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SENSITIVITY OF STREAMFLOW TO CLIMATE CHANGE: A CASE STUDY FOR NEBRASKA

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Abstract. *Climate change, whether natural or due to human action, will have an impact on many aspects of our environment. The nature of streamflow changes will depend on the magnitude and direction of the climate change. However, since the principal climatic factors that control streamflow are precipitation and evapotranspiration (which can be estimated from air temperature data), the sensitivity of streamflow to variations in climate can be studied through the use of plausible scenarios of climate change.*

A simple water budget model was used to reconstruct streamflow from monthly temperature and precipitation data for locations within and immediately surrounding the Little Blue River basin in south-central Nebraska. Then, climate change scenarios corresponding to changes in monthly temperature of 1°C and 3°C and to differences in monthly precipitation of 10% and 20% were used to estimate the sensitivity of streamflow to climate change. Results of this procedure show the sensitivity of streamflow to climate variability. For example, a 20% increase in precipitation would more than double the average annual streamflow, while a 20% precipitation decrease would almost halve the average annual streamflow. The effects of temperature changes are similar, with a 3°C increase resulting in an almost 60% decrease in streamflow, and a 3°C decrease causing streamflow to increase by more than 80%. Scenarios with both temperature and precipitation changes can either enhance or nullify the effects of a single change.

Changes in streamflow will affect water availability for agricultural, human consumptive, industrial, and recreational uses. For a region with critical water needs, such as the Great Plains, understanding the possible consequences of climate change on streamflow is necessary to ensure adequate future supplies. The simple streamflow model presented here can easily be applied to other streams in the Great Plains to evaluate the regional effects of climate change on water supply.

That the climate of the Earth has changed in the past and will change in the future is not in doubt. Since the Industrial Revolution, human alteration—both inadvertent and intentional—of the concentrations of radiatively-active trace gases in the atmosphere has risen to levels at which human activity may become a significant cause of climate change. However, even as there appears to be increasing evidence for and scientific consensus about the human impact on global climate, there remains considerable uncertainty concerning the regional patterns of climate change. For this reason, studies of the impact of climate change on natural and managed systems (e.g., hydrologic or agricultural systems), which by their very nature must be undertaken at a regional scale, have an inherently high degree of uncertainty. Therefore, the present study focuses not on the expected response of a hydrologic system to a given climate change, but rather on the sensitivity (i.e., range of responses) of streamflow to a range of hypothetical climate change scenarios.

To assess the effects of climate change on stream runoff in the United States several investigators (Stockton and Boggess 1979; Revelle and Waggoner 1983; Karl and Riebsame 1989) have used variations of the empirical statistical models relating precipitation, temperature, and runoff demonstrated initially by Langbein (1949). He studied annual runoff for different values of mean annual precipitation and precipitation-weighted mean annual temperature for 22 drainage basins in the coterminous United States. Langbein found that for any annual precipitation runoff diminishes rapidly with increasing temperature, and that for any given temperature the proportion of precipitation that runs off increases rapidly with increasing precipitation.

Stockton and Boggess (1979) used Langbein's empirical climate-runoff relationships to estimate the hydrologic effects of hypothetical changes in both temperature (2°C) and precipitation (10%) in the 18 United States water regions. They concluded that a change toward a warmer and drier climate would have the greatest effects on runoff nationwide. It was estimated that the most severe impacts would be felt in seven western regions where

a 2°C increase in temperature and a 10% decrease in precipitation would result in a 40 to 76% reduction in runoff.

In a study of the Colorado River Basin, Revelle and Waggoner (1983) combined Langbein's empirical climate-runoff relationships with climate model projections by Manabe and Wetherald (1980). Their study indicated that a 2°C increase in mean annual temperature combined with a 10% increase in mean annual precipitation would result in an 18% decrease in runoff. Thus, even with an increase in annual precipitation in their model, increased evapotranspiration reduced net annual runoff. They showed that a 2°C rise in temperature would decrease runoff nearly three times more than a 10% decrease in precipitation.

Karl and Riebsame (1989) performed an empirical analysis of actual climate fluctuations and the associated runoff changes over the past 50 years for 82 river basins in the United States. Their study indicated that changes in mean annual temperature of 1°C would have little effect on runoff, but that 10% changes in annual precipitation could alter runoff by more than 10%. They concluded that due to an overestimation of the role of evapotranspiration, temperature variability is not as great a factor in runoff as earlier studies suggest. Their research indicates that the precipitation-driven amplification of runoff should be of greatest concern when investigating the impacts of global climate change.

These empirical and statistical studies provide simple but crude approximations of runoff. However, they do not measure the impact of physical factors such as topography, soil, geology, vegetation, and size of the drainage basin. Therefore, only general conclusions concerning the impact of climatic changes on water resources can be drawn from their results (Gleick, 1986).

