare the simulation capabilities of the six chosen water balance models under historical climate and alternative climate with respect to the same input data series. The monthly streamflow data were from the Boluo stream gauging station, above which the drainage area is 25,325 km^2. Fig. 2 presents areal

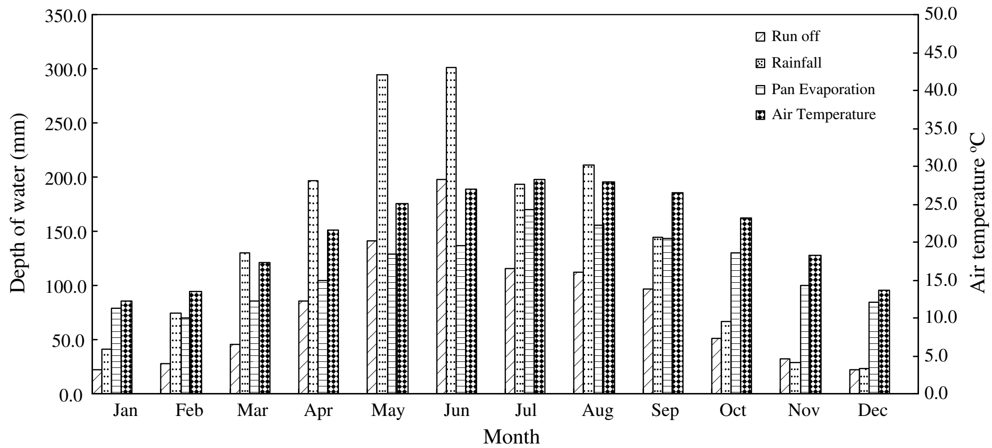

Figure 2 Mean monthly pan evaporation, rainfall, runoff and air temperature of the Dongjiang basin.

mean monthly values over the 29-year record for observed pan evaporation, rainfall, runoff, and air temperature for the Dongjiang basin.

Methodology

Selected monthly water balance models

Almost all monthly water balance models share the same general structure which includes the water balance equation at the monthly time scale:

$$S(t+1) = S(t) + P(t) - E(t) - Q(t) \quad (1)$$

in which $S(t)$ represents the amount of soil moisture stored at the beginning of the time interval t; $S(t+1)$ represents the storage at the end of that interval; and the flow across the control surface during the interval consists of precipitation P(t), actual evapotranspiration, $E(t)$, and soil moisture surplus, $Q(t)$, which supplies streamflow and groundwater recharge. Solution of this equation requires dealing with time series of four variables: S, P, E, Q, and possibly of other variables related to them.

Monthly water balance models represent the entire catchment hydrology as a series of moisture storages and flows. The water balance models differ in how E and Q are conceptually considered and mathematically represented. In order to estimate the actual evapotranspiration in the soil—water budget method many investigators have used a soil—moisture extraction function or coefficient of evapotranspiration which relates the actual rate of evapotranspiration E to the potential rate of evapotranspiration PE based on some function of the current soil moisture content and moisture retention properties of the soil (Xu and Singh, 2004).

It was desired to consider a broad spectrum of hydrological models and therefore six monthly models (i.e., Thornthwaite—Mather model (Alley, 1984), VUB model (Vandewiele et al., 1992), the monthly Xinanjiang model (Hao and Su, 2000), Guo model (Guo, 1992), WatBal model (Kaczmarek, 1993; Yates, 1996) and Schaake model (Schaake, 1990)) with 2—5 parameters, all of which have been widely employed in simulating water resources of stationary and changed climate conditions, were employed in this study. All six models are described in the Appendix (a complete list of equations can be found in the cited references); only their similarities and differences are briefly discussed here in order to provide background information in interpreting and understanding the similarities and differences in the simulation results. The structural characteristics of the models employed are shown in Table 1.

Table 1 Characteristics of the selected models

Model	No. of soil zones	No. of storages (deficit)	Types of storage (deficit)	Runoff components
TM	1	2	Soil moisture storage, Water surplus	Runoff
VUB	1	1	Soil moisture storage	Fast runoff, Slow runoff
XAJ	3	5	Upper layer tension storage, Lower layer tension storage, Deep layer tension storage, Free water storage, Groundwater storage	Fast runoff, Slow runoff
GM	1	2	Soil moisture storage, Groundwater storage	Surface runoff, Interflow, Groundwater flow
WM	1	1	Relative soil moisture storage	Direct runoff, Surface flow, Sub-surface flow, Baseflow
SM	1	1	Soil moisture deficit	Surface runoff, Groundwater flow

In all of the aforementioned models, three distinct components can be identified: monthly streamflow, evapotranspiration and soil moisture accounting, which are treated differently in each model. Soil moisture accounting is the most important component which expresses the balance between soil moisture content, incoming (precipitation), and outgoing (evapotranspiration and runoff) quantities.

Actual evapotranspiration in all of the models is treated as a function of potential evapotranspiration and soil moisture storage. The Thornthwaite–Mather (called TM model hereafter) Guo (called GM model hereafter) models assume that evapotranspiration takes place at potential rate for a given month when precipitation is equal to or greater than the potential evapotranspiration. In the WatBal (called WM model hereafter) and Schaake (called SM model hereafter) models, evapotranspiration is assumed to be equal to the potential evapotranspiration when soil moisture storage reaches the maximum capacity. The Belgium (called VUB model hereafter) model is an exceptional case, in which evapotranspiration is equal to the potential evapotranspiration in the case when the available water (precipitation plus available storage) is greater than the potential evapotranspiration. In all of these models, when precipitation or available water is less than the potential evapotranspiration, soil moisture deficit occurs. Then, the actual evapotranspiration is determined by using a non-linear function of potential evapotranspiration and soil moisture content in TM, VUB, GM and WM models. On the other hand, the SM model calculates actual evapotranspiration as a proportion to the ratio of the soil moisture content to the maximum storage capacity. In all of these five models soils are considered to be a one layer medium for moisture accounting. The monthly Xinanjiang (called XAJ model hereafter) model is the only one in which actual evapotranspiration is estimated from detailed relationships with plant root characteristics and moisture content distribution in the soil profile. Evapotranspiration in the monthly Xinanjiang model may decrease from the potential rate in the upper layer of very thin top soil to actual rate in the lower layers. The approaches used in all of these models do not consider the influence of vegetation and the interaction between vegetation and atmosphere. However, they have the advantage of minimal input data requirement and ease of application.

In the treatment of soil moisture accounting there are large differences among the models. All the models, except the VUB model, adopt a threshold value of soil moisture storage capacity and rainfall in excess of this value would become runoff. In the TM, GM and XAJ models, the precipitation for the current month does not produce any runoff unless soil moisture content reaches the storage capacity. In the WM model baseflow is a pre-determined threshold value and surface runoff emerges only when the current monthly effective precipitation exceeds the baseflow. In the SM model the threshold controls the generation of groundwater runoff.

In each model, with the exception of the TM model, the remaining content of soil moisture after evapotranspiration eventually contributes to various runoff components. The TM model does not differentiate surface runoff from groundwater flow.

A distinction is made between quick runoff and slow runoff in both the VUB model and the XAJ model. Quick (or fast) runoff and slow runoff correspond to surface runoff and groundwater flow, respectively, in the SM model. Although the two runoff components in all three models are the same, the mechanisms of runoff generation are completely different. In the VUB model, the mechanism of quick runoff can be seen as a translation of the variable-source-area concept of runoff generation: the greater the available water, the wetter the catchment; the larger the 'source-area' of surface runoff, the greater the part of the 'active' precipitation running off rapidly. The slow runoff component is proportional to the soil moisture content at the beginning of the month.

The runoff generation process described in the monthly XAJ model is based on the concept of soil moisture storage repletion. Specifically, runoff generation results from the soil moisture in the excess of field capacity in the aeration zone. Part of the generated runoff contributes to quick flow through a free water storage reservoir, and the rest is added to the groundwater storage. Groundwater contributes to the formation of slow flow through a linear reservoir with a time lag of one month.

