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Climate information based streamflow and rainfall forecasts for Huai River Basin using Hierarchical Bayesian Modeling

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Abstract

A Hierarchical Bayesian model for forecasting regional summer rainfall and streamflow season-ahead using exogenous climate variables for East Central China is presented. The model provides estimates of the posterior forecasted probability distribution for 12 rainfall and 2 streamflow stations considering parameter uncertainty, and cross-site correlation. The model has a multilevel structure with regression coefficients modeled from a common multivariate normal distribution results in partial-pooling of information across multiple stations and better representation of parameter and posterior distribution uncertainty. Covariance structure of the residuals across stations is explicitly modeled. Model performance is tested under leave-10-out cross-validation. Frequentist and Bayesian performance metrics used include Receiver Operating Characteristic, Reduction of Error, Coefficient of Efficiency, Rank Probability Skill Scores, and coverage by posterior credible intervals. The ability of the model to reliably forecast regional summer rainfall and streamflow season-ahead offers potential for developing adaptive water risk management strategies.

1 Introduction

The Huai River Basin (Fig. 1), located between the Yangtze and Yellow River basins is the most densely inhabited river basin and the main cropping area in China. The total drainage area of 270 000 km² is divided into the Huai River Catchment (190 000 km²) and the Yishusi River Catchment (80 000 km²) by a paleo-channel of Yellow River. The mean annual precipitation for the Huai River Basin is approximately 900 mm, of which 50–75 % occurs during the summer monsoon season. There are 36 large reservoirs in the basin primarily designed for water supply and flood control (Zhang et al., 2012). Natural variations in climate in conjunction with increasing population demands are causing severe water stress in the region. Moreover, the region is susceptible to droughts and floods with a recurrent frequency of four years on average (Yan et al.,

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2013; Cheng et al., 2012). Such variations and epochal changes are often related to large scale climatic patterns that may have similar inter-annual fluctuations. Such teleconnections are explored and formalized into a predictive model in a Bayesian framework so that applications for water, food and energy management can be informed ahead of the season. There is considerable interest today in strategies for climate risk mitigation and adaptation to a variable climate. The seasonal forecasting approach provides a way for institutional planning and action in this context.

Interest in the development and application of long-lead hydrologic forecasts has grown over the last decade primarily because of the improved monitoring of sea-surface temperature (SST) in the tropical Pacific and advances in experimental climate forecasts from General

Circulation Models (GCMs). Since GCM-predicted fields are usually available at large spatial scales, one needs to apply either dynamical or statistical downscaling to develop regional hydrologic forecasts (Robertson et al., 2004; Gangopadhyay et al., 2005). Alternately, one can develop a low-dimensional statistical model by relating the observed rainfall or streamflow to identified climatic precursors (e.g., El Nino Southern Oscillation (ENSO) indices) for the given site (Souza and Lall, 2003). A key aspect in developing such dynamic or statistical models is the ability to accurately represent the uncertainties, both at the model representation level and at the parameter estimation level. Developing statistical schemes that can address simultaneous prediction at multiple sites and for multiple variables while addressing their correlation structure is often a challenge.

Hierarchical Bayesian methods provide the opportunity to explicitly quantify the parameter uncertainty through each estimation stage using appropriate conditional and prior distributions. This allows a better representation of model and parameter uncertainties. Recently, Devineni et al. (2013) presented a Hierarchical Bayesian Regression strategy for estimating streamflow at multiple locations using various model structures to pool information across multiple sites to an appropriate degree such that the features that are common to the site regression and those that vary across sites can be identi-

fied for an overall reduction in parameter uncertainty while preserving the structure in errors across the stations. Here, a similar Hierarchical Bayesian approach is developed for regional rainfall and streamflow forecasts using appropriate climate indicators that could be derived from GCMs or observed climate fields. The application focuses on the upper and middle regions of the Huai River Basin which are of interest for local management, and may have similar climatic forcing. Section 2 provides a brief description of the study area, data sources and the climate predictor identification. The Hierarchical Bayesian Regression model is presented in Sect. 3. In Sect. 4, the cross-validated results are presented. A summary is finally presented.

1.1 Data description

1.2 Streamflow and rainfall data

We used streamflow data from two stations and rainfall data from 12 stations in this study to develop regional hydrologic forecasts. The streamflow data are from the Bengbu (117.38° E, 32.56° N) and the Lutaizi (116.79° E, 32.57° N) hydrological stations (shown as red dots in Fig. 1). Twelve rainfall stations with at least 50 yr of data were selected in the contributing section of the basin (shown as filled triangles in Fig. 1). The details of the hydrologic stations including the number of years of data records are shown in Table 1. Preliminary analysis of the seasonality of streamflow and area averaged rainfall show that more than 50 % of the annual rainfall and streamflow occurs in June-July-August (JJA) (Fig. 2). Consequently, a prediction of the summer monsoon rainfall and streamflow in June or earlier is of interest.

