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Climate, streamflow and water supply in the northeastern United States

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Abstract

Most previous investigations of the impact of potential climatic change on water supply systems have focused on individual systems so that their conclusions only apply to a particular system. Recent advances in computer technology, regional hydrology and our understanding of water supply system behavior allow for examination of the sensitivity of water supply system behavior to potential climatic change in a very general framework. A regional hydroclimatologic model of annual streamflow is developed for the northeastern United States which relates moments of annual streamflow to climatic and geomorphic characteristics at 166 gaging stations. The regional hydroclimatologic streamflow model is then combined with analytic relationships among water supply system storage, reliability, resilience and yield. The sensitivity of various water supply system performance indices such as yield, reliability and resilience are derived as a function of climatic, hydrologic and storage conditions. These results allow us to quantify, in general, the sensitivity of water supply system behavior to changes in the climatologic regime. Case studies for four watersheds in New York and one water supply system in Massachusetts indicate that our simple regional annual modelling approach can reproduce the results of much more detailed site-specific monthly hydroclimatologic modelling approaches. © 1997 Elsevier Science B.V.

1. Introduction

Most previous investigations of the relationship among climate, streamflow and water supply have combined general circulation models of the atmosphere (GCMs), rainfall-runoff models and a reservoir operations model to explore potential impacts of climatic

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change on the behavior of a particular water supply system. Investigations of this type in the northeastern U.S. include, but are not limited to, a study by Wolock et al. (1993) for the Delaware river basin, studies by Lettenmaier et al. (1994) and Kirshen and Fennessey (1995) for the water supply system which services eastern Massachusetts and a study by Fiering and Rogers (1994) for the Colebrook reservoir in Connecticut. Other studies in the northeast which only examine the impact of climatic change on monthly watershed runoff include a study by McCabe and Ayers (1989) for the Delaware River Basin and a study by Tung and Haith (1995) for four watersheds in New York. The literature on the impact of potential climatic change on hydrology and water supply systems is much too large to summarize here; comprehensive reviews may be found in Gleick (1989), Gleick (1990), Waggoner (1990), Chang et al. (1992), Rind et al. (1992), Ballentine and Stakhiv (1993), Leavesley (1994) and Loaiciga et al. (1996). Much of this literature describes the nesting of several detailed simulation models including a GCM, a rainfall-runoff model and sometimes a reservoir operations model for the purpose of performing climatic sensitivity analysis. Though validation procedures exist specifically for testing such nested hydro-climatologic modelling schemes (Klemes, 1986), they are rarely applied, and as a result, the complexity of the models along with associated issues of parameter and model uncertainty render most of the results questionable (Klemes, 1990). This study takes a different approach.

The behavior of many complex water supply systems is controlled primarily by year-to-year variations in hydrology and climate. Such systems, termed over-year systems, can be modelled using an annual time-scale, leading to remarkably simple modelling approaches which exploit both at-site and regional information about climate and streamflow. Vogel and Hellstrom (1988) have documented that an annual time-scale will suffice for modelling both streamflow and water supply system behavior for a system dominated by carry-over or over-year storage requirements. Hydroclimatic models with an annual time-scale are extremely simple when compared to models with monthly, weekly or daily time-scales. The much simpler structure associated with models of annual streamflow allows us to develop more general regional models which apply to a broad geographic region rather than to a single specific river basin, as is the case for monthly, weekly and daily models. The primary objective of this study is to document that a very simple regional hydroclimatologic model of annual streamflow can be used to evaluate the sensitivity of water supply systems to climate, leading to roughly the same results as much more detailed approaches which use monthly, weekly or daily time intervals. Another goal is to provide graphical relations which summarize the impact of changes in climate on water supply system behavior for a broad range of water supply systems throughout the northeastern U.S. analogous to the figures developed by Schaake (1990) for the southeastern U.S. and by Rogers and Fiering (1990) for an arbitrary river basin.

Our approach is empirical because we exploit observed relationships between streamflow and climate in a similar way to the empirical study by Langbein et al. (1949) and the study by Revelle and Waggoner (1983) for the western U.S. which is based on the relations of Langbein et al. (1949). Leavesley (1994) argues that since such empirical approaches only reflect climatic and basin conditions during the time period in which they were developed, it is questionable to extend those relations to climatic conditions different from those used in their development. The same criticism may be raised for any hydrologic model which is not tested for geographical and climatic transposability using the proxy-basin and differential split-sample tests recommended by Klemes (1986). Our approach is to develop an empirical regional hydroclimatologic model for such a broad range of climatic, streamflow and basin characteristics, that it would not be an extrapolation to perturb it with modest variations in climate.

2. Regional hydroclimatological database

Our approach is semi-empirical and our ability to relate streamflow, climate and water supply results from extensive use of regional streamflow, geomorphic and climatic data for the northeastern U.S., hence we begin by describing those information resources. A listing of the streamflow, geomorphic and climatic data may be found in Fennessey (1994) and Bell (1995). All annual streamflow and climatic records were based on water years (1st October–30th September).

2.1. Streamflow database

The streamflow dataset consists of records of daily streamflow at 166 sites located in the northeastern U.S. with drainage areas ranging from less than 2 mi^2 to nearly 7000 mi^2. This dataset is a subset of the Hydroclimatologic Data Network (HCDN) available on CD-ROM from the U.S. Geological Survey (Slack et al., 1993). Fig. 1 illustrates the location of the 166 sites.

The development of the HCDN was a large undertaking which included screening the data in a variety of ways. Streamflow records were reviewed on the basis of the following criteria: (1) availability of data in electronic form, (2) records of lengths in excess of 20 years unless site location is underrepresented, (3) accuracy ratings of records had to be at least "good" as defined by USGS standards, (4) no overt adjustment of "natural" monthly streamflows by flow diversion, augmentation, groundwater pumping, or other forms of regulation, and (5) only measured discharge values are tabulated; no reconstructed or estimated records are used.