In the SM model surface runoff is generated by the effective precipitation after evapotranspiration and replenishing soil moisture. The infiltration process is described simply as a linear function of the soil moisture storage deficit. The groundwater runoff is derived from the aquifer through a non-linear reservoir.

The GM and WM models consider more runoff components than other models. Surface runoff and interflow are generated when the soil moisture capacity is reached. Part of the excess water contributes to surface runoff and another part flows to the river as interflow, and the remaining percolates to the groundwater storage. Groundwater discharges to the river as baseflow with a time lag of one month. Runoff in the WM model is divided into four parts: direct runoff, surface runoff, sub-surface runoff, and baseflow. Baseflow is given a pre-set value, while direct runoff is a fraction of precipitation. Surface runoff and sub-surface runoff are expressed as functions of relative soil moisture storage. Actually there is no obvious distinction between surface runoff and interflow/sub-surface runoff at a monthly time interval, because the distribution of precipitation is unified over time as a result of the extended time scale.

Only the XAJ model considers a certain spatial variation of the basic variables used for the computation of surface runoff. This model assumes two parabolic distributions of tension water capacity and free water storage capacity over the catchment. None of the models, except the SM model, includes any infiltration or percolation equations for describing the movement of water from the surface detention storage to soil moisture storage and consequently to groundwater storage.

River flow routing is not considered in the models, and all the flow components run off directly at the basin outlet. This is because compared with daily or event-based hydrological models, monthly hydrological models may not make a distinction between runoff production and runoff routing, and therefore have a relatively simple framework.

Methods of model calibration and validation

Some of the parameters of the six selected models cannot be directly determined from field measurement or estimated from catchment characteristics and must, therefore, be estimated by model calibration. There are many objective functions that have been used for model calibration. In this study, model parameters were optimized by minimizing the values of the objective function given by Eq. (2):

$$\mathrm{OF} = \sum_{t=1}^{n} (\mathrm{qobs}_t - \mathrm{qsim}_t)^2 \tag{2}$$

where qobs_t and qsim_t are the observed streamflow and simulated streamflow, respectively, of month t, and n is the total number of months of simulation. The main reasons for using this criterion include: (1) it is simple and directly applicable to any model, as stated by Sorooshian and Dracup (1980); and (2) the main objective of the study is to compare the model performance with respect to a common criterion rather than to compare the advantages and disadvantages of the objective functions.

The Simplex method of Nelder and Mead (1965) was chosen for parameter optimization in this study. The Simplex method is a local, direct search algorithm that has been commonly used for conceptual hydrological models (e.g., Johnston and Pilgrim, 1976; Gan and Biftu, 1996; Hendrickson et al., 1988). In order to test the accuracy of modelling results, it is a common practice to employ some criterion for judging model performance. The first requirement of a model is that it should have the ability to reproduce the mean of observed streamflow. However, the mean value cannot fully indicate how well individual simulated values match observed values. To overcome this limitation, the root mean squared error (RMSE) and the coefficient of efficiency or the Nash–Sutcliffe efficiency E (Nash and Sutcliffe, 1970) were employed.

RMSE is simply the average of the squared errors for all simulation results and provides an objective measure of the difference between observed and simulated values:

$$\mathrm{RMSE} = \frac{1}{n}\sqrt{\sum_{t=1}^{1} (\mathrm{qobs}_t - \mathrm{qsim}_t)^2} \tag{3}$$

E, as shown in Eq. (4), is a dimensionless coefficient for measuring the degree of association between observed and simulated values:

$$E = \frac{\sum_{t=1}^{n} (\mathrm{qobs}_t - \overline{\mathrm{qobs}})^2 - \sum_{t=1}^{n} (\mathrm{qobs}_t - \mathrm{qsim}_t)^2}{\sum_{t=1}^{n} (\mathrm{qobs}_t - \overline{\mathrm{qobs}})^2}, \tag{4}$$

where $\overline{\mathrm{qobs}}$ is the mean of observed discharges. The value of E is always less than unity. A value of E equalling unity represents a perfect agreement between observed and simulated streamflows.

Determination of climate change scenarios

In simulating future water resources scenarios, future climate change scenarios are needed. Climate scenarios are sets of time series or statistical measures of climatic variables, such as temperature and precipitation, which define changes in climate. Many methods have been developed for generating climate scenarios for the assessment of hydrologic impacts of climate change, which include downscaled general circulation model (GCM) simulations and hypothetical methods. GCMs are used to generate projections of future climate change on a large spatial and temporal scale (several decades). Even as GCM grid sizes tend towards one or two degrees, there is still a significant mismatch with the scale at which many hydrologic and water resources studies are conducted (Varis et al., 2004). Given the limitations of GCMs grid-point predictions for regional climate change impact studies, an alternative option is to downscale GCM's climate output for use in hydrological models. Two categories of climatic downscaling, namely, dynamic approaches (in which physical dynamics are solved explicitly) and empirical (the so called 'statistical downscaling') are commonly employed.

Dynamic downscaling techniques have been used to develop regional climate models (RCMs) to attain a horizontal resolution on the order of tens of kilometres over selected areas of interest. This nested regional climate modelling technique consists of using initial conditions, time-dependent lateral meteorological conditions derived from GCMs (or analyses of observations) and surface boundary conditions to drive high-resolution RCMs (e.g., Cocke and LaRow, 2000; von Storch et al., 2000). Thus, the basic strategy is to use a global model to simulate the response of global circulation to large-scale forcings and an RCM to (a) account for sub-GCM grid-scale forcing (e.g., complex topographical features and land-cover inhomogeneity) in a physically-based way; and (b) enhance the simulation of atmospheric circulations and climate variables at fine spatial scales (upto 10–20 km or less). The main theoretical limitations of this technique that remain to be improved include (Hay et al., 2002; Varis et al., 2004): (1) the inheritance of systematic errors in the driving fields provided by global models. For example, boundary conditions from a GCM might themselves be so biased that they impact the quality of regional simulation, complicating the evaluation of the regional model itself (e.g., Hay et al., 2002); (2) lack of two-way interactions between regional and global climate; and (3) the algorithmic limitations of the lateral boundary interface. Other limitations are: (1) depending on the domain size and resolution, RCM simulations can be computationally demanding, which has limited the length of many experiments to date, and (2) there will remain the need to downscale the results from such models to individual sites or localities for impact studies (Wilby and Wigley, 1997; Xu, 1999). It is essential that the quality of GCM large-scale driving fields continues to improve, as those impact regional climate simulations (Varis et al., 2004).

A second, less computationally demanding, approach is statistical downscaling. In this approach, regional-scale atmospheric predictor variables (such as area-averages of precipitation or temperature, and circulation characteristics (such as mean sea level pressure or vorticity) are related to station-scale meteorological series (Hay et al., 1991, 1992; Karl et al., 1990; Kim et al., 1984; Wigley et al., 1990). The statistics involved can be simple or extensive, but the final relationships typically arrived at are as with some form of regression analysis. Statistical downscal-

Table 2 Hypothetical climate change scenarios

Scenario no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
ΔT (°C)	1	1	1	1	1	2	2	2	2	2	4	4	4	4	4
ΔP (%)	−20	−10	0	10	20	−20	−10	0	10	20	−20	−10	0	10	20

ing methods can be classified into three categories (Wilby and Wigley, 1997; Xu, 1999), namely, regression methods (e.g., Kim et al., 1984; Wigley et al., 1990; von Storch et al., 1993); weather-pattern based approaches (e.g., Lamb, 1972; Hay et al., 1991; Bardossy and Plate, 1992; Wilby, 1995); and stochastic weather generators (e.g., Richardson, 1981; Wilks, 1992; Gregory et al., 1993; Katz, 1996). In reality, many downscaling approaches embrace the attributes of more than one of these methods and therefore tend to be hybrid in nature. Compared with dynamic downscaling, statistical downscaling methods have the following advantages (von Storch et al., 2000): (1) they are based on standard and accepted statistical procedures, (2) they are computationally inexpensive, (3) they may flexibly be crafted for specific purposes, and (4) they are able to directly incorporate the observational record of the region. However, the following disadvantages have also been summarized by Goodess et al. (2001): (1) They assume that predictor/predictand relationships will remain unchanged in the future, (2) they require long/reliable observed data series, and (3) they are affected by biases in the underlying GCM. The last point also exists in the dynamic downscaling methods. Furthermore, the skill of statistical downscaling depends on climatic region and season (Wetterhall et al., 2006, 2007).