1.3 Climate teleconnection and predictor identification

Xu et al. (2007) and Kwon et al. (2009) developed season ahead streamflow forecasts for the Yangtze River on the Three Gorges Dam using exogenous climate indices from eastern Indian Ocean and western Pacific Ocean. Recently, Liu et al. (2013) and Lin-

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derhorm et al. (2013) investigated the relation between East Asian monsoon rainfall and north Atlantic Sea Surface Temperature conditions using observations and paleo-reconstructed records. So far little work has been done for climate informed hydrologic prediction for the Huai River Basin. To identify predictors that influence the regional hydroclimate in the basin during JJA season, we consider SST anomaly conditions during February–March–April (FMA, 3 months lag) and October–November–December (OND, 6 months lag) obtained from the Hadley Center SST dataset (HADSST2) (Rayner et al., 2006). Figure 3 shows the Spearman’s rank correlation between the observed streamflow during JJA at the Bengbu hydrologic station and the pre-season SST conditions. The 3-month lag correlation (i.e. JJA streamflow with FMA SSTa) and the 6-month lag correlation (i.e. JJA streamflow with OND SSTa) are shown in Fig. 3a and b respectively. From Fig. 3a, we see that the SST1 region (155° E–175° E and 40° N–50° N) influences the summer streamflow in the Huai River Basin. This region is associated with the Kuroshio Current which is the west side of the clockwise North Pacific Ocean gyre. This phenomenon was previously identified by Geng (1997). The warmer SSTs in the Kuroshio Current region and the mid latitude central pacific in late spring can generate a large scale circulation pattern over the Asia-Pacific region that is favorable for precipitation in the North China region. Hence, warm conditions in this region result in above-normal inflow conditions in the Huai River. Similarly, from Fig. 3b, we can see that 6-month prior conditions in the North Atlantic Ocean identified as SST2 (15° W–5° E and 35° N–55° N) influence the summer flows in the basin. This is in line with an earlier finding (Gu et al., 2009a) who showed that the East Asian Summer Monsoons are strongly related to triple mode of the North Atlantic SST anomalies in the preceding winter that typically enhance the stationary wave-train propagating from west Eurasia to East Asia. The correlations with SST1 and SST2 are statistically significant at the 95 % level.

In addition to the SST anomaly conditions, we also considered the 4-month lagged (February–March–April) North Atlantic Oscillations (NAO) (Hurrell et al., 2003), Summer North Atlantic Oscillation (SNAO) (Folland et al., 2009) and 6-month lagged

(November–December–January) Atlantic Multi-decadal Oscillation (AMO) (Knight et al., 2006) as candidate predictors. Gu et al. (2009b) showed that the January and March NAO modulates the summer rainfall patterns over China. Following Linderholm et al. (2011) we selected the SNAO as one of the predictors as it influences the storm tracks over China with wetter than normal conditions during positive SNAO phase and drier than normal conditions during negative phase. Similarly, AMO has been shown as an important covariate for climate in central Asia with positive AMO phase leading to strong southeast summer monsoons and late retrieval (Lu et al., 2006). We also checked the relationship between ENSO indices, Pacific Decadal Oscillation and the Northern hemisphere snow cover with summer streamflow and rainfall in the basin. There was no statistically significant correlation between these covariates and streamflow or rainfall in the region. Table 2 summarizes the correlations between the two streamflow stations, area averaged rainfall and the climate predictors selected for the study.

2 Hierarchical Bayesian Model: methodology

In this paper, we follow an approach similar to that used by Devineni et al. (2013) for tree ring based streamflow reconstruction, but for multivariate seasonal forecasts for regional streamflow and rainfall. The basic idea is that a particular climate predictor may inform the rainfall or streamflow anomaly at each of the sites in the region in a similar way. If the response were exactly the same, predicting the average of the station values or pooling all the data into the same regression would be effective since that would reduce the uncertainty associated with parameter estimation. However, the response across the rainfall and the streamflow stations may vary systematically due to local conditions or averaging scale (e.g., for a large river basin vs small or point rainfall). The hierarchical model can be used for partial pooling of this common information, by considering multiple levels of modeling. The individual regression coefficients for each site on each climate predictor are estimated at the first level. The second level estimates

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the average regression coefficient across sites and its variance, thus allowing variation in the response across sites, but also its potential shrinkage to an appropriate degree through an estimation of the variance. If the variance estimated is large, then the model tends towards a model that would be formed if each site was regressed independently on the predictor. If the variance is small, then the model tends towards a fully pooled regression model, and the responses are deemed homogeneous. Partial-pooling reduces the equivalent number of independent parameters, resulting in lower uncertainty in parameters estimates, and therefore reduced uncertainty in the final forecasts. The general modeling framework is presented as follows.