Fennessey (1994) performed a non-parametric Mann Kendall trend test to determine whether discernible anthropogenic influences were evident at the 166 streamgage locations. Helsel and Hirsch (1992) describe the application of the Mann–Kendall trend test. The null hypothesis is that each of the 166 flow records do not contain a time trend. Using a 5% level test, the null hypothesis was rejected at only ten (or 6%) of the 166 sites. This result suggests that there is a strong possibility that those sites that failed the Mann–Kendall temporal trend test, did so simply by chance, the result of a type I error. However, such tests should not be taken too seriously, since annual streamflows are known to be serially correlated, and trend tests are known to lack power and yield more type I errors than one expects when applied to serially correlated time-series (von Storch and Navarra, 1995). Lettenmaier et al. (1994) document that significant trends in streamflow are apparent for some regions of the country, and that those trends are not consistent with changes in climatic variables. Hence they may be due to a combination of both climatic and water management effects.

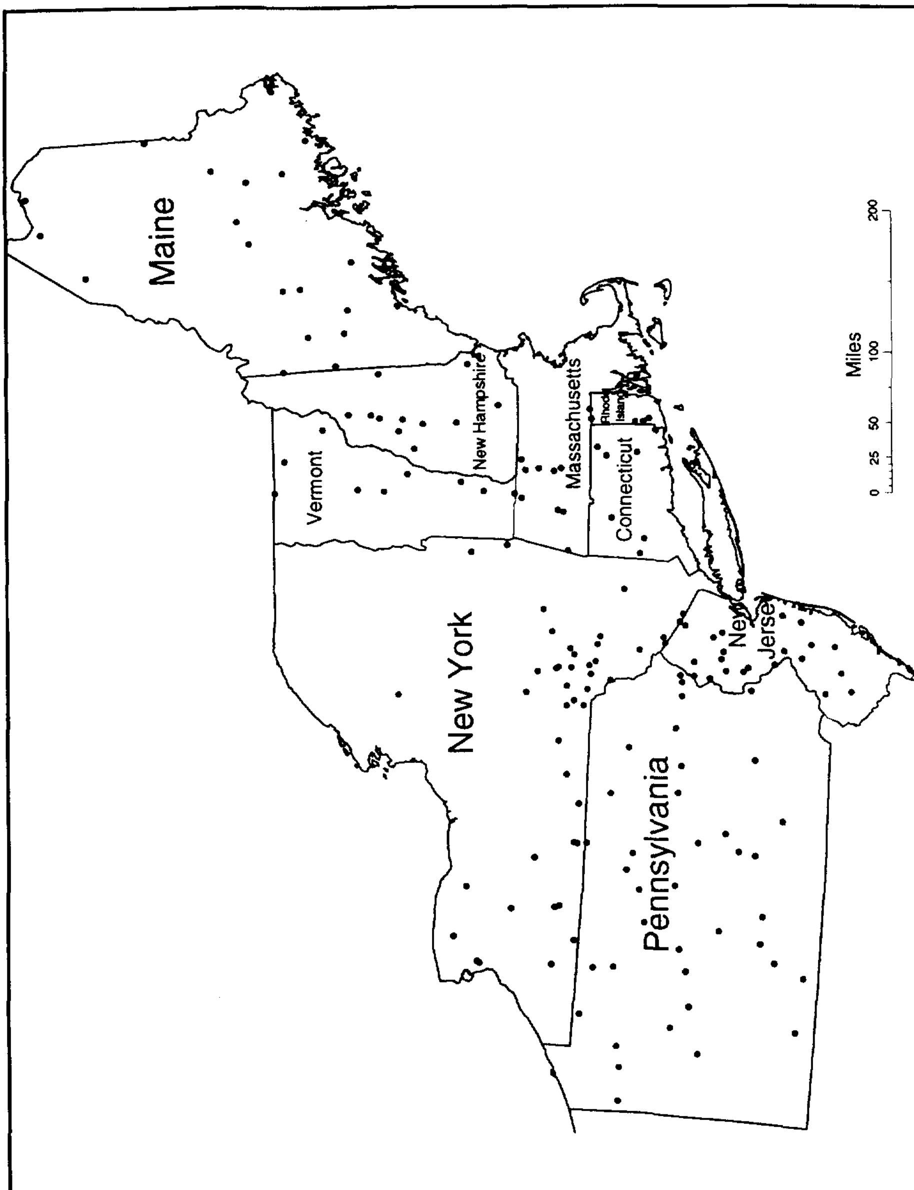

Fig. 1. Location of 166 streamflow gages (from Slack et al., 1993).

2.2. Climatic database

Daily precipitation and temperature data for the period-of-record 1 October 1951 through 30 September 1981, were obtained from the 323 NOAA Summary-of-the-Day climate stations illustrated in Fig. 2. Daily data were then aggregated into annual values for those years with no missing data. Mean annual precipitation P, and temperature T, values were then estimated. The mean values of P and T were interpolated to the 166 streamgage location coordinates using an algorithm developed by Fennessey (1994). Using climatic data from the nearest five climatic stations, annual mean P and T values were interpolated to a given streamgage with interpolation weights proportional to the record length and inversely proportional to the square of the distance between the climatic station and the streamflow gage. Ideally, the average climatic data would be interpolated to the centroid of each watershed, however these coordinates were unavailable. The streamflow and climatic data in this study were subject to detailed quality control and assurance procedures described by Fennessey (1994).