Given the deficiencies of GCM predictions and downscaling techniques, the use of hypothesized scenarios as input to catchment-scale hydrological models is widely used (e.g., Nemec and Schaake, 1982; Xu, 2000; Graham and Jacob, 2000; Engeland et al., 2001; Arnell and Reynard, 1996; Leavesley, 1994; Boorman and Sefton, 1997; Panagoulia and Dimou, 1997). Various hypothetical climate-change scenarios have been adopted and the techniques for developing climate scenarios are continuously progressing. Hulme and Carter (1999), as cited by Varis et al. (2004), have presented a typology for scenario construction, which consists of the following eight stages: Scenarios based on expert judgment; Equilibrium $2\times CO_2$ scenarios; Time-dependent climate change; Multiple forcing scenarios; Climate system unpredictability; Natural climate variability; Scenarios combining uncertainties based on Bayesian logic; and Sub-grid scale variability.

Hypothetical scenarios are based on reasonably but arbitrarily specified changes in climate variables. Changes in climate variables, such as temperature and precipitation, are adjusted, often according to a qualitative interpretation of climate model predictions or the analyses of changes in climate characteristics that occurred in the past during particularly warm or cool periods of hydrometeorological observations. Adjustments might include, for example, changes in mean annual temperature of 1 °C, 2 °C, 3 °C, 4 °C or changes in annual precipitation of 5%, 10%, 15%, 20%, 25%, relative to the baseline climate. Adjustments can be made in temperature and precipitation independently or the combination thereof.

In this study hypothetical scenarios were used which were drawn from the analysis of the state-of-the-art estimates of future climate change for the region (IPCC, 2001). In order to cover a wide range of climate variability, fifteen hypothetical climate change scenarios were derived from combinations of three temperature increases and five precipitation changes (Table 2).

Results

Comparison of models results in reproducing historical records

Statistical comparisons and visual comparisons of observed and simulated values were conducted to evaluate the performance of the selected models. Table 3 gives a summary of statistics of simulations by the selected models for both calibration and validation periods. The values of *RMSE* and *E* in Table 3 indicate that all the models produce good results for calibration and validation periods and the XAJ model has the highest E value and the lowest RMSE value, followed by GM, VUB, TM, SM and WM. The values in Table 3 demonstrate the basic capability of each model to reproduce long-term mean annual observed runoff in the Dongjiang basin.

Comparison of mean monthly values of simulated runoff, evapotranspiration and soil moisture content for the period 1960—1988 are shown in Fig. 3. It is seen that (1) all six mod-

Table 3 Comparison of the results of the six models in reproducing historical discharge in the Dongjiang basin (1960—1988)

Calculation and validation		QOBS	TM	VUB	XAJ	GM	WM	SM
Calibration 1960—1974	$\overline{Q}$ (mm/month)	73.3	71.4	72.2	71.3	70.5	77.4	75.3
	RMSE		23.61	24.65	21.88	22.28	29.01	27.03
	E (%)		90.9	90.1	92.2	91.9	86.3	88.1
Validation 1975—1988	$\overline{Q}$ (mm/month)	82.8	83.4	83.2	83.3	82.6	86.4	81.8
	RMSE		22.57	22.44	20.19	21.01	27.34	25.15
	E (%)		89.9	89.9	91.8	91.1	85.0	87.3

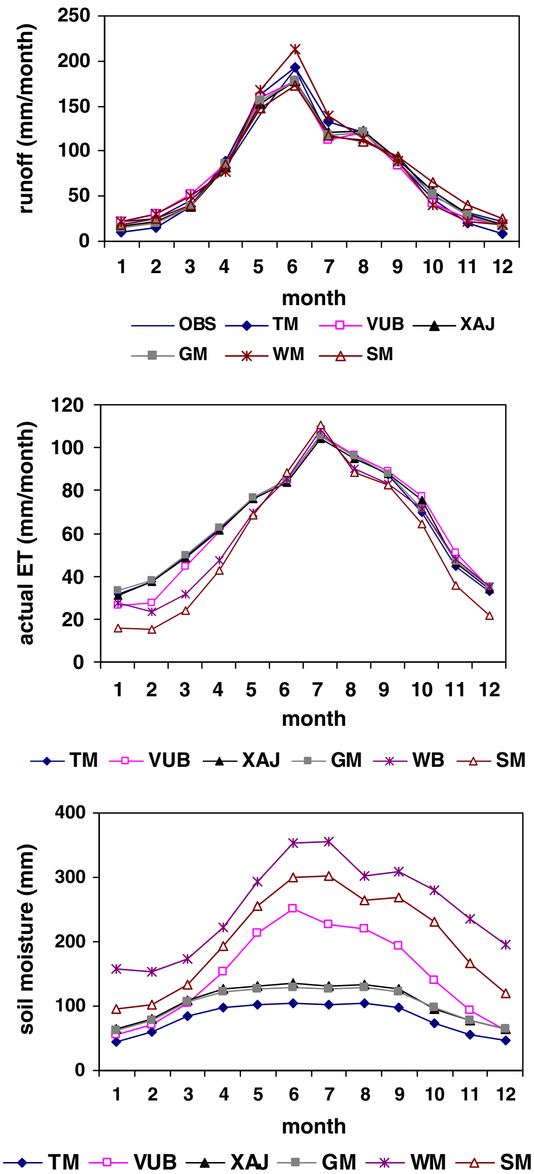

Figure 3 Comparison of mean monthly runoff (upper), actual evapotranspiration (middle) and soil moisture (lower) simulated by the six models for the period of 1960—1988.

els simulate quite well the mean monthly runoff except the WM model for the months of June and July, and the SM model for the months of October and November (Fig. 3 upper). (2) There is a good agreement in the mean monthly evapotranspiration simulated by the six models, except that the WM and SM models yield smaller values for winter and spring months (Fig. 3 middle). (3) Larger differences are found for the simulated monthly soil moisture. Similar results have been reported by Alley (1984) and Vandewiele et al. (1992) that 'state variables simulated by the conceptual monthly water balance models may be quite different', and this is partially because the soil depth is usually not explicitly defined in such models.

In order to compare the models' capabilities in simulating the dynamics of monthly runoff series, the monthly runoff values computed using different models are analyzed and correlated with observed values using a linear regression equation $Y = mX + c$. In the equation, Y represents the model calculated runoff, and X is the observed runoff, and m and c are constants representing the slope and intercept, respectively. The scatter plot and the results of regression analysis are shown in Fig. 4. It is seen from Fig. 4 that: (1) as far as the R^2 values are concerned, all model predictions correlated well with observed values resulting in R^2 values $\geqslant 0.90$ in all cases. (2) When the values of slope and intercept are compared, the differences among the models are also not so great. The values of intercept are generally less than 10% of the mean monthly values, and the slopes (m) are all within ±10% of the 1:1 line. A close look at the differences among the models reveals that the TM and GM models have the smallest bias in the regression slope, while the GM and SM models have the smallest bias in the intercept. (3) When looking at the monthly peak values, it is found that the TM, GM and WM models exhibit a bigger negative bias when compared with observed values.

Generally speaking, the results show that all six models can reproduce historical monthly runoff series with an acceptable accuracy. It must, however,be kept in mind that the purpose here is not to discuss in detail which model is superior. The main purpose is to check how diverse the models' results are with respect to historical and alternative climates.