Given that the station rainfall at location i for year t is R_{it} , and station streamflow at location i for year t is S_{it} , we form $\mathbf{Y} = \{\mathbf{R}, \mathbf{S}\}$. The streamflow and rainfall data \mathbf{Y} are assumed to come from a distribution (process model) with probability density function $f(\mathbf{Y}|\theta)$, where θ is a parameter vector. In the application presented here, we consider that $\log(\mathbf{Y})$ is normally distributed. This assumption was checked using a Shapiro-Wilks test. Where a linear model is considered for $\log(\mathbf{Y})$ in terms of a set of climate predictors, the regression coefficients are interpretable as the fractional change in \mathbf{Y} given the corresponding predictor. This allows a consideration of partial pooling of response to a particular climate predictor across rainfall and streamflow data series that may have a disparate range or scale of values. The first level of the model considers that at each site i , $\log(Y_i)$ is described by a Normal distribution with time varying mean μ_{it} that is informed by a regression on the 5 predictors \mathbf{X}_t with intercept α_i and a (5×14) regression coefficient matrix $\boldsymbol{\beta}$. The “errors” from the regression model are considered to be spatially correlated with a (14×14) covariance matrix $\boldsymbol{\Sigma}$. The second level of the model considers that the regression coefficient matrix $\boldsymbol{\beta}$ can be modeled as coming from a multivariate Normal distribution with a (5×1) mean vector $\boldsymbol{\mu}_\beta$ which represents the average regression coefficient across the 14 sites for each predictor, and a (5×5) covariance matrix $\boldsymbol{\Sigma}_\beta$ that takes into account the correlation across the predictors associated with the common effect across the 14 sites. The $\boldsymbol{\mu}_\beta$ and $\boldsymbol{\Sigma}_\beta$ are called hyper-parameters. The model and the priors associated with the parameters and

the hyper-parameters are presented below:

$$\text{Level 1 : } \log(\mathbf{Y}_t) \sim N(\boldsymbol{\mu}_t, \boldsymbol{\Sigma}) \dots \quad (1)$$

$$\mu_{it} = \alpha_i + \mathbf{X}_i \boldsymbol{\beta} \dots \quad (2)$$

$$\text{Level 2 : } \boldsymbol{\beta} \sim \text{MVN}(\boldsymbol{\mu}_\beta, \boldsymbol{\Sigma}_\beta) \dots \quad (3)$$

With priors modeled as:

$$\alpha_i \sim N(0, 10\,000)$$

$$\boldsymbol{\mu}_\beta \sim N(0, 10\,000)$$

$$\boldsymbol{\Sigma}_\beta \sim \text{Inv - Wishart}_{\nu_0}(\boldsymbol{\Lambda}_0)$$

$$\boldsymbol{\Sigma} \sim \text{Inv - Wishart}_{\nu_1}(\boldsymbol{\Lambda}_1) \dots \quad (4)$$

The prior for the covariance matrix $\boldsymbol{\Sigma}_\beta$ is taken to be the Inverse-Wishart distribution with a scale matrix $\boldsymbol{\Lambda}_0$ and ν_0 degrees of freedom. In our applications, the scale matrices $\boldsymbol{\Lambda}_0$ and $\boldsymbol{\Lambda}_1$ were specified as an identity matrix (**I**) and the degrees of freedom ν_0 and ν_1 were set to one more than the dimension of the matrix (i.e. the total number of predictors, five for $\boldsymbol{\Sigma}_\beta$ and total number stations, 14 for $\boldsymbol{\Sigma}$) to induce a uniform prior distribution on the variance (Gelman and Hill, 2007). This choice of priors was made for computational convenience and represents a simpler model than could be formulated if all parameter covariance were to be modeled. The joint posterior distribution $p(\boldsymbol{\theta} | \text{data})$, of the complete parameter vector $\boldsymbol{\theta}$ is derived by combining the prior distributions and

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the likelihood functions. The parameters θ are estimated using WinBUGS (Speigelhalter et al., 1996) which employs the Gibbs sampler, a Markov Chain Monte Carlo (MCMC) method for simulating the posterior probability distribution of the parameters conditional on the current choice of parameters and the data. A discussion on such model constructs and their comparison to a no pooling model that estimates independent regressions across sites and to a full pooling model that ignores the cross-site variations in response is presented in Devineni et al. (2013). Several hydrologic applications using Bayesian model constructs have also been developed and demonstrated in Lima and Lall (2009, 2010), and Kwon et al. (2008). Renard et al. (2013) present a useful tutorial and examples of related Bayesian models for hydroclimatic applications.