3. Regional regression models for annual streamflow

3.1. Review of literature

Regional hydrologic relationships between climate, geomorphology and streamflow have been developed by many investigators for the purpose of estimating floodflows and lowflow statistics at ungaged sites. For example, regional hydrologic relationships are so well developed in the U.S. for floodflows that a computer program is now available to implement them for all regions in the U.S. (Jennings et al., 1994). Regional hydrologic models for annual streamflow are not nearly as well developed as they are for flood and lowflows. Perhaps the earliest regional hydroclimatologic model of annual streamflow is the graphical relationship between annual runoff, annual precipitation and annual temperature for the U.S. introduced by Langbein et al. (1949). Orsborn (1974) documents a graphical relation between average annual streamflow, precipitation and drainage area for stations near Vancouver, Washington. Other regional hydrologic models of annual streamflow in the U.S. include regional regression models developed by Lull and Sopper (1966) and Johnson (1970) for the New England region, Hawley and McCuen (1982) in the western U.S. and Thomas and Benson (1970) for selected regions throughout the U.S.. Other regional hydrologic models of annual runoff include runoff maps introduced by Church et al. (1995) for the northeastern U.S. and other regions of the U.S. (see Church et al., 1995, for other citations). Regional models of annual streamflow have received considerably more attention from a global perspective as evidenced from the books by Kalinin (1971), McMahon et al. (1992) and Finlayson and McMahon (1992).

3.2. Multivariate hydroclimatologic regression models of annual streamflow

This section describes the development of regional hydroclimatologic relationships for

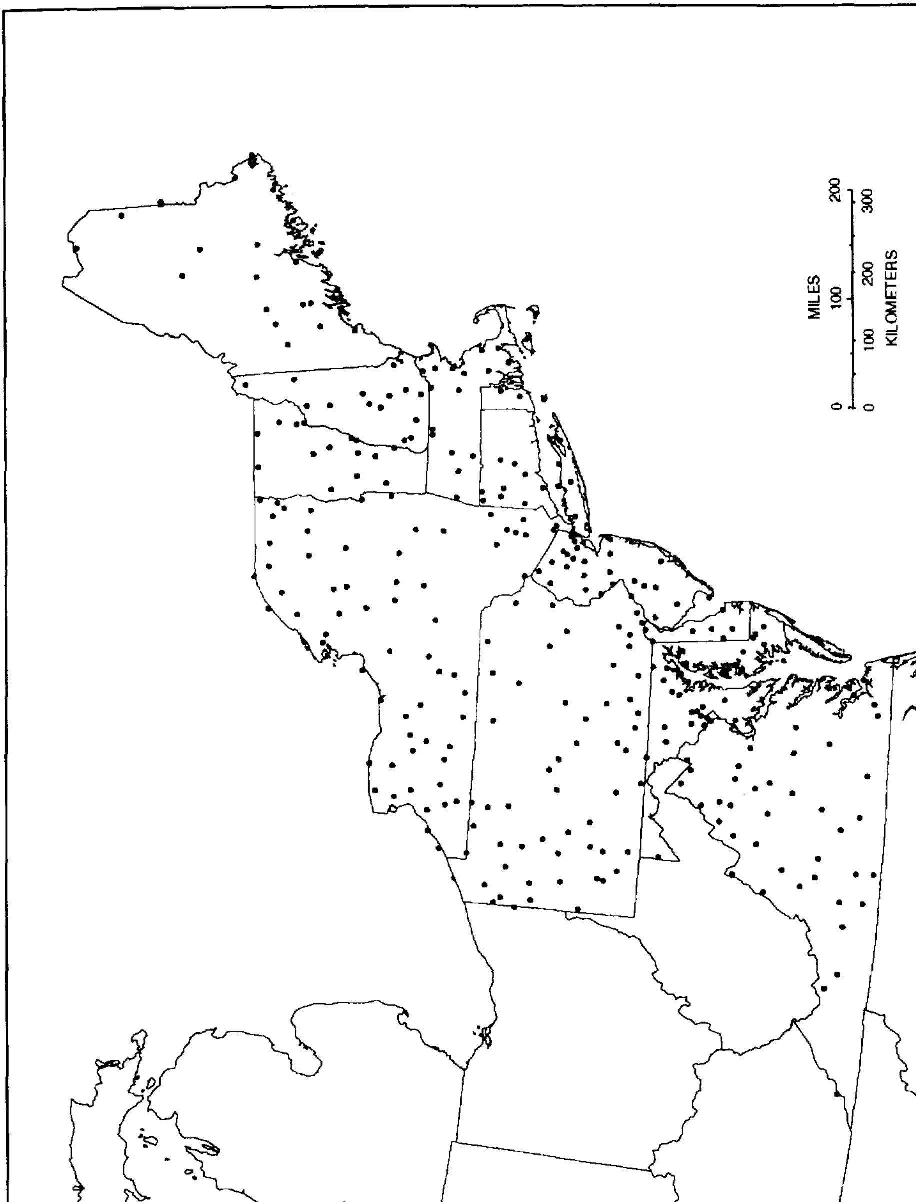

Fig. 2. Location of 323 NOAA Summary-of-the-Day climate observatories.

the mean, μ, and standard deviation, σ, of annual streamflow. Regional regression procedures led to multivariate relationships between μ and σ and various independent climatic and geomorphic variables. The climatic variables included: mean annual precipitation P (in inches) and mean annual temperature T (in °F). The independent geomorphic variables included watershed area A (in square miles), basin relief H (in feet) and numerous others which were not found to significantly improve the multivariate regressions. Basin relief was approximated as the difference between the mean basin elevation and the streamgage elevation. All basin characteristics were obtained (or computed) from the basin characteristics file included with the HCDN (Slack et al., 1993).