Other investigators (Nemec and Schaake 1982; Wigley and Jones 1985; Gleick 1986, 1987a, 1987b; Bultot, Coppens, Dupriez, Gellens, and Meulenberghs 1988; Bultot, Dupriez, and Gellens 1988; Thomas 1990) have used deterministic, physically-based hydrologic models and analyses. These physically-based hydrologic models have also generally shown that annual runoff is more sensitive to fluctuations in precipitation than to temperature. For example, Wigley and Jones (1985) presented theoretical and empirical modeling arguments and controlled environmental experiments to show that precipitation changes are more dominant than evapotranspiration changes in affecting runoff, particularly for high runoff ratios.

Nemec and Schaake (1982) addressed the sensitivity of stream runoff to hypothetical climate change scenarios in one arid and one humid basin in

the southern United States. They used a hydrologic model that includes parameterizations of the physical processes affecting stream runoff and found that moderate variations in climatic variables (1°C and 10% annual precipitation) would result in significant changes in runoff. For a temperature rise of 1°C and a decrease in precipitation of 10%, they found a 25% reduction of average annual runoff in the humid basin and a 50% decrease in the arid basin, with similar increases in runoff resulting from increased precipitation.

Gleick (1986, 1987a) reviewed the use of modified water-balance models to evaluate the regional hydrologic impacts of global climatic changes. Some advantages of water balance models are that they can incorporate monthly or seasonal climatic variables, soil moisture characteristics, groundwater fluctuations, snowfall, and snowmelt. Gleick (1987a, 1987b) developed a modified water-balance modeling technique that he applied to the Sacramento Basin in California to examine the effects of several climatic change scenarios on water availability. His study showed that annual runoff is more sensitive to changes in precipitation rather than temperature. However, the seasonal distribution of runoff and soil moisture was affected by changes in mean monthly temperature.

Bultot, Coppens, Dupriez, Gellens, and Meulenberghs (1988) and Bultot, Dupriez, and Gellens (1988) applied a physically-based hydrologic model to three river basins in Belgium to estimate actual evapotranspiration, soil moisture, and runoff for present climatic conditions and those predicted for the case of CO_2 doubling. Their model predicted increases of potential and actual evapotranspiration throughout the year, a reduction in snowcover, inter-basin differences in groundwater storage and soil moisture related to sub-surface soil types and infiltration rates, and increases in winter runoff and flooding with reductions in summer streamflow.

The present study utilizes a modified Thornthwaite-Mather water budget model to estimate the sensitivity of discharge from the Little Blue River in south-central Nebraska to hypothetical climate change scenarios expressed as changes in monthly temperature and precipitation. Thompson (1992) used the Thornthwaite-Mather water budget approach to simulate the effects of a single climate change scenario on climatological water balances in Missouri and Kansas. He found that a 2.5°C temperature increase coupled with a 10% increase in precipitation could cause a 25% or more decrease in total annual runoff. The present study differs from Thompson's in that the effects of multiple climate change scenarios are investigated in terms of time-series of water budget data rather than a climatological water balance.

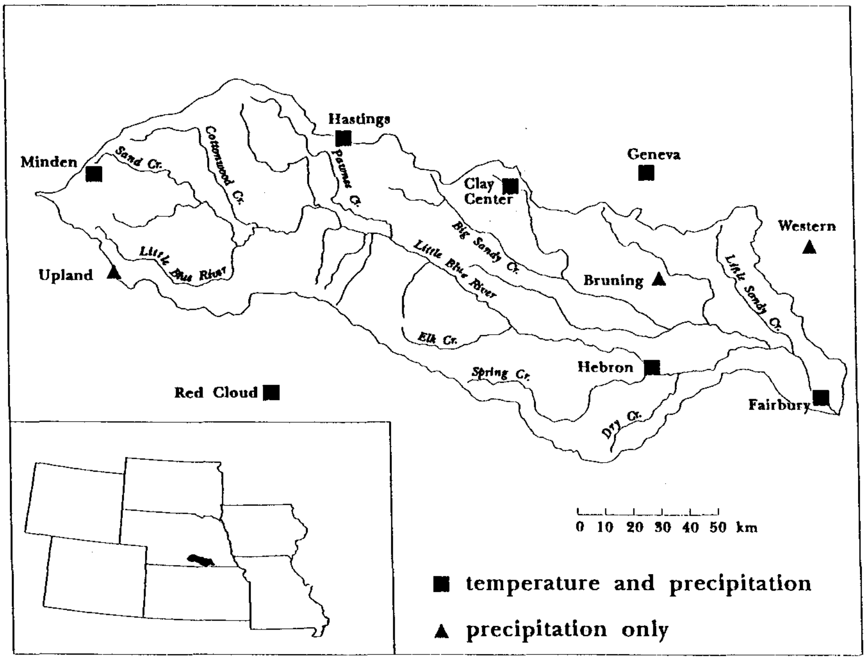

Figure 1. The Little Blue River Basin above Fairbury, NE. Major tributaries of the Little Blue River are shown, as are the locations of the climatological stations used in this study.