3 Results

The posterior distribution of the regression coefficients (β) for each climate predictor and the mean of the vector of regression coefficients across sites (μ_β) from the joint normal distribution, is shown in Fig. 4 through boxplots of the values simulated from the posterior density functions. All the predictors except NAO have positive coefficients for the mean response across sites. Note that the spread on each μ_β covers the median of the 14 corresponding β s, as would be expected from partial pooling. With the exception of the AMO, the β for streamflow tend to be higher and distinct from those for rainfall. This is expected given that the streamflow represents a spatial averaging of the rainfall process. One could consider modeling these as separate groups to be pooled. However, given that we have only 2 streamflow stations, pooling across them would not provide much improvement. Modeling them together provides a larger sample size (14) for the estimation of the coefficients of the Level 2 model, and leads to a higher spread in the posterior distribution of μ_β than would result if we modeled the rainfall and streamflow stations in separate groups. An averaging of a larger number of rainfall stations with streamflow stations that individually represent a spatial averaging of rainfall is attractive from a conceptual perspective to regularize or reduce the uncer-

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tainty in the estimates of the response for the noisier rainfall stations. The simulations of the covariance matrix across predictors, Σ_{β} (results not shown), have non-zero off-diagonal elements. If two predictors are highly correlated then their regression coefficients cannot be uniquely identified through classical or Bayesian regression. However, their mean and covariance can be estimated and simulations of the $\log(Y)$ generated from the Bayesian model would be based on this covariance across the associated regression coefficients. Hence, an appropriate range of $\log(Y)$ values will be generated for each prediction, even though the individual β are not uniquely identified due to predictor correlation.

The posterior probability distributions of the forecasts from the model for the streamflow at Bengbu station and rainfall at Shouxian station during the period 1996–2010 are shown as boxplots in Fig. 5a and b respectively. While the Bayesian model is developed using all the data, the forecasts are shown for the last 15 yr to make a cleaner figure. Subsequent performance metrics are evaluated under cross-validation. The plots show streamflow and rainfall values as the percentage difference each year from their long term average ($1696 \text{ m}^3 \text{ s}^{-1}$ for the flow at Bengbu station and 455 mm for rainfall at Shouxian). We see that the directional indication of the forecast is generally quite accurate while the uncertainty varies from year to year. From a forecast utility point of view, knowing this uncertainty is useful since it can be used for developing probability based risk management models for optimizing reservoir operations or agricultural decision models. To provide insight as to how this could be approached, we present, in Fig. 6 the Receiver/Relative Operating Characteristic (ROC) plot (Mason, 1982) considering the decile categorical thresholds on the forecast posterior probability distribution for each of the 14 forecasts. Forecasts with better discrimination from random chance typically exhibit ROC curve approaching the upper-left corner of the diagram as opposed to the 45° diagonal lines where the forecast has little ability to discriminate from a 50–50 probability that occurs by chance. From Fig. 6, we see that the ROC curve for all the 14 stations is well beyond the diagonal line and approaches the left corner indicating that the forecasts exhibit hit rates higher than the false alarm rate and are

well calibrated to predict anomalous events using exogenous climate precursors. In reality a decision maker could prescribe their own thresholds of interest and evaluate the consequences of the forecast relative to the uncertainty and the threshold prescribed, as part of the decision process.

5 The correlation of the forecasts across the stations is important to maintain for hydrologic applications. Here, it is estimated from the posterior distribution of the streamflow and rainfall and compared to the observed cross-site correlation. The results for two stations (station1: streamflow from Bengbu station, station 11: rainfall from Shouxian station) are shown in Fig. 5c and d respectively. The boxplots in Fig. 5c and d present
10 the posterior probability distribution of the correlations for these station with each of the 13 other stations. The observed correlation for each station is shown as a triangle.

Thus far, we develop the above Bayesian model using all the data in the record and evaluate the performance probabilistically with respect to historical data. Although, one could argue that in a Bayesian context, cross validation is not required, cross-validation statistics computed over different blocks of data can reveal how well the model can
15 perform in truly out of sample predictions recognizing that different climate epochs may lead to different model fits and performance. We present results from an m -fold cross validation here. A sample is formed by leaving out m randomly selected data points from the observational data set for validation and the Bayesian model is developed
20 using the remaining $(n - m)$ observations. This process is repeated several times to obtain an ensemble of validation metrics resulting from each randomly selected model. In the applications presented m was 10, n was 50, and 30 sample models were fit. We used three traditional performance metrics, Reduction of Error (RE) and Coefficient of Efficiency (CE) and the Rank Probability Skill Score (RPSS), as measures of
25 model performance to compare the forecasted posterior mean and the distribution of the streamflow and rainfall estimates with the actual streamflow and rainfall data.