Ordinary least-squares (OLS) and weighted least-squares (WLS) multivariate linear regression procedures were used to fit models of the form

$$\ln(Y) = c_0 + c_1\ln(X_1) + c_2\ln(X_2) + \cdots + c_n\ln(X_n) + \epsilon \tag{1}$$

which, when transformed into linear space, yields the equivalent model

$$Y = e^{c_0}X_1^{c_1}X_2^{c_2}\cdots X_n^{c_n}e^{\epsilon} \tag{2}$$

where Y represents the dependent variable, X_i represents the various independent variables, and the ϵ values are normally distributed errors with zero mean and variance, σ_ϵ^2. With OLS regression procedures, each basin is treated equally, implying that all observations of the dependent variable are "equally reliable". Tasker (1980) introduced WLS regression to account for the fact that each basin has a different streamflow record length leading to estimates of the dependent variable with varying degrees of reliability. Using WLS regression, Tasker (1980) shows that the weight w_j assigned to each set of observations of the dependent variable and its associated independent variables is proportional to the reciprocal of the variance of an estimate the dependent variable. Ignoring model error, the reciprocal of the variance of $\bar{x}$ and s, sample estimates of the dependent variables, μ and σ, are given by $\mathrm{Var}^{-1}(\bar{x}) = n/\sigma^2$ and $\mathrm{Var}^{-1}(s) \cong (2n)/\sigma^2$. Since both variances are proportional to the record length n, we use the simple weighting scheme:

$$w_i = \frac{n_i}{\sum_{i=1}^{s} n_i} \tag{3}$$

where w_i is weight for site i; n_i is length of streamflow record, in years, for site i and s is number of sites, 166. Eq. (3) places a weight on each dependent variable in proportion to the record length used to estimate those variables.

Table 1 and 2 summarize the OLS and WLS regression models for the mean annual streamflow, μ, and the standard deviation of annual streamflow, σ, respectively. Tables 1 and 2 document that OLS regression models for the natural logarithm of μ and σ are very precise with adjusted R^2 values of 99.1 and 98.8 respectively. The OLS models are only included for comparative purposes, because they allow computation of the goodness-of-fit statistic R^2 which cannot be computed for the WLS models. Table 1 reports three different WLS regression models for μ. In row 2 of Table 1, the constant term, a, was not significantly different from zero using a 5% significance level t-test. Tables 1 and 2 report the t-ratios of each estimated model parameter in parentheses, where the critical value for the t-ratio is equal to 1.98 using a 5% significance level. Dropping the constant term leads to

Table 1
Summary of regression models for the mean annual streamflow: $\hat{\mu} = aA^{b}P^{c}T^{d}H^{e}$

Model	a	b Area	c Precipitation	d Temperature	e Relief	SE%	r_ϵ	R^2 (adj)
OLS	0.139 (−2.43)	0.987 (118.4)	1.306 (11.50)	−0.752 (−4.79)	0.092 (5.06)	13.75	0.9947	99.1
WLS	0.242 (−1.82)	0.986 (115.39)	1.212 (10.93)	−0.791 (−5.27)	0.085 (4.76)	13.79	0.9956	
WLS	1	0.982 (118.26)	1.095 (12.03)	−1.013 (−11.57)	0.068 (4.44)	14.03	0.9939	
WLS	1	1.000 (129.22)	1.177 (12.53)	−1.005 (−10.90)	0	14.84	0.9874	

The values in parentheses are t-ratios of model parameters. Standard error of estimates are given by SE% = $100[\exp(\sigma_\epsilon^2)-1]^{1/2}$. OLS stands for ordinary least squares regression. WLS stands for weighted least squares regression.

the second WLS model in row 3 of Table 1 which we recommend as the most precise of the four models in Table 1. This model is:

$$\hat{\mu} = A^{0.982}P^{1.095}T^{(-1.013)}H^{0.068} \tag{4}$$

where $\hat{\mu}$ is WLS regression estimate of mean annual streamflow in cfs; A is basin area in square miles; P is mean annual precipitation in inches; T is mean annual temperature in °F; and H is basin relief in feet. All estimated model parameters in Eq. (4) are very precise as evidenced by their large t-ratios shown in parentheses in Table 1.

Table 2 documents two OLS models for σ in rows 1 and 2. In this case, the coefficient

Table 2
Summary of regression models for the standard deviation of annual streamflow: $\hat{\sigma} = aA^{b}P^{c}T^{d}H^{e}$

Model	a	b Area	c Precipitation	d Temperature	e Relief	SE%	r_ϵ	R^2(adj)
OLS	0.0050 (−5.89)	0.971 (104.51)	1.147 (9.06)	−0.033 (−0.19)	0.069 (3.40)	15.35	0.9896	98.8
OLS	0.0043 (−10.64)	0.971 (105.48)	1.147 (9.09)	0	0.071 (3.85)	15.35	0.9897	98.9
WLS	0.0062 (−10.35)	0.966 (105.25)	1.056 (8.77)	0	0.072 (4.07)	15.39	0.9891	
WLS	0.0124 (−9.20)	0.978 (108.79)	0.976 (7.92)	0	0	16.04	0.9875	

The values in parentheses are t-ratios of model parameters. Standard error of estimates are given by SE% = $100[\exp(\sigma_\epsilon^2)-1]^{1/2}$. OLS stands for ordinary least squares regression. WLS stands for weighted least squares regression.

for annual temperature is not significant using a 5% significance level, hence we drop temperature from the model. Again, OLS models are only provided for comparative purposes because they allow for estimation of the adjusted R^2. Row 3 of Table 2 summarizes our recommended regional regression model for estimating σ at ungaged sites in this region. This model is:

$$\hat{\sigma}=0.0062A^{0.966}P^{1.056}H^{0.072} \tag{5}$$

where $\hat{\sigma}$ is the WLS regression estimate of the standard deviation of annual streamflow (in cfs) and A, P and H are as defined earlier in Eq. (4).

Perhaps the best summary of the goodness-of-fit of the regression estimators $\hat{\mu}$ and $\hat{\sigma}$ in Eqs. (4) and (5) is provided in Figs 3 and 4, respectively, which compare the regression estimates of both μ and σ with the original sample estimates upon which those regressions were based. Figs 3 and 4 verify that Eqs. (4) and (5) are remarkably precise estimators for the entire northeastern U.S.

Tables 1 and 2 also document the normal probability plot correlation coefficient, r_ϵ, which provides a measure of the normality of the regression model residuals in Eq. (1). Using a 5% significance level test, the critical value of r_ϵ is approximately 0.987 for n = 166 sites (Vogel, 1986). Using this test of normality, one cannot reject the null hypothesis of normality for the residuals for all regression models reported in Tables 1 and 2. Bell (1995) uses numerous influence statistics to document that none of the 166 sites used to develop the regression models described in Tables 1 and 2 exhibit either unusual influence or leverage.