The Little Blue River Basin

The Little Blue River is the principal tributary of the Big Blue River. Its drainage basin includes all or part of ten counties in south-central and southeastern Nebraska as well as small portions of three counties in northern Kansas. The basin is bounded by the Republican River Basin on the west and southwest, the Middle Platte River Basin on the north and the Big Blue River Basin on the east.

The Little Blue River Basin, upstream of Fairbury, Nebraska (Fig. 1) has a total drainage area of 6,016 km^2 (2,350 mi^2) and is composed of plains, rolling hills, stream valleys, and some marshy depressions. The upper (western) portion of the basin is a loess plain with poorly defined drainage patterns. In the central part of the basin, the plains are gently rolling, with a

well-defined drainage pattern on the south side of the river. Bedrock is near the land surface in the lower basin, with outcroppings on the steeper slopes where severe erosion has occurred (Corps of Engineers 1971; Nebraska Natural Resources Commission [NNRC] 1976).

Elevation varies from 400 m (1,310 ft) above mean sea level at Fairbury, to 580 m (1,900 ft) in the western headwaters. The stream slopes about 1 m km^{-1} (5 ft mi^{-1}), with the channel averaging about 90 m (300 ft) wide through the study area (Corps of Engineers 1971). Twenty-four main tributaries enter the Little Blue River above Fairbury. There are no large impoundments in the basin, although there are numerous farm ponds and some marshy depressions throughout the watershed.

The Little Blue River Basin has a varied climate with wide fluctuations in temperature, precipitation, and humidity (Corps of Engineers 1971). The climate is characterized by two well-defined seasons, winter and summer, and two transitional seasons, spring and autumn, which are chiefly composites of winter and summer patterns. The winters are usually long and cold with snow cover often present. Spring is usually brief, cool, rainy, and often windy. Summers are long, hot, and accompanied by many thunderstorms and high evapotranspiration losses. The autumn is long and warm with occasional periods of rain. Mean annual precipitation ranges from about 585 mm (23 in) in the western portion of the basin to 735 mm (29 in) in the east. About 75% of the precipitation comes as rainfall during the growing season, but the distribution of this largely convective precipitation is not always conducive to good crop production (NNRC 1976). Evapotranspiration consumes most of the precipitation falling on the land surface. Mean annual temperature across the basin is approximately 11.7°C (53°F) with little spatial variation. The mean frost-free period ranges from approximately 160 days in the northwestern portion of the basin to 175 days in the southeastern region (NNRC 1976).

Soils in the Little Blue River Basin have developed from loess parent material (NNRC 1976; Dugan 1984) and are relatively impermeable. These loess soils have low rates of infiltration of surface water into the ground. However, slopes over most of the basin are gentle so that there is sufficient time for water to infiltrate into the soils. In fact, ponding is common on many soils in the basin following precipitation events. Areas of steeper slope, such as stream valleys, are characterized by soils having higher permeability (Dugan 1984).

The natural vegetation changes from tall grass prairie in the eastern part of the basin to mixed grass prairie in the west. Approximately 1.5% of the basin is forested, primarily along the Little Blue River and its tributaries,

where the elm-ash-cottonwood forest type predominates, and with some stands of bur oak and black walnut in the lower portions of the basin (NNRC 1976).

Crop irrigation provides the primary use of surface water in the basin. While the amount of surface water used for irrigation varies from year to year, depending on the amount of rainfall that occurs during the growing season, the greatest source of irrigation water in the region is groundwater. In 1975, for example, the ratio of groundwater to surface-water used for irrigation was 34:1 in the Little Blue River Basin (Bentall and Shaffer 1979). Return flows of irrigation water to the Little Blue are small and studies of the stream discharge records plus the use of an analog groundwater model have failed to disclose any appreciable effect of the groundwater use on stream discharge (Kansas-Nebraska Big Blue River Compact Commission 1968).

Methods

Thornthwaite-Mather Water Budget

Since its development by C. W. Thornthwaite in the 1940s, the climatic water budget has been used to evaluate a number of stream basin hydrologic parameters including evapotranspiration, soil moisture, and runoff (Thornthwaite 1948; Thornthwaite and Mather 1955). Thornthwaite's water budget procedures have been found to be useful in many applications that require streamflow estimates. Thornthwaite's empirical approach to estimating potential evapotranspiration (PE) has been criticized for underestimating PE, especially in semi-arid or arid regions (Pelton et al. 1960; Pruitt 1960, 1964; Stanhill 1961; Hashemi and Habibian 1979). However, Pereira and Paes de Camargo (1989) have recently offered evidence that this criticism is a result of misapplication or misunderstanding of Thornthwaite's method. They note that Thornthwaite's PE estimates are most often compared to pan evaporation or evapotranspiration from small experimental plots. In semi-arid and arid regions these small sites are affected by advection of sensible heat from the surrounding environment, resulting in a large added energy source for evapotranspiration from the moist oasis (i.e., the so-called "oasis effect"). Thornthwaite's PE computations were developed for situations in which sensible heat advection would be minimal (i.e., in areas with adequate fetch) and are thus better suited for estimating regional evapotranspiration from a relatively homogeneous area rather than from a site not representative of its surroundings. When compared to evapotranspiration measured at sites with adequate fetch, Thornthwaite's approach has been found to produce quite

reasonable estimates of PE (Pereira and Paes de Camargo 1989). Furthermore, when estimating soil moisture deficit, Calder et al. (1983) found that the method used to estimate PE was less critical than the techniques used to simulate soil moisture extraction and recharge. They found that sophisticated PE models did not produce significantly better results than simple climatological estimates of PE.