Differences of models results in predicting hydrological response of changed climate

It is important for water resources managers to be aware of and prepared to deal with the effects of climate change on hydrological variables. Obviously, streamflow is essential in order to provide an indication of the extent of impacts of climatic change on water resources, which represents an integrated response to hydrologic inputs on the surrounding drainage basin area and therefore affords a good spatial coverage. Since it is expected that climate change will result in a diversity of environmental responses, actual evapotranspiration and soil moisture are also included in this study.

Mean annual changes

The mean annual response of hydrological variables to the 15 climate scenarios is first evaluated. The percent changes of mean annual runoff in response to the 15 climate change scenarios are shown in Fig. 5. In general, Fig. 5 shows that even on the annual level there is a wide range of differences between runoff responses simulated by the six models when perturbed climate scenarios are used to drive the models. The runoff changes in response to temperature changes for a given precipitation change are shown in Fig. 5a. Six lines in each branch of lines represent the results of six models. The slope of each line represents the changing rate for runoff as temperature increases. The differences between the five branches represent the influence of different precipitation changing scenarios. It is seen that (1) when there is no change in precipitation (six middle branch lines), one, two and four degree increase in the air temperature result in a reduction in annual runoff of about 3%, 5—8% and 8—15%, respectively, depending on the model. When precip-

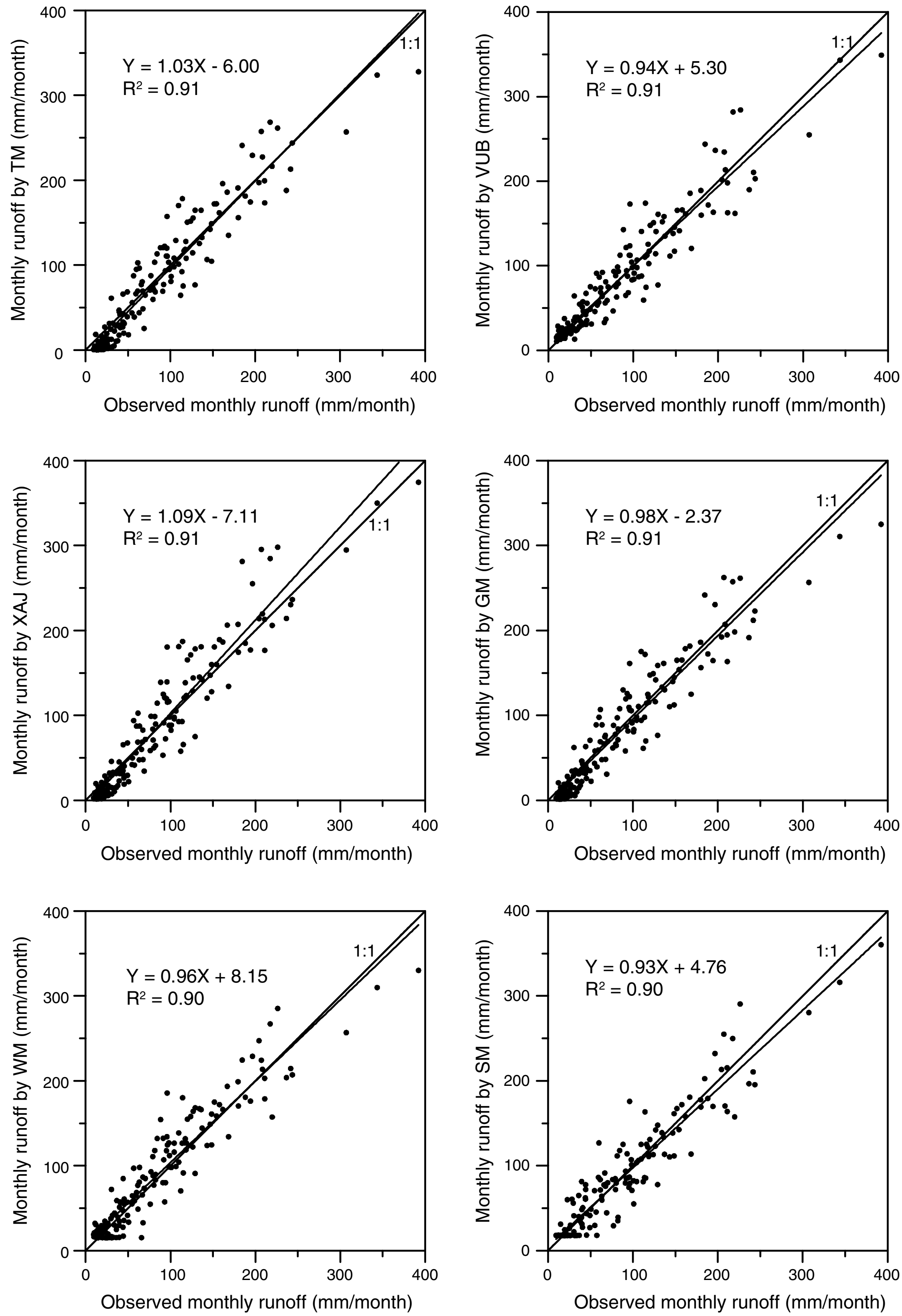

Figure 4 Scatter plot and regression equations of the monthly runoff calculated by the six models with the observed runoff.

itation decreases by 20% (six left branch lines), annual runoff decreases by about 25—40%, 28—45% and 30—50%, respectively, for temperature increases of 1 °C, 2 °C and 4 °C, depending on the model. When precipitation increases by 20% (six right branch lines), annual runoff increases by about 20—35%, 17—27% and 15—20%, respectively, for temperature increases of one, two and four degrees, depending on the model; (2) the differences between models increase as the precipitation change increases in both directions. The decrease in precipitation results in larger differences than

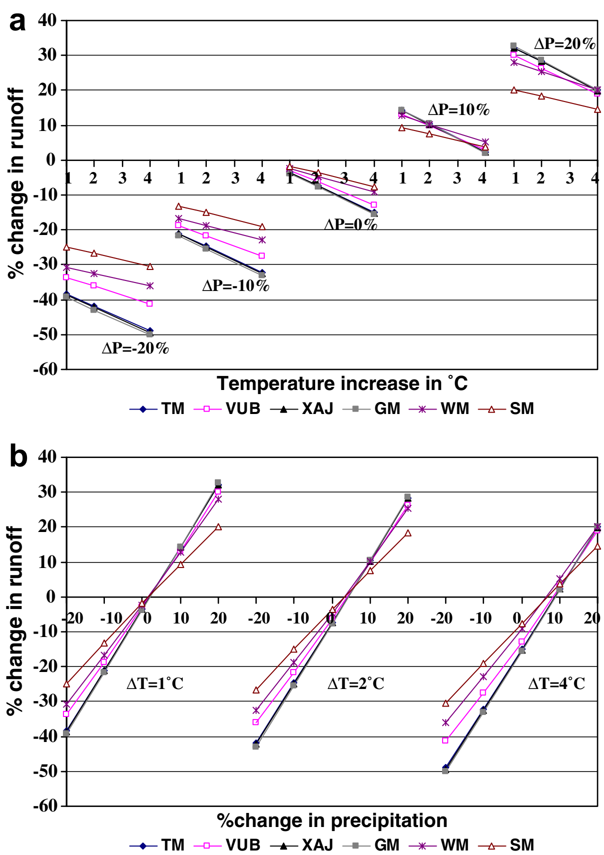

Figure 5 Comparison of mean annual changes in runoff in response to temperature increases for a given precipitation change (a) and comparison of mean annual changes in runoff in response to mean annual changes in precipitation for a given temperature change (b).