In the last row of Tables 1 and 2, simplified WLS models are documented which do not employ the geomorphic variable H. These models are:

$$\hat{\mu}=A^{1.000}P^{1.177}T^{(-1.005)} \tag{6}$$

$$\hat{\sigma}=0.0124A^{0.978}P^{0.976} \tag{7}$$

with both $\hat{\mu}$ and $\hat{\sigma}$ in cfs. These simplified models are used in the climatic sensitivity analyses which follow.

Eqs. (4)–(7) may be used to describe the moments of annual streamflow throughout the northeastern U.S., with the exception of coastal areas of New Jersey, Connecticut, Rhode

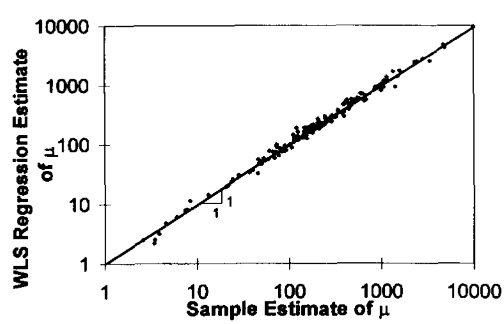

Fig. 3. Comparison of WLS regression estimates of mean annual streamflow, μ (Eq. (4)), with original sample estimates of μ at 166 sites.

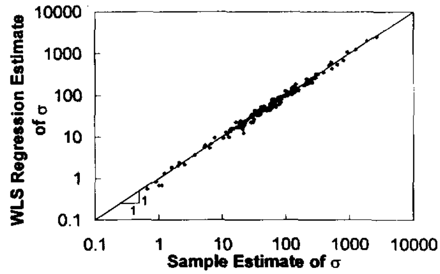

Fig. 4. Comparison of WLS regression estimates of standard deviation of annual streamflow, σ, (Eq. (5)) with original sample estimates of σ at 166 sites.

Island and Massachusetts, where watershed runoff is dominated by groundwater mechanics.

3.3. Summary of regional hydroclimatologic model of annual streamflow

Vogel et al. (1995) perform numerous statistical tests to show that the year-to-year variability and persistence of streamflow in the northeastern U.S. is remarkably homogeneous. Vogel et al. (1995) use hypothesis tests based on L-moment ratios (Hosking, 1990), L-moment diagrams (Hosking, 1990) and probability plot correlation coefficient tests (Vogel, 1986) to show that annual streamflow in the northeastern United States is approximately normally distributed. Vogel et al. (1995) perform a simple experiment to show that observed time-series of annual streamflow at these 166 sites could not be distinguished from synthetic, normally distributed, annual streamflow series generated with a fixed lag-one serial correlation coefficient of $\rho = 0.19$ across the entire geographic region illustrated in Figs 1 and 2.

In summary, the regional hydroclimatologic model of annual streamflow is described by Eqs. (4) and (5) or Eqs. (6) and (7) along with the assumption that annual streamflow follows an AR(1) normal model with a fixed lag-one serial correlation coefficient of $\rho =$ 0.19. This model applies to basins in northeastern U.S. with values of drainage area in the range 1.5–6780 mi^2, values of annual average precipitation in the range 31–63 inches, and values of annual average temperature in the range 35.4–54.5°F.

4. Comparison of annual watershed model with monthly watershed models: four case studies in New York

Tung and Haith (1995) report the impact of climatic change on monthly and annual watershed runoff for four watersheds in New York. The following section evaluates the ability of the model reported in Eq. (6) to reproduce the results of their more detailed studies. Tung and Haith (1995) simulate monthly streamflow using a daily water balance model fed by daily precipitation and temperature measurements for the four watersheds summarized in Table 3. Table 3 reproduces the drainage areas and historical climate from Table 1 in Tung and Haith (1995). Table 3 also compares the historical streamflow statistics reported in Table 4 of Tung and Haith (1995) with estimates of the mean annual streamflow obtained from Eq. (6) in this study. The historical climatic information summarized by Tung and Haith (1995) in their Table 1, and reproduced here in our Table 3, is based on the periods 1972–1980, 1974–1979, 1972–1975 and 1972–1977 for the Fall Creek, Oatka Creek, Wappinger Creek and W. Branch Delaware River watersheds, respectively. Our regression model is based on climatic conditions over the longer period 1950–1980, hence it is not surprising to see modest differences. Overall, the agreement between our regional regression formula and their results is acceptable, though this is to be expected from our earlier results in Fig. 3.

The main objective of this comparison is to evaluate the ability of our regional regression models to predict the sensitivity of streamflow to changes in climatic conditions. For that purpose, we use Eq. (6) to reproduce the GCM scenarios performed by Tung and Haith (1995). Tung and Haith (1995) predict streamflow under future climatic conditions using monthly temperature and precipitation data obtained from mean monthly climatic predictions for $2 \times CO_2$ conditions, from the Goddard Institute for Space Studies (GISS) and the Geophysical Fluid Dynamics Laboratory (GFDL) models. Fig. 5 compares the streamflow under current (historical) and future ($2 \times CO_2$) climatic conditions predicted by Tung and Haith (1995) (Table 4) and by this study. The results for this study were obtained by inserting the annual temperature and precipitation values reported by Tung and Haith (1995) for the GISS and GFDL scenarios, into Eq. (6). Fig. 5 illustrates that our methodology compares favorably with the results of the more detailed monthly modelling approach employed by Tung and Haith (1995). Our procedure predicts reductions in annual streamflow under future climatic conditions of about the same magnitude as the procedure employed by Tung and Haith (1995). Naturally since Tung and Haith (1995) model monthly streamflow rather than annual streamflow, their study provides much more detailed information than this study can, regarding the seasonal impact of climatic change on precipitation, evapotranspiration, surface runoff and streamflow. Nevertheless, Fig. 5