While the Thornthwaite-Mather water budget procedure can be used to provide month-to-month streamflow values, annual values have proven to be more reliable under a wide array of climatic and hydrologic conditions (Mather 1979). One of the primary advantages of water budget methods over more complex streamflow models is the minimal data requirement—temperature, precipitation, and a few hydrologic parameters. It is precisely these minimum data needs that make a water budget method extremely well-suited for studies of the impact of climate change and variability on streamflow.

Streamflow estimation using the climatic water budget is essentially a bookkeeping procedure that tracks inputs and withdrawals of moisture to and from the soil. Inputs of moisture are represented by rainfall and snowmelt while withdrawals of soil moisture include evapotranspiration and drainage. Thornthwaite's original water budget formulation did not account for snowmelt, nor did it explicitly include drainage. However, drainage is implied in Thornthwaite's surplus term, which is non-zero whenever soil moisture exceeds the field capacity of the soil. Several methods for including snow accumulation and melt in the water budget have been developed in the past 40 years. The most useful methods for water budget studies are those that do not increase the data requirements; that is, those that estimate snowfall, accumulation, and melt using only the temperature and precipitation data already needed to calculate the water budget.

The method used to compute the climatic water budget in the present study is that of Willmott et al. (1985), which closely follows the Thornthwaite-Mather (1955) procedure. The differences introduced by Willmott et al. (1985) are the inclusion of a separate, interacting snow-cover budget and the use of the soil moisture availability function of Mintz and Serafini (1984). Details of the governing equations and computational procedures employed in the present study are outlined in the appendix. The present study, however, uses a field capacity of 254 mm (10 in) which is characteristic of the deep-rooted vegetation and loamy soils of the study region (Thornthwaite and Mather 1957; Palmer 1968; Main 1979) rather than the value of 150 mm (6 in) used worldwide by Willmott, et al. (1985).

Monthly moisture surpluses predicted by the water budget method are not used directly as estimates of monthly streamflow. This is because not all

of the available surplus moisture will have sufficient time to move through the soil and into a stream channel. In other words, the physical characteristics of the basin such as size, slope, vegetation cover, soil type, and subsurface material will introduce a delay between the time of surplus soil moisture and its appearance as streamflow. The simple approach used here is to have a constant fraction of surplus moisture contribute to streamflow in the month when it occurs, while the remainder is held over to the following month and added to any surplus for that month (van Hylckama 1958; Mather 1978, 1979). This procedure produces the lag necessary to simulate realistic streamflow, and helps to maintain a base flow during seasons with no soil moisture surplus, except in times of severe drought. For the current study it was assumed that 25% of the monthly surplus contributes to the streamflow of that month while the remainder (75%) is added to the succeeding month's surplus, which is then subject to the same delay factor. This factor cannot be determined by direct physical argument, but it is in agreement with van Hylckama's (1958) findings for basins with little topographic relief.

Climate Change Scenarios

When attempting to evaluate the response or sensitivity of any physical (or biological) system to climate change, one of the largest uncertainties introduced is our current level of understanding (or lack thereof) of the magnitude, or even the direction of future climate change. Even if global climate change could be modeled using today's general circulation models (GCMs), much climatic variation takes place at regional and smaller scales that are unresolved and will remain so for the foreseeable future. Because of this, studies of the effects of climate change on hydrologic systems are limited to the use of climate change scenarios that may or may not match future climate realities. However, these scenarios are useful for investigating the response of hydrologic systems to climate change and variability since they are easily constructed and employed as inputs to other models.

A number of different approaches to developing climate change scenarios have been devised in recent years. These include GCM output, analog climates (historical, paleoclimatic or spatial), synthesis scenarios ("scenarios by committee"), arbitrary change scenarios, or scenarios based on physical or statistical arguments (World Meteorological Organization 1987). While GCM output can provide some indication of the direction as well as the possible magnitude of a climate change associated with some forcing (e.g., doubled CO_2), the uncertainties associated with GCMs, as well as their poor spatial resolution, reduce their usefulness for studies of regional hydro-

logic consequences of climate change. Although resource managers and planners may desire indications of climate change direction and magnitude, GCM output must be used cautiously. Hypothetical, arbitrary climate change scenarios can be developed at much lower cost than GCM scenarios, and can provide useful information on the response of hydrologic systems to plausible levels of climate change and variability.