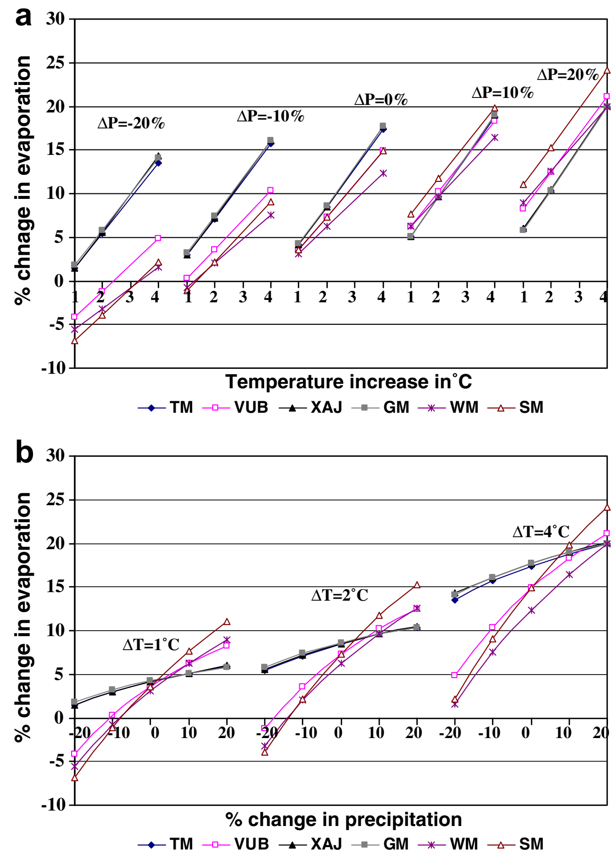

Figure 6 Comparison of mean annual changes in evapotranspiration in response to temperature increases for a given precipitation change (a) and comparison of mean annual changes in evapotranspiration in response to mean annual changes in precipitation for a given temperature change (b).

the increase in precipitation by the same amount; (3) the TM, XAJ and GM models behave similarly and produce more changes in runoff for a given climate change scenario, while the SM model produces the smallest changes. The runoff changes in response to precipitation changes for a given temperature change are shown in Fig. 5b. The slopes of the lines in each branch of lines reveal the changing rate of runoff in response to precipitation changes. The difference between the three branches represents the influence of different temperature change scenarios. Comparing Fig. 5a with Fig. 5b it can be seen that runoff changes are more sensitive to precipitation changes than to temperature.

Similar to Fig. 5, evapotranspiration changes in response to the climate change scenarios are shown in Fig. 6. This figure also shows that even on the annual level there is a wide range of differences between the evapotranspiration responses simulated by the six models when perturbed climate scenarios are used to drive the models. From Fig. 6a and b, two groups of model simulated evapotranspiration changes in response to climate change scenarios can be distinguished. The first group consists of three models, i.e., TM, XAJ and GM, which respond almost identically in all the climate change scenarios. The other models show different responses.

Each branch of lines in Fig. 6a represents the difference of evapotranspiration changes caused by precipitation changes. Comparing Fig. 6a with Fig. 5a it is seen that the effect of precipitation changes on evapotranspiration is less than on runoff. Comparing Fig. 6b with Fig. 5b one can see that the effect of temperature changes on evapotranspiration and runoff has a different sign. For example, when precipitation has no change, one, two and four degree increases in temperature cause an increase in evapotranspiration of about 3%, 6–8% and 12–17%, respectively.

The percent changes of mean annual soil moisture in response to the 15 climate change scenarios are shown in Fig. 7. In general, Fig. 7 shows that (1) the differences in soil moisture changes simulated by the VUB model are larger than those simulated by other five models when perturbed climate scenarios are used to drive the models. This is probably due to the fact that the VUB model is the only model under comparison that does not have an upper threshold limit for soil moisture. The soil moisture can change freely in the VUB model, while this is not the case in other five models. (2) The differences in soil moisture changes under alternative climates are smaller than the runoff changes in the five models (exclude VUB model), when an upper threshold limit is used for soil moisture simulation in the model equations.

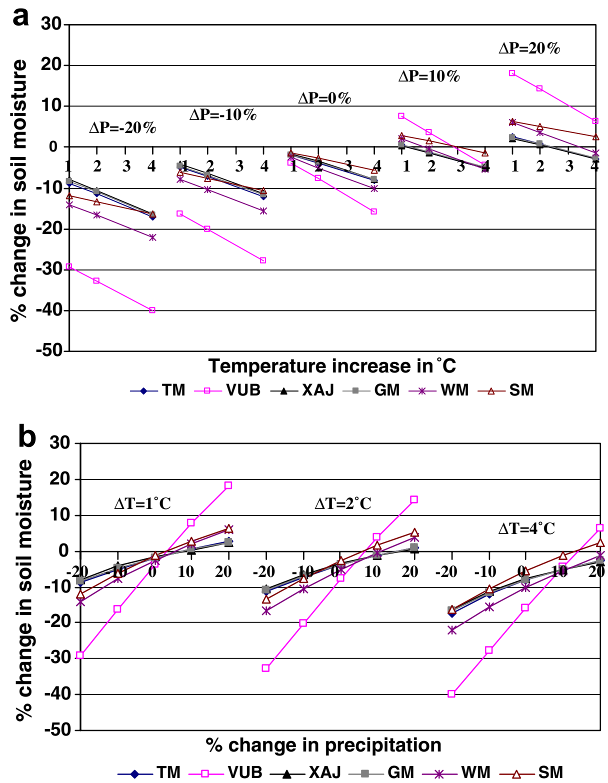

Figure 7 Comparison of mean annual changes in soil moisture in response to temperature increases for a given precipitation change (a) and comparison of mean annual changes in soil moisture in response to mean annual changes in precipitation for a given temperature change (b).

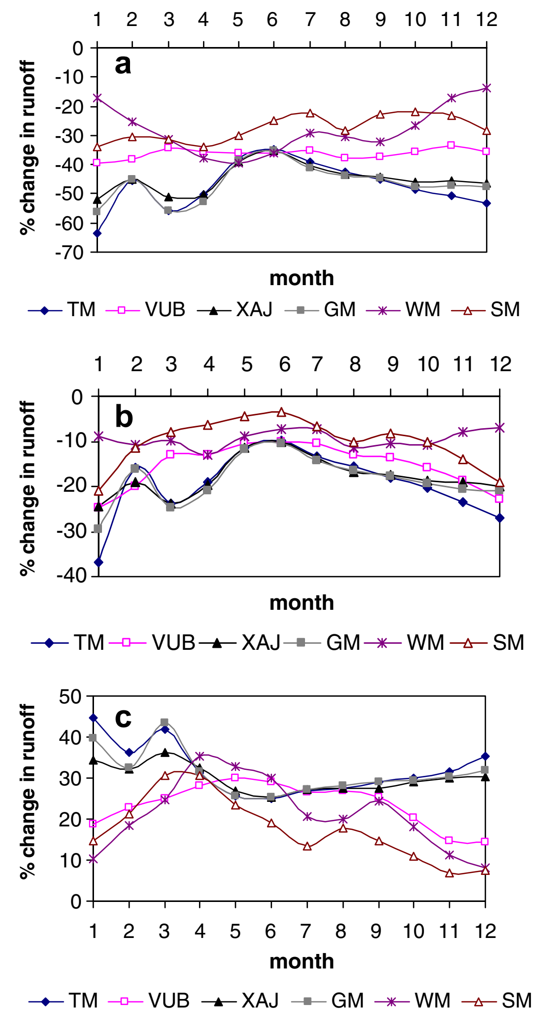

Figure 8 Comparison of mean monthly changes in runoff simulated by the six models for three climate change scenarios (a) ΔT = +2 °C and ΔP = − 20%, (b) ΔT = +2 °C and ΔP = 0%, and (c) ΔT = +2 °C and ΔP = 20%.

Mean monthly changes

To evaluate seasonal and inter-annual changes, differences in mean monthly runoff, actual evapotranspiration and soil moisture simulated by the six models with various climate change scenarios are compared. Due to the limit of the length of the paper and for illustrative purposes, changes in mean monthly runoff, actual evapotranspiration and soil moisture simulated by the six models for three climate change scenarios (i.e., combination of temperature increases by 2 °C (ΔT =+2 °C) and precipitation changes by ±20% and 0%) are plotted in Figs. 8–10, respectively.