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The Reduction of Error (RE) ranges from $-\infty$ to $+1$ and is similar to the R^2 statistic (Lorenz 1956; Fritts 1976).

$$RE = 1.0 - \frac{\sum_{t=1}^n (O_t - S_t)^2}{\sum_{t=1}^n (O_t - \bar{o}_c)^2} \quad (5)$$

In Eq. (5), O_t and S_t are the observed and the predicted posterior mean of the streamflow (transformed back to real space by taking anti-logs) in year t of the validation period and \bar{o}_c is the mean of the observational data in the calibration period. $RE > 0$ indicates that the simulated streamflow contains useful information not contained in the calibration period. Similarly $RE < 0$ indicates that the simulations are poorer than climatology, i.e. the simulations are not better than the mean flows in the calibration period. The coefficient of efficiency (CE) is defined as

$$CE = 1.0 - \frac{\sum_{t=1}^n (O_t - S_t)^2}{\sum_{t=1}^n (O_t - \bar{o}_v)^2} \quad (6)$$

In Eq. (6), O_t and S_t are the observed and the predicted posterior mean of the streamflow in year t of the validation period and \bar{o}_v is the mean of the observational data in the validation period. $CE < 0$ indicates that the simulations are poorer than validation climatology, i.e. the simulations are not better than the mean flows in the validation period. CE is similar to RE, but used as a measure to evaluate the model under the validation period, it is a more rigorous metric.

In addition to RE and CE that measure the error in predicting the conditional mean, we also verified the RPSS to quantify the error in estimating the entire probability distribution of the forecast (Wilks, 2011; Candille and Talagrand, 2005; Gangopadhyay et

al., 2005). The RPSS is based on the ranked probability score (RPS) computed for each forecast and observation pair at each station in each year:

$$RPS = \sum_{m=1}^n (S_m - O_m)^2 \quad (7)$$

Where S_m is the cumulative probability of the forecast for category m and O_m is the cumulative probability of the observation for category m . This is implemented as follows.

First, the observed time series is used to distinguish 3 possible categories for forecasts of streamflow and rainfall, naturally taking 33, 33 and 33 %. These categories are determined separately for each year and stations in the basin. Next, for each forecast-observation pair, the number of ensemble members forecast in each category is determined (out of 1000 ensemble members given by the HRB model), and their cumulative probabilities are computed. Similarly, the appropriate category for the observation is identified and the observation's cumulative probabilities are computed (i.e., all categories less than the observation's position are assigned zero and all categories equal to and greater than the observation's position are assigned 1). Now the RPS is computed as the squared difference between the observed and forecast cumulative probabilities, and the squared differences are summed over all three categories.

The RPSS is then computed as

$$RPSS = 1 - \frac{RPS_{\text{forecast}}}{RPS_{\text{climatology}}} \quad (8)$$

Where RPS_{forecast} is the mean ranked probability score for model forecast and $RPS_{\text{climatology}}$ is the mean ranked probability score for climatologically forecast. RPSS represents the level of improvement of the forecast in comparison to reference forecasts which is usually assumed to be climatology. Similar to RE and CE, an $RPSS > 0$ indicates that the forecasts have skill better than the climatology, and vice versa.

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The results for RE, CE, and RPSS performance under m-fold cross-validation for each station are shown in Fig. 7. We observe that typically, the Hierarchical Bayesian model leads to values of RE, CE and RPSS greater than zero for all the stations (except Huoshan station –14) indicating that the seasonal forecasts developed using the climate precursors contain useful information.

In addition to computing these traditional performance measures, the performance of the posterior probability distribution is assessed by examining the model's ability to cover the observed rainfall and flows within a specified credible interval under m-fold cross validation. We estimated the coverage rates for the 90 % credible intervals under cross-validation. For each validation period, we count the number of failures or the number of observations that are outside the 5th and 95th percentile of the posterior distribution for each station resulting from the model developed using the fitting period. By computing the total number of failures from all the randomly selected models, we estimated the coverage rate as the percentage of failures for a total of 300 (30 × 10 yr) forecasts. The average coverage rate across the stations is 92 % for the corresponding to the 90 % coverage interval indicating the robustness of the fitted Bayesian models under cross-validation.

4 Summary and discussion

This study investigated the predictability of the summer rainfall and streamflow in the upper and mid Huai River basin using 5 selected large scale climate indices as predictors. The partial-pooling hierarchical Bayesian regression model provided a useful way to model spatial co-variability in seasonal hydrological predictions, while considering the potentially common effects of the predictors on regional hydrologic response. An advantage of the approach is that it allows appropriate grouping of information in the region, and explicit modeling of the covariance of the model errors and the regression coefficients to better represent the uncertainty in both the model parameters and the final streamflow and rainfall forecasts. Cross-validated model results show good pre-

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dictive skill, and the common effects as well as the at site effects of each predictor are identified well even under leave 20 % out cross validation. Our future work in the region will focus on developing dynamic rules for operating multiple reservoir systems in the basin using both multi-site seasonal hydrologic forecasts and changing demand through adaptive human behavior to better manage deficits from the reservoirs. The refinement of the method applied here to disaggregating the rain and streamflow in time over the season to properly capture monsoon breaks and the amplitude, duration and spatial structure of rainfall events will be a goal for these applications.