Only two climatic inputs (temperature and precipitation) are needed to compute the Thornthwaite-Mather climatic water budget. Scenarios with mean annual temperature changes of 0°C, 1°C and 3°C and annual total precipitation changes of 0%, 10%, and 20% were constructed with the assumption that all months experienced the same change (i.e., constant temperature change or percentage precipitation change). While not all of the resulting twenty-five scenarios are equally likely, and real climate changes will undoubtedly affect the seasonal cycle as well as the mean climate, these scenarios offer a simple basis on which to evaluate the impacts of climate change and variability on streamflow.

Data

Climatic data were obtained through the High Plains Climate Center (HPCC) from two different sources. The majority of the data was taken from a CD-ROM that includes daily climatic information for around 5,000 climatological stations in the United States (U.S. West 1990). Additional data for stations or time periods not included on the CD-ROM were obtained from the Climatological Data Annual Summaries for Nebraska published by the National Oceanographic and Atmospheric Administration (NOAA). These data were encoded and added to the data extracted from the CD-ROM. From these sources daily minimum and maximum temperature for seven different climatological stations and daily precipitation totals for these and three additional stations (Table 1) were selected for the period 1925-1988. Using a program developed at the HPCC monthly averages of temperature and totals of precipitation were computed. Monthly values could not be computed for some stations for certain months because of missing data. Missing monthly temperatures were replaced with that month's average computed for the entire period. Because of the greater temporal and spatial variability of precipitation, missing monthly precipitation totals were replaced with the average monthly precipitation for the remaining stations. No more than a single station had missing values for either temperature or precipitation in any given month.

TABLE 1

CLIMATOLOGICAL STATIONS FOR THIS STUDY, WITH WEIGHTING FACTORS USED TO COMPUTE BASINWIDE TEMPERATURE AND PRECIPITATION

				Temperature		Precipitation	
Station	lat	lon	elev (m)	area (km^2)	weight	area (km^2)	weight
Bruning	40°20'N	97°34'W	482	-----	-------	727.3	0.11888
Clay Center	40°30'N	97°56'W	530	1113.2	0.18196	1074.6	0.17566
Fairbury	40°07'N	97°10'W	408	346.8	0.05669	269.6	0.04407
Geneva	40°32'N	97°36'W	497	364.1	0.05953	61.9	0.01012
Hastings	40°35'N	98°21'W	588	1208.0	0.19746	1222.7	0.19987
Hebron	40°10'N	97°35'W	451	1619.0	0.26465	1129.8	0.18467
Minden	40°30'N	98°57'W	661	937.1	0.15317	573.2	0.09369
Red Cloud	40°06'N	98°31'W	524	529.4	0.086544	385.1	0.06296
Upland	40°19'N	98°54'W	658	-----	-------	497.5	0.08133
Western	40°24'N	97°12'W	451	-----	-------	175.9	0.02875

Thiessen (1911) polygons were used to estimate basinwide temperature and precipitation for each month. Thiessen polygons are constructed around a set of climatological stations in such a way that all locations within a given polygon boundary are closer to the station enclosed by the polygon than to any other station. The entire area contained within a polygon was considered to have monthly temperatures and precipitation equal to that of the enclosed station. By weighting each station's temperature and precipitation time series by the fractional area of the stream basin enclosed by its

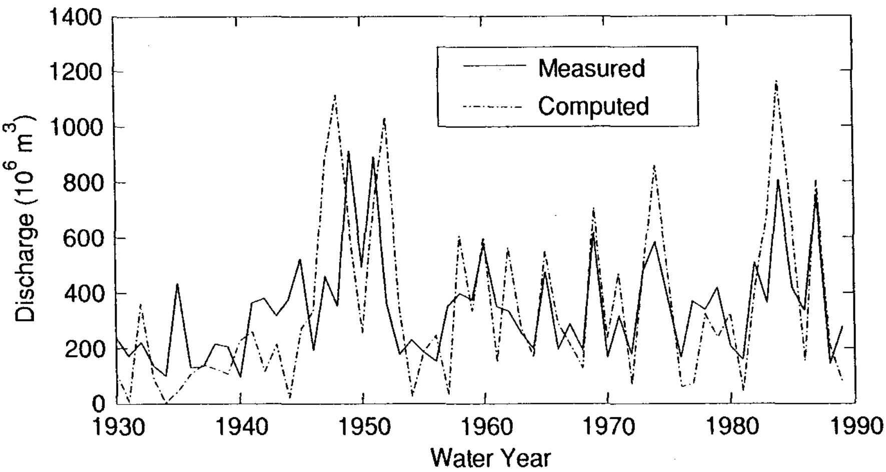

Figure 2: Time series of annual measured and computed discharge of the Little Blue River at Fairbury, NE for 1930-1989. Monthly values have been summed over water years (October-September).

polygon, basinwide estimates of monthly temperature and precipitation were obtained. These basinwide time series were used as the input for the streamflow model described below.