Fig. 8 shows that (1) the differences in model predicted mean monthly runoff resulting from the use of different hydrological models are large. The figure shows a great diversity in the simulation results. (2) The larger differences in percent changes in runoff for winter months may be caused by smaller absolute values in runoff in winter months (see Fig. 2). (3) The TM, XAJ and GM models respond similarly in predicted changes in monthly runoff, while other three models show a similar pattern of seasonal variation in the predicted changes of monthly runoff for changing climate. (4) On average, when temperature increases by 2 °C the mean monthly runoff changes by −30 ~ −50%, −5 ~ −10%, and 10 ~ 30%, respectively for precipitation changes of −20%, 0% and 20%, depending on the model.

Fig. 9 shows that (1) the differences in model predicted mean monthly actual evapotranspiration resulting from the use of different hydrological models are also remarkable. (2) Two groups of evapotranspiration reaction behaviour with respect to climate changes can be distinguished. The first group of models (i.e., TM, XAJ and GM) react very similarly, while differences among other three models are considerable. (3) In summer months (rainy season in the region) the model-predicted actual evapotranspiration does not change significantly with the change in precipitation. A two degree increase in air temperature causes about a 10% increase in evapotranspiration in all three cases, meaning that actual evapotranspiration is energy controlled in the rainy season in the humid area.

It is seen from Fig. 10 that (1) the differences in model-predicted mean monthly soil moisture content resulting from the use of different hydrological models are also considerable. (2) Again, two groups of model reaction behaviour with respect to climate changes can be distinguished. The first group of models (i.e., TM, XAJ and GM) react very similarly, while differences among the other three models are considerable, especially the VUB model behaves very differently from other models. As discussed

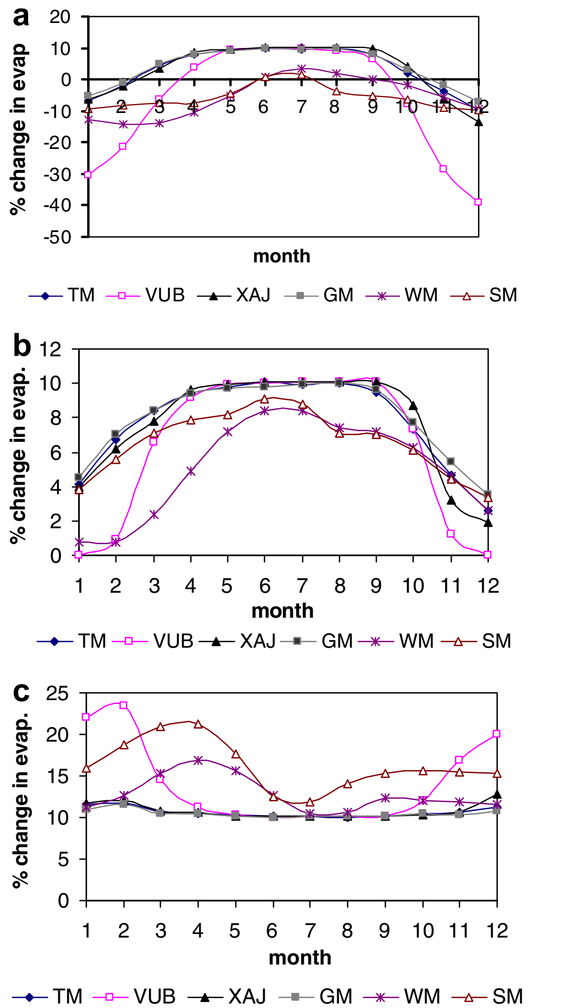

Figure 9 Comparison of mean monthly changes in actual evapotranspiration simulated by the six models for three climate change scenarios (a) $\Delta T = +2\,^{\circ}C$ and $\Delta P = -20\%$, (b) $\Delta T = +2\,^{\circ}C$ and $\Delta P = 0\%$, and (c) $\Delta T = +2\,^{\circ}C$ and $\Delta P = 20\%$.

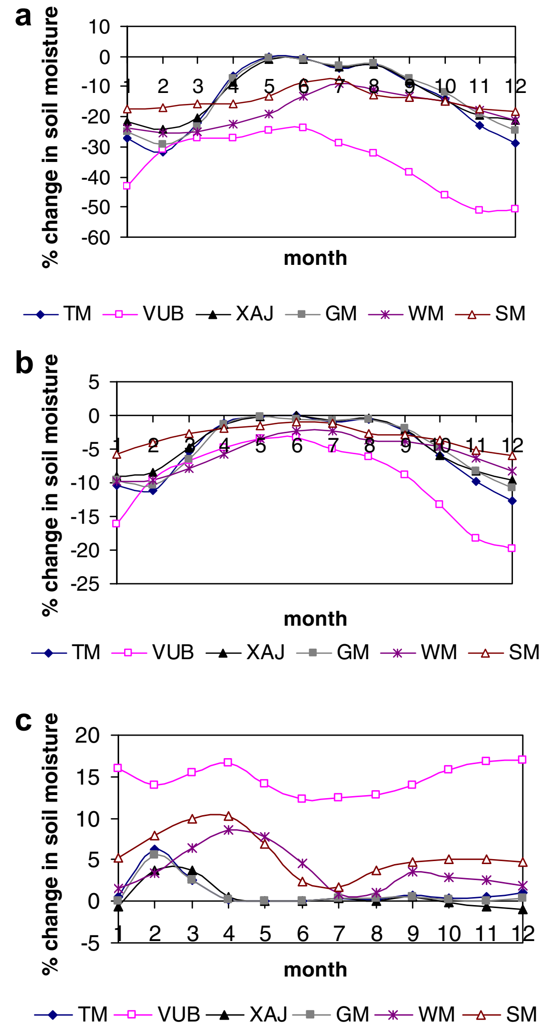

Figure 10 Comparison of mean monthly changes in soil moisture simulated by the six models for three climate change scenarios (a) $\Delta T = +2\,^{\circ}C$ and $\Delta P = -20\%$, (b) $\Delta T = +2\,^{\circ}C$ and $\Delta P = 0\%$, and (c) $\Delta T = +2\,^{\circ}C$ and $\Delta P = 20\%$.

in the last section, this is because the VUB model is the only one that does not have an upper threshold limit in the calculation of soil moisture.

Summary and conclusions

In this study, six monthly water balance models are used and their differences in reproducing historical water balance components and in predicting hydrological impacts of 15 perturbed climate change scenarios are compared. The study is performed in the Dongjiang catchment, a tributary of the Pearl River (Zhujiang) located in a subtropical humid region in southern China. The main focus of the study is to test how large differences one can expect when using different rainfall-runoff models to simulate hydrological response of climate changes as compared to their capacities in simulating historical water balance components.

This study shows that when using the class of models considered in this paper for hydrological simulation of changing/changed climate the following conclusions can be drawn.

- All the six tested models can reproduce almost equally well the historical runoff data series, while large differences exist in the model simulated soil moisture.
- Using alternative climates as input to the tested models, large differences exist in model predicted runoff, actual evapotranspiration and soil moisture. The differences depend on the climate scenarios, the season, and the hydrological variables under examination.
- Using an upper threshold limit in the soil moisture simulation by the models has a significant influence on the model simulated soil moisture under both historical and alternative climates. The model without a threshold in soil moisture simulation results in greater changes in model predicted soil moisture with respect to alternative climates than the models with a threshold soil moisture.

The results confirm the findings of the previous studies using fewer models than the present study. Boorman and Sefton (1997) compared two rainfall-runoff models and concluded that ''Clearly the two models do little to support each other in estimating the possible effects of a changed climate''. Panagoulia and Dimou (1997) also reported that significant differences exist in the predicted runoff and soil moisture with respect to alternative climates by using two different models, and the differences depend on the climate scenarios and on the seasons.