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Table 1. Detail information for streamflow and rainfall station used in this study.

| Station ID | Station Name | Abbreviation | Elevation (m) | Category | Actual data record & used for reconstruction |
|------------|--------------|--------------|---------------|------------|--|
| 1 | Bengbu | BB | 10.0 | streamflow | 1951–2010 |
| 2 | Lutaizi | LTZ | 19.0 | streamflow | 1951–2010 |
| 3 | Xuchang | XC | 66.8 | rainfall | 1952–2010 |
| 4 | Xihua | XH | 52.6 | rainfall | 1955–2010 |
| 5 | Zhumadian | ZMD | 82.7 | rainfall | 1958–2010 |
| 6 | Xinyang | XY | 114.5 | rainfall | 1951–2010 |
| 7 | Shangqiu | SQ | 50.1 | rainfall | 1953–2010 |
| 8 | Gushi | GS | 57.1 | rainfall | 1952–2010 |
| 9 | Bozhou | BZ | 37.7 | rainfall | 1953–2010 |
| 10 | Fuyang | FY | 30.6 | rainfall | 1953–2010 |
| 11 | Shouxian | SX | 22.7 | rainfall | 1955–2010 |
| 12 | Bengbu | BB | 18.7 | rainfall | 1952–2010 |
| 13 | Liuan | LA | 60.5 | rainfall | 1956–2010 |
| 14 | Huoshan | HS | 68.1 | rainfall | 1954–2010 |

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Table 2. The correlation between streamflow/area rainfall and climate predictors chosen in this study^a.

| | SST1 | SST2 | AMO | NAO | SNAO |
|---------------|------------|------------|------------|-------------|------------|
| Lutaizi | 0.44 (FMA) | 0.45 (OND) | 0.28 (NDJ) | −0.26 (FMA) | 0.39(FMA) |
| Bengbu | 0.44 (FMA) | 0.47 (OND) | 0.21 (NDJ) | −0.22(FMA) | 0.34 (FMA) |
| Area rainfall | 0.46 (FMA) | 0.51 (OND) | 0.36(NDJ) | −0.27 (FMA) | 0.36 (FMA) |

^a () is the selected period of the predictor for streamflow prediction; Area rainfall is the average rainfall of 12 stations in the study region.

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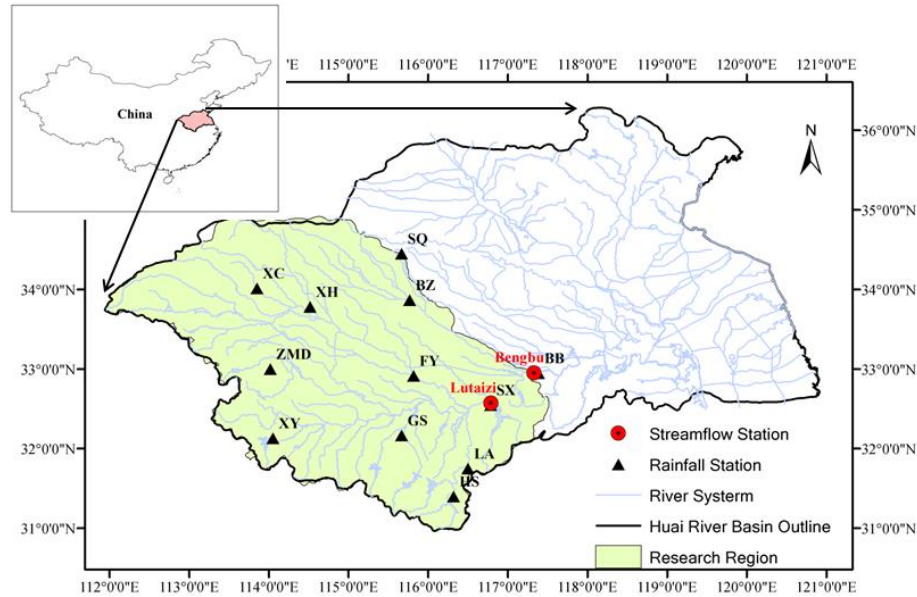


Fig. 1. Location of the study area.

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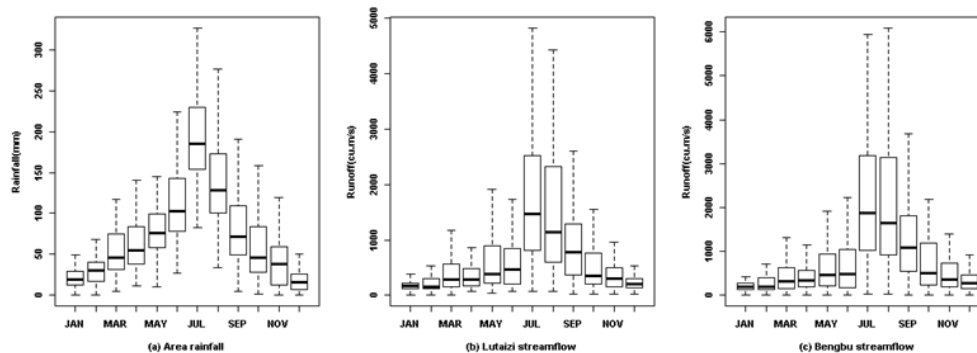


Fig. 2. Seasonality of area rainfall and streamflow in study region.