Only one active U.S. Geological Survey (USGS) stream gaging station in the Little Blue River Basin has long-term flow records. This gage furnishes a reliable estimate of streamflow from the basin as it includes all major tributaries except Rose Creek (downstream of Fairbury). USGS provided the monthly streamflow data for water years 1930 through 1989.

Results

Under present climate conditions, the simple water budget model described here is able to estimate the average annual stream discharge quite well, but the interannual variability of the model is too large (Figs. 2 and 3; Table 2). This may be, in part, due to an unmodeled baseflow in the Little Blue River and some of its tributaries that is not greatly influenced by interannual climate variability. This high year-to-year variability in the modeled discharge is reflected in the mean absolute error (MAE) and root mean square error (RMSE) of the model, with almost all (> 99 percent) of the RMSE due to unsystematic (i.e., random) errors (Table 2). The model pre-

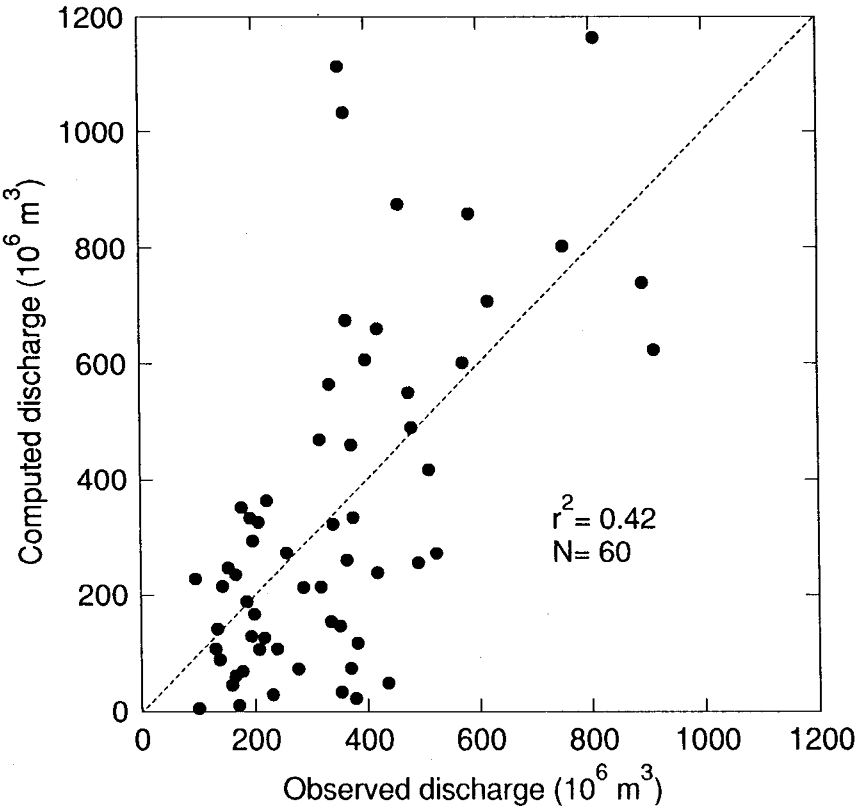

Figure 3: Computed versus measured annual discharge of the Little Blue River at Fairbury, NE for 1930-1989. Monthly values have been summed over water years (October-September). The best-fit linear regression is shown and does not differ significantly from the 1:1 line.

dicts discharge well below average during the drought periods of the 1930s and early 1940s, mid-1950s and 1970s, and much higher than average discharge during the wet periods that occurred in the late 1940s to early 1950s as well as in the mid-1970s and mid-1980s (Fig. 2). However, the model tends to predict more extreme droughts and longer recovery periods following droughts than were observed. Once again, this could be due to unmodeled baseflow from groundwater sources.

TABLE 2

MODEL VALIDATION STATISTICS FOR THE SIMPLE STREAM DISCHARGE MODEL

Statistic[1]	Value[2]	Confidence limits (95%)[3] lower	Confidence limits (95%)[3] upper
o	340.9605	299.7001	398.6277
σ_o	187.9098	129.7393	229.9155
p	339.9277	266.7738	412.3053
σ_p	292.6178	221.0314	334.3962
MAE	167.1948	132.7246	202.3701
RMSE	221.2123	172.2835	269.5191
$RMSE_s$	1.7852		
$RMSE_u$	221.2051		

[1] o and p are the observed and predicted mean discharges; σ_o and σ_p are the observed and predicted standard deviations; MAE is mean absolute error; RMSE, $RMSE_s$ and $RMSE_u$ are root mean square error and its systematic and unsystematic components (see Willmott et al. 1985).

[2] all values are expressed as 10^6 m^3.

[3] confidence limits were estimated using a non-parametric bootstrapping technique (see Willmott et al. 1985).