Table 3
Description of historical annual climate and streamflow for four New York watersheds (from Tung and Haith, 1995)

Watershed name	Drainage area (mi^2)	Mean annual precipitation (inches)	Mean annual temperature (°F)	Average Annual Streamflow	
				Tung and Haith (1995) (inches)	This study Eq. (6) (inches)
Fall Creek	130.0	38.97	46.2	19.7	21.5
Oatka Creek	200.0	33.85	47.7	14.2	17.6
Wappinger Creek	181.0	39.75	48.9	19.7	20.8
W. Branch Delaware River	333.0	43.30	45.7	23.2	24.6

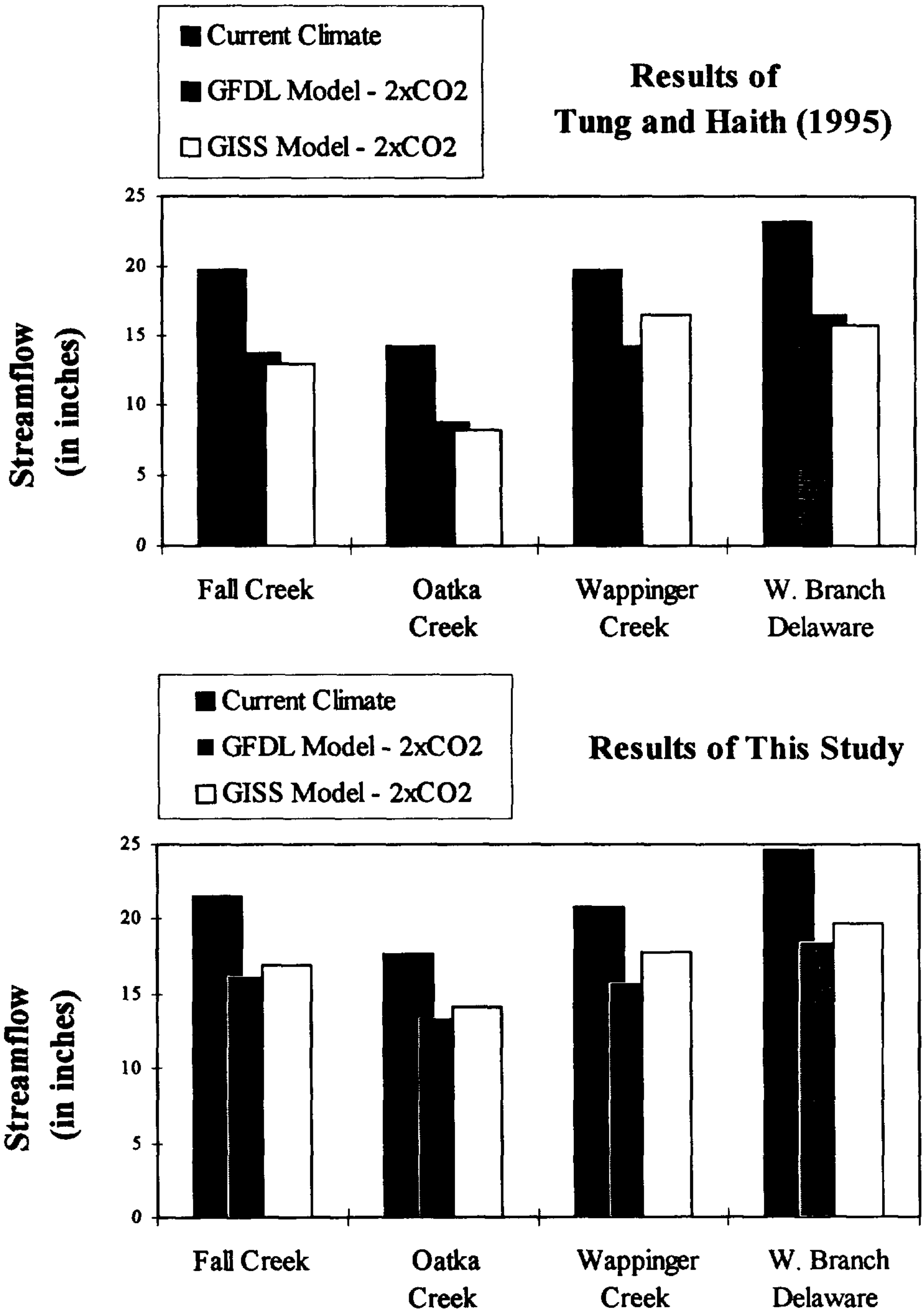

Fig. 5. Comparison of our approach for predicting average annual streamflow under future climate conditions with the approach used by Tung and Haith (1995) for four watersheds in New York.

documents that the regional hydroclimatologic modelling approach developed here can provide results comparable to those of much more detailed modelling approaches, at the annual level.

5. Storage–reliability–resilience–yield relationships

Vogel and Bolognese (1995) and others have introduced analytic relationships which approximate the behavior of water supply systems dominated by over-year or carry-over storage requirements. Using four case-studies of reservoir systems located in the north-eastern U.S., Vogel et al. (1995) show that SRRY relationships can be very useful for comparing and understanding the behavior of complex multiple reservoir systems. Similar to this study, the Vogel et al. (1995) study combines regional hydrologic regression models of annual streamflow with an analytic SRRY model for the purpose of generalizing water supply system behavior in the northeastern U.S. This study differs from Vogel et al. (1995) in two important ways: (1) more detailed and accurate regional hydrologic models are derived here which include the impacts of precipitation and temperature; (2) detailed comparisons are made regarding the impact of potential climatic change on water supply systems.