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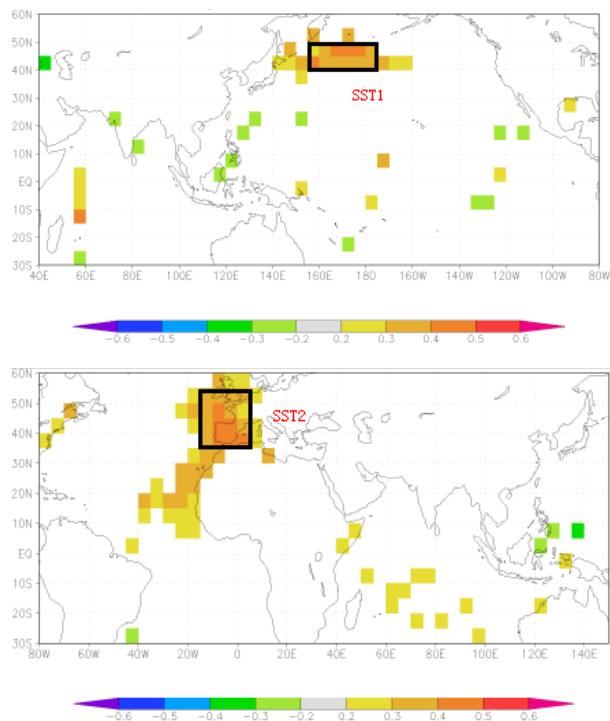


Fig. 3. SST regions (SST1 and SST2) that influence the rainfall and streamflow in the Huai River Basin. SST regions that have significant correlation at 95 % confidence interval (> 0.25 or < -0.25) are considered as predictors for the Hierarchical Bayesian Model (SST1: $40^{\circ}\text{N} \sim 50^{\circ}\text{N}$, $155^{\circ}\text{E} \sim 175^{\circ}\text{E}$; SST2: $35^{\circ}\text{N} \sim 55^{\circ}\text{N}$, $15^{\circ}\text{W} \sim 5^{\circ}\text{E}$).

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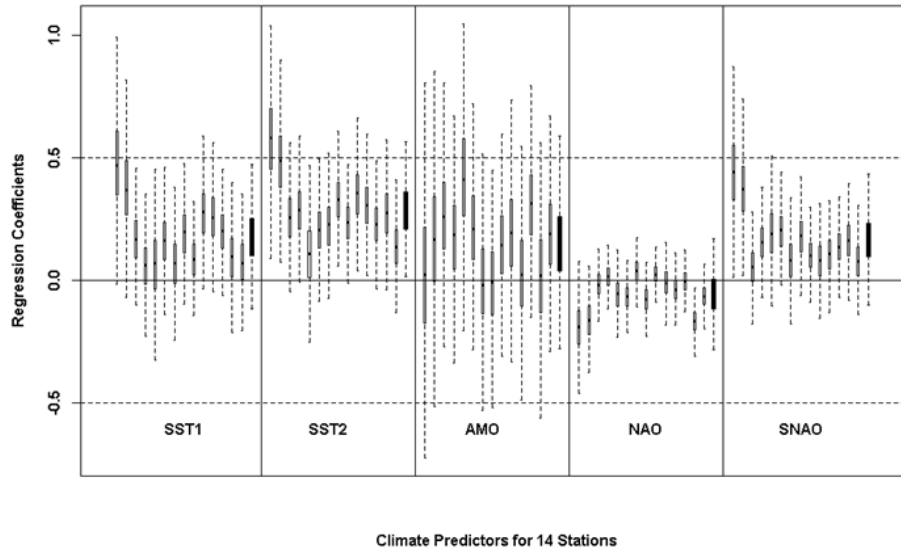


Fig. 4. Boxplots of the regression coefficients β_{ij} (open box) and the μ_{β_j} mean of the regression coefficients (filled box) for the five predictors. The first two boxplots for each predictor correspond to the streamflow stations and the next 12 to the rainfall stations.