Model results for the climate change scenarios (Table 3) indicate the sensitivity of streamflow to climate variability. For example, a 20% increase in precipitation would more than double average annual streamflow, while a 20% precipitation decrease would almost halve the average annual streamflow. The effects of temperature changes are similar, with a 3°C increase resulting in an almost 60% decrease in streamflow, and a 3°C decrease increasing streamflow by over 80%. Scenarios that include both a temperature change and a precipitation change can result in either an enhancement or a nullification of the effects of a temperature or precipitation change

TABLE 3

AVERAGE WATER-YEAR DISCHARGE FROM THE LITTLE BLUE RIVER AT FAIRBURY, NE

Results are presented for the 25 hypothetical climate change scenarios (combinations of a temperature change, ΔT, and a precipitation change, ΔP) as well as observed discharge (1930-1989).

Average water-year discharge for climate scenarios (10^6 m^3)							Observed Discharge (10^6 m^3)
				ΔP			
		-20%	-10%	0%	+10%	+20%	
	+3° °	77.17	97.60	140.87	223.79	374.19	
	+1° °	134.68	175.37	271.24	413.51	635.28	
ΔT	0°	170.87	238.46	339.93	525.08	784.71	340.96
	-1°	222.16	288.86	418.93	644.19	939.51	
	-3°	304.96	421.06	626.29	918.96	1242.34	

alone. That is, a 20% increase in precipitation coupled with a 3°C temperature decrease increases computed streamflow nearly fourfold, while a 20% precipitation decrease and 3°C temperature increase reduces computed streamflow to less than 25% of the present average flow. Alternatively, increasing (or decreasing) temperature by 3°C and precipitation by 20% together has little net effect on computed mean annual streamflow (approximately 10%).

These results are consistent with those reported in previous studies, both in magnitude and sign, and appear to support the findings that streamflow is more sensitive to changes in precipitation than to changes in temperature. However, upon closer inspection it is apparent that the changes in PE caused by the hypothetical temperature changes are smaller (as a percentage

TABLE 4

EFFECTS OF CLIMATE CHANGE SCENARIOS ON POTENTIAL (PE) AND ACTUAL (AE) EVAPOTRANSPIRATION IN THE LITTLE BLUE RIVER BASIN

		ΔPE (%)	ΔAE (%) ΔP				
			-20%	-10%	0%	+10%	+20%
	+3°	18.54	-14.91	-4.58	5.22	14.00	21.10
	+1°	5.66	-16.43	-6.60	1.90	9.15	14.45
ΔT	0°	0.00	-17.45	-8.32	0.00	6.18	10.41
	-1°	-5.24	-18.75	-9.63	-2.02	3.06	6.53
	-3°	-14.55	-21.07	-13.21	-7.54	-3.91	-1.83

of current PE) than the precipitation changes (Table 4). Therefore, it is not surprising that the modeled discharge seems to be more sensitive to precipitation variation. Changes in actual evapotranspiration (AE) are smaller than those for PE, when precipitation is unchanged, since AE is affected by both the atmospheric demand for water (PE) and the supply of water (precipitation and soil moisture). This leads to amplification factors (Wigley and Jones 1985; Karl and Riebsame 1989) for temperature (expressed in terms of PE or AE) of the same magnitude as those for precipitation. This is in agreement with the results of Wigley and Jones (1985) for basins with small runoff ratios. The runoff ratio for the Little Blue River is approximately 0.10, as is appropriate for the semi-arid nature of the region, and sensitivity to both temperature and precipitation variations is to be expected.

Conclusions

Changes in streamflow affect water availability for agricultural, human consumptive, industrial, and recreational uses. For a region with critical

water needs, such as the Great Plains, understanding the possible consequences of climate change on streamflow is necessary to ensure adequate future supplies. The simple method presented here for evaluating the sensitivity of streamflow to possible climatic change can easily be applied to other stream basins in the Great Plains or other regions to evaluate the regional effects of climate change on water supply. Our results are in general agreement with previous research in that relatively small changes in temperature or precipitation can cause much larger changes in streamflow. However, our findings indicate that for south-central Nebraska the effects on streamflow of temperature changes are as important as those due to changes in precipitation, when the temperature change is expressed as a percent change in potential evapotranspiration.

The present study only considers the sensitivity of streamflow to hypothetical climate changes that are constant over the annual cycle. However, as we increase our understanding of the altered seasonality that will undoubtedly accompany any global climate change, we will be able to incorporate that information into this model. Because actual evapotranspiration and streamflow are influenced by both temperature and precipitation, changes in the seasonality of temperature and precipitation will have as yet undetermined effects on streamflow due to non-linearities inherent in the system. Furthermore, when GCM simulations of the regional patterns of climate change become credible, temperature and precipitation outputs from those models can be utilized to generate water-budget estimates of streamflow so that the impact of a specified climate change on streamflow might be determined.