This study suggests that attention must be paid when using existing hydrological models for simulating hydrological responses of climate changes. Future water resources scenarios predicted by any particular hydrological model represent only the results of that model. More studies using different hydrological models on different catchments need to be carried out in order to provide more general conclusions.

Acknowledgements

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Appendix A. Model description

A.1. Thornthwaite–Mather model (TM)

The Thornthwaite–Mather model (called TM model hereafter) was developed by Thornthwaite and Mather (1955) and a detailed study of the model was done by Alley (1984). It is a model with two storages (Fig. A1): 'soil moisture index' and 'water surplus'. The model has two parameters: soil moisture capacity and storage constant. In this model, the soil is assumed to have a maximum soil moisture capacity, S_{max}. Moisture is either added to or subtracted from the soil, depending on whether the precipitation of month t, $P(t)$, is greater or less than the potential evapotranspiration for the month, PET(t). When $P(t) \geqslant \mathrm{PET}(t)$, the soil moisture content is updated by $S(t) = \min[P(t) - \mathrm{PET}(t) + S(t), S_{max}]$, and the excess precipitation is assumed to contribute to water surplus, $\Delta Q = [P(t) - \mathrm{PET}(t)] + S(t-1) - S_{max}]$. When $P(t) < \mathrm{PET}(t)$, the soil moisture storage is described by $S(t) = S(t-1)e^{-\frac{\mathrm{PET}(t)-P(t)}{S_{max}}}$ and $\Delta Q = 0$. Actual evapotranspiration is computed by $E_a(t) = \mathrm{PET}(t)$, when $P(t) \geqslant \mathrm{PET}(t)$. Otherwise, $E_a(t) = P(t) + S(t-1)\left|e^{-\frac{\mathrm{PET}(t)-P(t)}{S_{max}}} - 1\right|$. Streamflow is derived from the water surplus. This model assumes that a fraction λ of the water surplus remains in the soil and recharges the groundwater storage. Thus runoff for month t is $R(t) = (1-\lambda)[Q(t-1) + \Delta Q]$, and the water surplus at the end of the month is updated by $Q(t) = \lambda[Q(t-1) + \Delta Q]$. Parameter λ varies with the depth and texture of the soil, size and physiography of the basin, and characteristics of groundwater system. Parameters that need to be calibrated are S_{max} and λ.

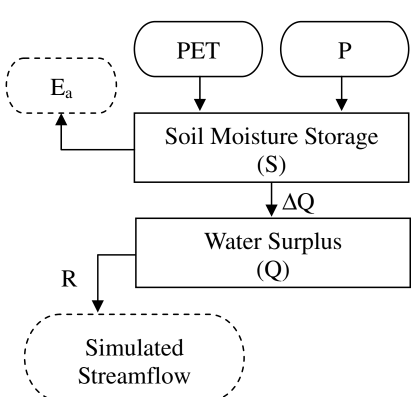

Figure A1 Schematic representation of Thornthwaite–Mather model (TM).

A.2. Belgium model (VUB)

Vandewiele et al. (1992) proposed a series of monthly water balance models on a basin scale. One of the models, referred to here as VUB model (Fig. A2), is selected for use in the study. The model is based on the water balance equation $S(t) = S(t-1) + P(t) - E_a(t) - R(t)$. $S(t)$ and $S(t-1)$ represent the states of the soil moisture storage at the end and beginning of month t, respectively. $P(t)$, $E_a(t)$, $R(t)$ are precipitation, actual evapotranspiration, and discharge during month t, respectively. Actual evapotranspiration is computed as $E_a(t) = \min\left[W(t)(1 - e^{-\frac{\mathrm{PET}(t)}{a_1}}), \mathrm{PET}(t)\right]$, where a_1 is a non-negative parameter and $W(t) = P(t) + S(t-1)$ is the available water. Monthly discharge is distinguished between slow runoff $R_s(t) = a_2[S(t-1)^+]^{0.5}$ and quick runoff $R_f(t) = a_3\ (S(t-1)^+)^{0.5}N(t)$, where $N(t) = P(t) - \mathrm{PET}(t)\left[1 - e^{-\frac{P(t)}{\max(\mathrm{PET}(t),1)}}\right]$ is defined as the effective precipitation. The total runoff is $R(t) = R_s\ (t) + R_f(t)$. Parameters that need to be calibrated are a_1, a_2, and a_3.

A.3. The monthly xinanjiang model (XAJ)

The Xinanjiang model was originally developed on a hourly time scale for flood forecasting or on a daily time scale for continuous hydrological simulation (Zhao, 1992). Various

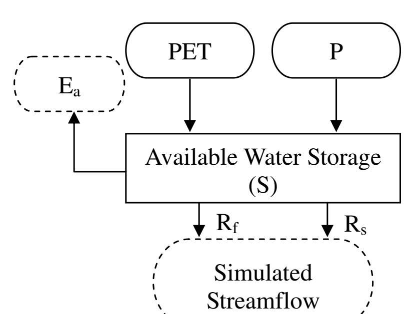

Figure A2 Schematic representation of Belgium model (VUB).

simplified model versions for monthly simulation were reported (e.g., Hao and Su, 2000) according to the characteristics of monthly runoff hydrograph for basin hydrological modeling and water resources assessment (Fig. A3).

In the monthly Xinanjiang model, the soil moisture storage during month t, $W(t)$, is divided into three layers: $WU(t)$ in the upper layer, $WL(t)$ in the lower layer, and $WD(t)$ in the deep layer, of which the upper threshold limit values are WUM, WLM and WDM, respectively. Soil moisture storage is depleted and replenished, respectively, through evapotranspiration and precipitation.

Actual evapotranspiration is related to both potential evapotranspiration and soil moisture status. In the upper layer, evapotranspiration of month t, $EU(t)$, takes place at the rate of potential evapotranspiration. Once the moisture content in the upper layer has been depleted, evapotranspiration proceeds to the lower layer, $EL(t)$, depending on the ratio of the moisture content to the storage capacity. When the lower layer storage falls below some pre-set fraction of WLM, evapotranspiration is assumed to continue at a rate $ED(t)$. The actual evapotranspiration is the total of the evapotranspiration of the three layers.

Precipitation first satisfies the demand for evapotranspiration and runoff generation depends on the difference between precipitation and potential evapotranspiration. If the difference is positive, runoff generates; otherwise, there is no runoff production. It is proposed that the runoff production is based on the partial area concept by considering the non-uniform distribution of soil moisture storage capacity over the basin. The remainder of runoff production becomes an addition to the groundwater storage. The groundwater storage contributes to the slow runoff. The simulation procedures of the monthly Xinanjiang model are outlined in Fig. A3. The model has 7 parameters, of which 4 parameters are sensitive ones and need to be calibrated.

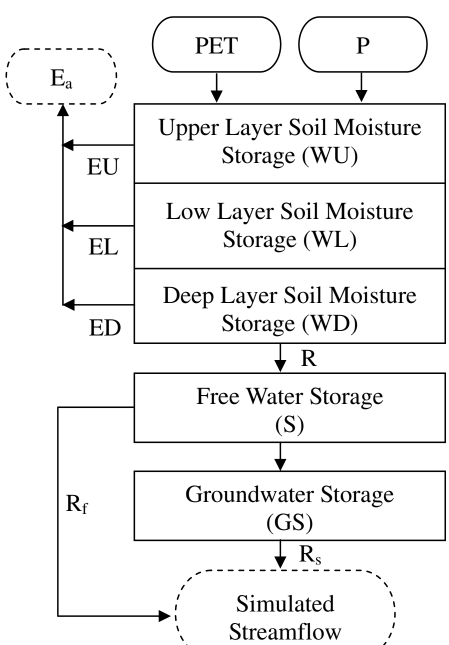

Figure A3 Schematic representation of the monthly Xinanjiang Model (XAJ).

A.4. Guo model (GM)

Guo (1992) developed a five-parameter monthly water balance model. The basic structure of this model is presented in Fig. A4. At a monthly time step, the model divides runoff into surface runoff, interflow, and groundwater flow. Rainfall first satisfies the demand for evapotranspiration and replenishes soil moisture, and the remaining amount would become surface runoff and interflow.