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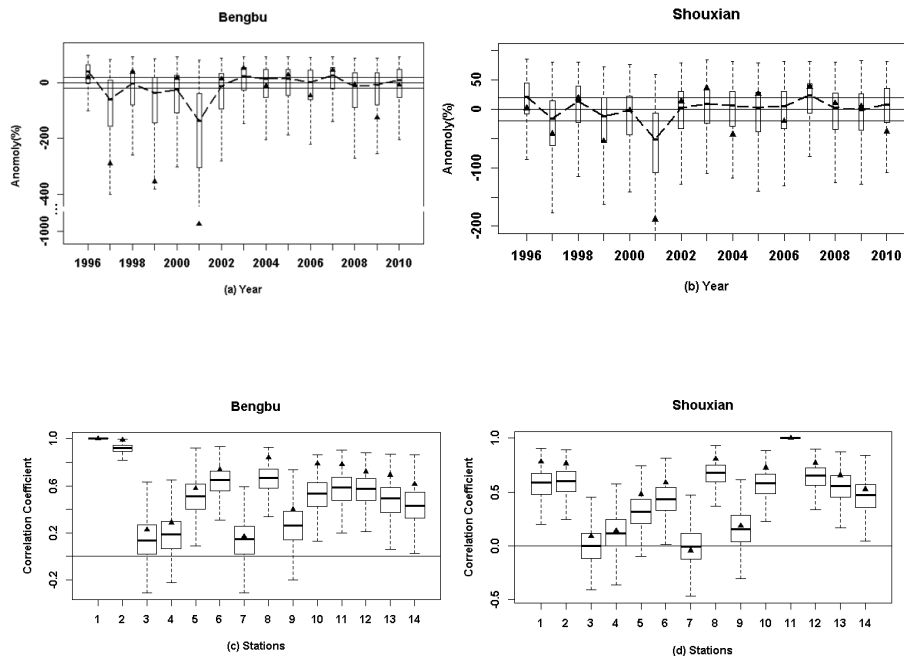


Fig. 5. The posterior probability distributions for JJA averaged streamflow and JJA total rainfall for **(a)** standardized streamflow at Bengbu station and **(b)** standardized rainfall at Shouxian station. The posterior distribution of the cross-site correlation for Bengbu station and for Shouxian station is presented in **(c)** and **(d)** respectively. Each boxplot has the 25th, median and 75th percentile of the posterior distribution, with whiskers extended to the extreme values sampled. The solid triangle denotes the observation.

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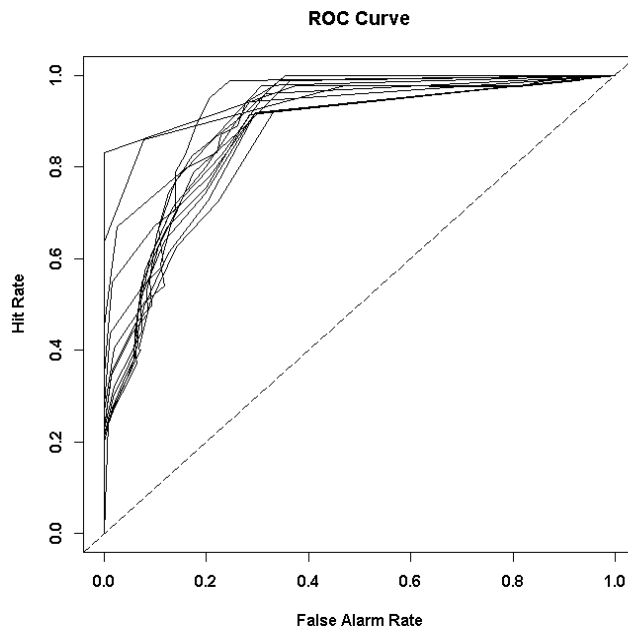


Fig. 6. ROC plot from the forecasts based on decile categorical thresholds for all the 14 stream-flow and rainfall stations.

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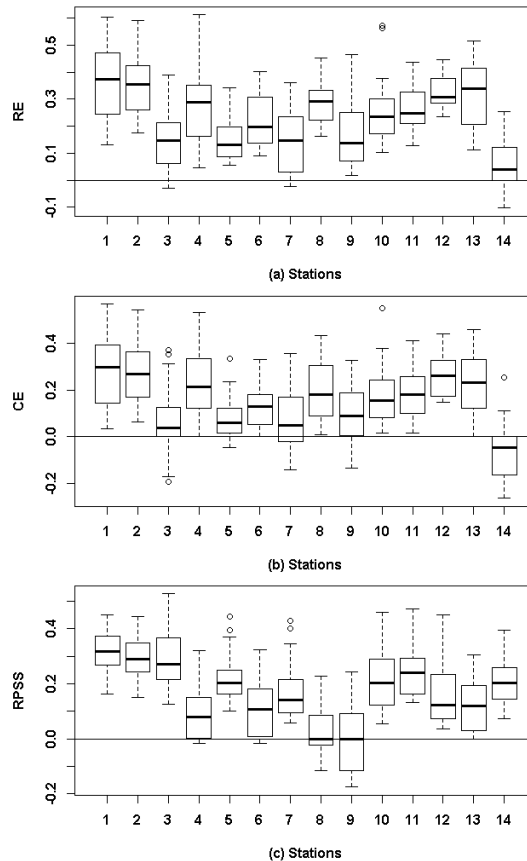


Fig. 7. Performance under “leave 10 out cross validation” for all 14 stations from the hierarchical Bayesian model from 30 random simulations ((a) Reduction of Error (RE), (b) Coefficient of Efficiency (CE), (c) Rank probability skill score (RPSS)).

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