APPENDIX: The Thornthwaite-Mather Water Budget

The climatic water budget, as devised by Thornthwaite and Mather (1955, 1957), is a bookkeeping procedure that tracks the inputs (i.e., precipitation, snowmelt) and withdrawals (i.e., evaporation, transpiration, drainage, runoff) of water to and from the root zone of the soil. Thornthwaite's original formulation was developed for monthly computations of the inputs and outputs, although the bookkeeping procedure can be used for any time interval of interest. If the initial soil moisture is known or can be estimated, month-to-month changes can be determined by measurements or calculations of the various inputs and withdrawals. The amount of water that can be stored in the root zone of any soil against the pull of gravity is finite and is called the soil's field capacity (w^*, mm). Any additional water added to the

soil after it has reached field capacity will drain from the root zone and is classified as surplus. This surplus water is then available for runoff and eventually for streamflow. Withdrawals of moisture from the soil depend on the climatic demand for water (i.e., potential evapotranspiration) and the rate at which moisture can be extracted from the soil. In general, as the soil dries, it becomes more difficult to extract soil moisture at a rate sufficient to meet the climatic demand. Thus, in addition to tracking the demand for and supply of water, a soil-moisture extraction function must be determined.

The procedure adopted in the present study utilizes Thornthwaite's original definition of potential evapotranspiration (PE) for a standardized month with 30 days and 12 hours of daylight (PE', mm/month) as

$$PE' = \begin{cases} 0, & T < 0°C \\ 16(10T/I)^a, & 0°C \leq T < 26.5°C \\ -415.85 + 32.24T - 0.43T^2, & T \geq 26.5°C \end{cases} \tag{1}$$

where T is the mean monthly air temperature (°C). The heat index (I), which accounts for the seasonal variation in temperatures, is given by

$$I = \sum^{12} (T/5)^{1.514}, \qquad T \geq 0°C \tag{2}$$

and the exponent a is

$$a = 6.75\times10^{-7}I^3 - 7.71\times10^{-5}I^2 + 1.79\times10^{-2}I + 0.49 \tag{3}$$

Potential evapotranspiration (PE) is then adjusted for the actual number of days (D) in the given month as well as the mean daylength for that month (h, hours) as

$$PE = PE'\left[\frac{D}{30}\frac{h}{12}\right]. \tag{4}$$

Water to meet the demand from potential evapotranspiration can be supplied by rainfall and snowmelt or by extraction from soil moisture storage. Thornthwaite's original formulation did not include a separate snow budget, but this was added by Willmott et al. (1985) in their global water balance investigation. All soil moisture (w, mm) and snow water equivalent (w^s, mm) budget computations are performed on a pseudo-daily basis (i.e., 30 times per month, subscript d) to better represent the physical processes taking place. Daily temperature and precipitation are held constant for the month (i.e., T_d=T and P_d=P/30). First, monthly precipitation must be classified as either rain (P^r) or snow (P^s); following Thornthwaite and Mather (1955), a threshold temperature of -1°C was used so that

$$P_d = \begin{cases} P_d^r, & T \geq -1°C \\ P_d^s, & T < -1°C. \end{cases} \tag{5}$$

Daily snow melt (M_d, mm/day) is estimated (Willmott et al., 1985) as

$$M_d = 2.63 + 2.55\,T_d + 0.912\,T_d P_d^r, \qquad 0 \leq M_d \leq (w_{d-1}^s + P_d^s) \tag{6}$$

Thus, the daily budget of snow water equivalent is represented as

$$w_d^s = w_{d-1}^s + P_d^s - M_d, \qquad w_d^s \geq 0 \tag{7}$$

Thornthwaite and Mather assumed that any water supplied by precipitation and, by extension, by snow melt is immediately available to meet the potential evapotranspiration demand. Any additional demand will be obtained by extraction of soil moisture, subject to the condition of increasing difficulty of extraction as the soil dries. When precipitation and snow melt exceed the demand for water, the excess precipitation is added to the soil moisture store until field capacity is reached, at which time the water becomes surplus. Thus, the daily soil moisture store can be given as

$$w_d = w_{d-1} + \beta_d D_d, \qquad 0 \leq w_d \leq w^* \ , \tag{8}$$

where β_d is the soil moisture extraction function (Mintz and Serafini, 1984)

(9)
$$\beta_d = \begin{cases} 1 - \exp\left(-6.68 \frac{w_{d-1}}{w^*}\right), & D_d < 0 \\ 1, & D_d \geq 0 \end{cases}$$

and D_d (mm/day) represents the difference between moisture supplied by precipitation and snow melt and the demand by PE

(10)
$$D_d = M_d + P_d^r - PE_d$$

with PE_d=PE/30. When D_d is negative, it represents a demand for soil moisture; when positive, it represents recharge of soil moisture.

Monthly actual evapotranspiration (AE, mm/month) is

(11)
$$AE = P^r + \sum^{30} M_d - \Delta w - \sum^{30} S_d$$

where Δw is the net change in soil moisture storage over the month (mm) and S_d is the daily surplus moisture. For streamflow calculations, monthly surplus (S, mm) is given by the final term in (11).

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