Vogel et al. (1995) document that the index, m, introduced by Hazen (1914), is useful for classifying the behavior of water supply systems. The index, m, is defined as:

$$m=\frac{(1-\alpha)\mu}{\sigma}=\frac{(1-\alpha)}{C_v}=\frac{\mu-Y}{\sigma} \tag{8}$$

where Y is the average annual demand, α ($\alpha = Y/\mu$) is the annual demand as a fraction of the mean annual inflow to the reservoir, μ and σ are the mean and standard deviation of the annual inflows, and C_v is the coefficient of variation of the annual inflows ($C_v = \sigma/\mu$). As long as the index, m, is in the range, $0 < m < 1$, the system will be dominated by year-to-year or carry-over storage requirements. Similarly, systems with $m > 1$ will be dominated by within-year storage requirements. Actual systems will be subject to both within-year and over-year variations in storage, hence there is no unique division between these two classes of behavior. In general, within-year systems are expected to refill each year whereas over-year systems contain long multi-year draw-down periods and are seldom full. In this study, we only consider the behavior of systems dominated by carry-over storage, hence we assume $0 < m < 1$.

Vogel and Bolognese (1995) summarize analytic functions which describe the relationship among reservoir system storage, yield, reliability and resilience for systems fed by AR(1) normal inflows. Vogel and Bolognese (1995), in Appendix A of their paper, summarize relations among reservoir system storage capacity S, planning horizon N, index m, lag-one serial correlation of the inflows ρ, and N-year no-failure reliability R_N, in the form

$$S/\sigma=f(m,N,\rho,R_N) \tag{9}$$

for systems fed by AR(1) normally distributed inflows, with m defined in Eq. (8). Eq. (9) was developed from Monte-Carlo experiments which routed AR(1) streamflows through a reservoir using the sequent peak algorithm and is too complex to reproduce here. Vogel and Bolognese (1995), in Eq. (17) of their paper show that N-year no-failure reliability R_N can be related to annual reliability R_a using

$$R_N = R_a[1 - r(1 - R_a^{-1})]^{N-1} \tag{10}$$

where r is an index of resilience which can be estimated using

$$r = \Phi\left[\frac{1}{\sqrt{1-\rho^2}}\left(m - \frac{\rho}{\Phi(-m)\exp(\frac{m^2}{2})\sqrt{2\pi}}\right)\right] \tag{11}$$

where $\Phi(\text{arg})$ denotes the cumulative normal density function applied at arg. The index, r, is defined as the probability that the reservoir system will be able to provide the stated yield Y, in a year following a failure year. Here a failure year is one in which the reservoir system is unable to deliver its prespecified yield Y. No-failure reliability R_N is the probability that a given system will provide a constant yield Y, without failure, over an N-year period. Annual reliability R_a is the steady-state probability, in a given year, that the reservoir system will deliver the stated yield. Therefore, the reservoir system fails to deliver its yield $(1 - R_a)$% of the time.

Vogel and Bolognese (1995) document that Eqs. (8)–(11) agree well with theoretical storage–reliability–yield relationships introduced by other investigators for over-year systems. Vogel et al. (1995) document that Eqs. (8)–(11) are useful for describing the behavior of the water supply systems which service New York City; Providence, Rhode Island; Springfield, Massachusetts and the Boston, Massachusetts, metropolitan areas. Vogel and Hellstrom (1988) also show that Eqs. (8)–(10) compare favorably with the application of a detailed monthly reservoir simulation model for the Boston metropolitan water supply system.

6. Sensitivity of water supply system behavior to climate

One goal of this study is to document that a simple annual regional hydroclimatologic model when linked with SRRY relations can be used to determine the sensitivity of complex reservoir systems and river basins to potential climatic change. Another goal is to document that our simple annual modelling approach is comparable to much more costly and complex monthly operations studies. The following sections evaluate the ability of our approach to meet these goals.

6.1. Comparison of annual and monthly hydroclimatologic models of water supply: a case study

Kirshen and Fennessey (1995) describe a recent investigation of the impact of climatic change on the water supply system which serves the Boston metropolitan area. Kirshen and Fennessey (1995) used the Sacramento soil moisture accounting model, the National Weather Service River Forecast System (NWSRFS) snow accumulation and ablation model, modified Penman-equation estimates of reservoir evaporation and Penman–Monteith estimates of potential evapotranspiration to generate monthly streamflows. The simulated streamflows were routed through a monthly reservoir operations model developed for the Massachusetts Water Resources Authority (MWRA, 1986) and summarized by Vogel and Hellstrom (1988). The following section provides a comparison of our approach with the approach taken by Kirshen and Fennessey (1995).

6.2. Description of case study: the MWRA water supply system

The MWRA system serves 47 communities and approximately 2.5 million residents primarily in the Boston metropolitan area located in eastern Massachusetts. MWRA (1986), Vogel and Hellstrom (1988) and Kirshen and Fennessey (1995) include detailed descriptions of the water supply system, hence we only provide a cursory overview here. The water supply system, is composed of three watersheds: the Swift River–Quabbin Reservoir watershed; the Ware River watershed; and the Nashua River–Wachusett Reservoir watershed. The three watersheds encompass an area of approximately 389.9 mi^2. The Quabbin and Wachusett Reservoirs have an active storage of approximately 265 billion gallons. The three watersheds are connected by a single underground tunnel known as the Quabbin Aqueduct. Several previous "safe-yield" investigations, summarized in MWRA (1986) and Vogel and Hellstrom (1988) have reported that the system could deliver a 50-year no-failure yield of approximately 300 mgd, if the 1960s drought were to repeat itself again. Vogel and Hellstrom (1988) use a stochastic streamflow model to show that considering sampling uncertainty associated with the observed streamflow record, the 50-year no-failure yield could easily range from 232 to 370 mgd. Recent efforts by the MWRA including price increases, leak detection and repair, public education and other programs, have led to significant reductions in water use, so that current demands are approximately 260 mgd.