If precipitation is greater than potential evapotranspiration, then the soil moisture is calculated as $S(t) = S(t-1) + P(t) - \mathrm{PET}(t)$, otherwise $S(t) = S(t-1)\,e^{-\frac{\mathrm{PET}(t)-P(t)}{S_{\max}}}$. When precipitation exceeds potential evapotranspiration and soil moisture storage attains its capacity, the excess water is equal to $S(t)$ minus $S_{\max}$, a part of which contributes to surface runoff $R_s(t) = c[S(t) - S_{\max}]$, and c is the surface runoff coefficient. The remaining amount of excess water is determined from $WS(t) = (1-c)[S(t) - S_{\max}]$, and the interflow is $R_i(t) = k_1 WS(t)$. Where $k_1$is a watershed lag coefficient. Groundwater is assumed to behave like a reservoir receiving a part of the excess water and discharging at a specified rate to the river with a time lag of one month. The groundwater flow is calculated by $R_g(t) = k_2 G(t-1)$, where k_2 is a groundwater reservoir coefficient, and $G(t-1)$ is the groundwater storage at the beginning of the month. At the end of each month t, groundwater storage is updated by $G(t) = G(t-1) + (1-k_1)WS(t) - k_2 G(t-1)$ for $WS(t) \geqslant 0$, and $G(t) = G(t-1) - k_2 G(t-1)$, for $WS(t) < 0$. The total monthly runoff $TR(t)$ is the sum of surface runoff, interflow, and groundwater flow $TR(t) = R_s(t) + R_i(t) + R_g(t)$. Parameters that need to be calibrated are maximum soil moisture storage, $S_{\max}$, surface runoff coefficient, c, watershed lag coefficient, k_1 and groundwater reservoir coefficient, k_2.

A.5. WatBal model (WM)

The WatBal model was originally developed for Colorado Subalpine watersheds (Leaf and Brink, 1973). Sequentially it was modified for assessing the hydrological impact of climate change. The conceptualization of the WatBal model is shown in Fig. A5. The uniqueness of the WatBal model is the use of continuous functions of relative storage to represent surface outflow, sub-surface outflow, and evapo-

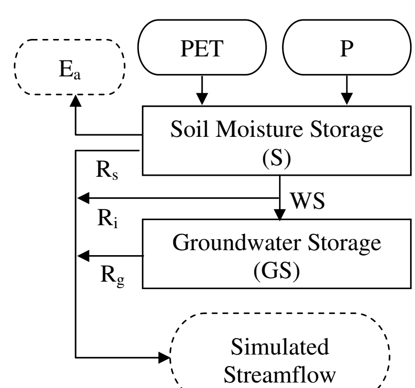

Figure A4 Schematic representation of Guo model (GM).

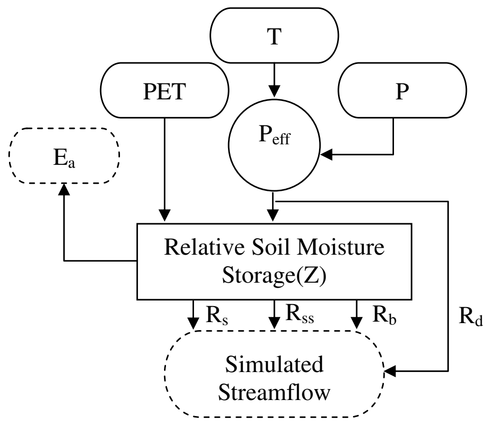

Figure A5 Schematic representation of WatBal model (WM).

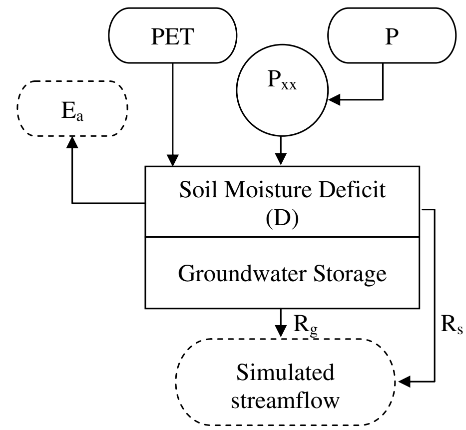

Figure A6 Schematic representation of Schaake model (SM).

transpiration. Given direct runoff $R_d(t) = \beta P_{eff}(t)$, the water balance is written as $S_{max}\frac{dz}{dt} = P_{eff}(t)(1-\beta) - R_s(z,t) - R_{ss}(z,t) - E_a(z,t) - R_b$, where t is time, β is a direct runoff coefficient, S_{max} is the maximum storage capacity, $P_{eff}(t)$ is the effective precipitation, $R_s(z,t)$ is the surface runoff, $R_{ss}(z,t)$ is the sub-surface runoff, $E_a(z,t)$ is the evapotranspiration, R_b is the baseflow, $z = S(t)/S_{max}$ is the relative soil moisture storage, and $S(t)$ is the soil moisture storage. This differential equation can be solved by using a predictor-corrector method.

Evapotranspiration is a function of potential evapotranspiration and the relative storage, $E_a(z,t) = \text{PET}(t)\frac{5z-2z^2}{3}$. Surface runoff and subsurface runoff are calculated as functions of relative storage and effective precipitation, respectively. The model has four parameters that need calibration, i.e., maximum storage capacity, S_{max}, direct runoff coefficient, β, a parameter related to surface runoff, ε, and a parameter related to sub-surface runoff, α.

A.6. Schaake model (SM)

Schaake and Liu (1989) developed a simple water balance model for assessing the impact of climate change. Schaake (1990) improved groundwater algorithms of the linear model using a nonlinear reservoir. The enhanced model, referred to as the SM model, has the ability to simulate runoff over a range of climate conditions. The uniqueness of the model is to introduce soil moisture deficit in the expression of runoff and evapotranspiration. Runoff is divided into surface runoff and groundwater flow. The schematic representation of the Schaake model is given in Fig. A6.

In the SM model, actual evapotranspiration is assumed to occur at a potential rate when the storage deficit is zero, whereas actual evapotranspiration is zero in the case when the storage deficit reaches the maximum limit. In the intermediate case, actual evapotranspiration $E_a(t)$ in month t is calculated as $E_a(t) = \text{PET}(t)\frac{D_{max}-D(t)}{D_{max}}$, and D_{max} is the maximum limit of soil moisture storage deficit, and $D(t)$ is the current deficit.

To compute surface flow, the effective precipitation, $P_{xx}(t)$, is defined as $P_{xx}(t) = P(t) - \Theta E_a(t) - zD(t)$, Θ is a parameter representing the proportion of $E_a(t)$ that must be satisfied from precipitation in the current month before runoff or infiltration can occur; z is a parameter that controls infiltration of precipitation through the surface of the earth. If $P_{xx}(t)$ is positive, surface runoff is calculated as $R_s(t) = P_{xx}(t)\frac{P_{xx}(t)}{P_{xx}(t)+D_{max}}$. Groundwater runoff is assumed to vary with deficit $D(t)$. It is assumed that the groundwater table rises to streams and groundwater runoff generates when $D(t)$ is less than a certain value. The equation used to compute groundwater runoff is $R_g(t) = kk[G_{max} - D(t)]$, where kk is a parameter, and G_{max} is a threshold value. If $D(t)$ exceeds G_{max}, $R_g(t)$ is zero and streams would percolate to groundwater. However, no attempt is made in the model to account for such losses from streams to groundwater. The storage deficit at the end of the month is computed by mass balance as $D(t+1) = D(t) - P_{xx}(t) + E_a\ (t) + R_s(t)$. Parameters that need to be calibrated are maximum limit of soil moisture storage deficit, D_{max}, infiltration parameter, z, proportion of $E_a(t)$ that must be satisfied before runoff or infiltration can occur, Θ, groundwater parameter, kk, and a threshold parameter, G_{max}.

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