6.3. Description of base case or historical system yield simulations

Kirshen and Fennessey (1995) employed monthly temperature, precipitation, incident solar radiation, wind speed and relative humidity data over the historical period of record 1950–1979 to simulate streamflows entering the reservoirs. The NWSRFS model requires monthly estimates of precipitation and temperature. The Sacramento soil moisture accounting model requires precipitation and land surface evapotranspiration ET, with the ET values estimated from the Penman–Monteith model which, in turn, required monthly estimates of temperature, solar radiation, wind speed, relative humidity and vegetation specific parameters. The simulated streamflows were then routed through the monthly reservoir operations model resulting in an estimated base case yield of 305 mgd over the 30-year planning period with an estimated annual reliability $R_a = 0.985$. By comparison, MWRA (1986) and Vogel and Hellstrom (1988) obtained a yield of about 295 mgd with $R_a = 0.985$. When no monthly failures are allowed, the historical yield rises to about 300 mgd.

In order to reproduce the base case simulation performed by Kirshen and Fennessey (1995) we assume a yield $Y = 305$ mgd with annual reliability $R_a = 0.985$, planning horizon $N = 30$ years, average annual historical precipitation $P = 41.3$ inches, and average annual temperature $T = 48.8$°F in Eqs. (6)–(11). We essentially calibrate our model to reproduce, exactly, the base case simulation performed by Kirshen and Fennessey (1995). We calibrate our model to theirs for the base case scenario, so that later comparisons of the sensitivity of our modelling approach to climatic change with their approach are not confounded by having modeled different base case scenarios. Our model is calibrated by noting that it requires an assumption about required downstream releases and diversions from the three watersheds. Let R be defined as the sum of those releases, then the net annual average inflow μ in Eq. (8) is defined by $\mu = \hat{\mu} - R$, with $\hat{\mu}$ given in Eq. (6). Solving Eqs. (6)–(11) leads to $R = 78.3$ mgd and a net annual inflow $\mu = 325.8$ mgd. Interestingly, Vogel and Hellstrom (1988) used a completely independent method to obtain a net annual inflow of $\mu = 328$ mgd for this system.

6.4. Description of general circulation model scenarios

General circulation models (GCMs) are detailed numerical models of atmospheric circulation (see Loaiciga et al., 1996, for a recent review). GCM model output is used here to provide estimates of annual average temperature and precipitation under future potential climatic conditions resulting from a doubling of atmospheric carbon dioxide. Kirshen and Fennessey (1995) used GCM model output to provide estimates of monthly temperature and precipitation as well as monthly incident solar radiation, wind speed and specific humidity under future potential climatic conditions. The GCM model runs employed by Kirshen and Fennessey (1995) include runs from GISS made in 1982; GFDL runs from 1988; Oregon State University (OSU) runs from 1984 and 1985, and the United Kingdom Meteorological Office (UKMO) 1986 runs. Since the focus of this section is on a comparison of our approach with that of Kirshen and Fennessey (1995), the fact that the GCM model scenarios are outdated is unimportant.

Table 4 compares the average annual values of precipitation P and temperature T for the base case (current) climate and the changed climate assuming a doubling of atmospheric carbon dioxide ($2 \times CO_2$). All of the GCM models predict a significant increase in T, whereas there is considerable disagreement among the models in terms of the direction of changes in P. The GISS and GFDL models predict a drop in precipitation while the OSU and UKMO models predict a considerable increase in precipitation.

Table 4
Comparison of base case climate and $2 \times CO_2$ GCM climates for the northeastern United States (from Kirshen and Fennessey, 1995)

GCM	Mean annual precipitation P (inches)	Mean annual temperature T (°F)
GISS	39.9	55.4
GFDL	38.0	57.6
OSU	46.8	54.4
UKMO	49.4	63.7
Base Case	41.3	48.8
Range of values at 166 basins in northeast	31–63	35.4–54.5

Also shown in Table 4 are the range of historical average annual values of P and T corresponding to the 166 basins used in this study. The hydroclimatologic models summarized in Eqs. (6) and (7) are based on a range of climatic fluctuation in P which is far in excess of that reported by these four GCMs, therefore no extrapolation of Eqs. (6) and (7) is required to evaluate changes in streamflow which result from changes in P. The values of T predicted by the GISS, GFDL and UKMO models lie outside the range of current climates observed in the northeast. Since the GFDL and UKMO scenarios involve temperature increases which are significantly outside the range of values of T observed for these basins, extrapolation of the regressions in Eqs. (6) and (7) will be required to evaluate the impact of T on streamflow for those scenarios. When regression equations are extrapolated, the results are suspect.

6.5. Comparison of system yields predicted by this study and by Kirshen and Fennessey (1995)

The values of precipitation P, and temperature T, in Table 4 are used in Eqs. (6) and (7) to estimate the mean, μ, and standard deviation, σ, respectively, of the annual streamflows under a $2 \times CO_2$ climate. Those values of μ and σ are used in Eqs. (8)–(11) to obtain system yields corresponding to the altered climatic conditions and the results are reported in Table 5 and Fig. 6. The agreement is generally good, especially considering the simplicity of our modelling approach. With the exception of the UKMO model, our modelling results agree in both the direction and magnitude of the changes in yield. Kirshen and Fennessey (1995) report an increase in yield under the UKMO scenario whereas we report a decrease in yield. In this unusual circumstance, we actually trust our simple modelling procedure more than the more complex one employed by Kirshen and Fennessey (1995). This is because the relative humidity values predicted by the UKMO model are unrealistically high (see Fennessey and Kirshen, 1994), leading to unrealistically low estimates of evaporation and evapotranspiration; hence the increased yields reported by Kirshen and Fennessey (1995). Our modelling approach is not sensitive to spurious changes in relative humidity.

The best agreement is obtained with the GISS and OSU models. This is due to the fact that little or no extrapolation of the regression Eqs. (6) and (7) was required to evaluate the

Table 5
Comparison of yields estimated by this Study and by Kirshen and Fennessey (1995)

GCM	Yield (mgd)		Difference (%)
	This study	Kirshen and Fennessey (1995)	
GISS	244	236	+3.4
GFDL	213	173	+23
OSU	318	328	−3
UKMO	278	421	−34
Base Case	305.0	305.0	0

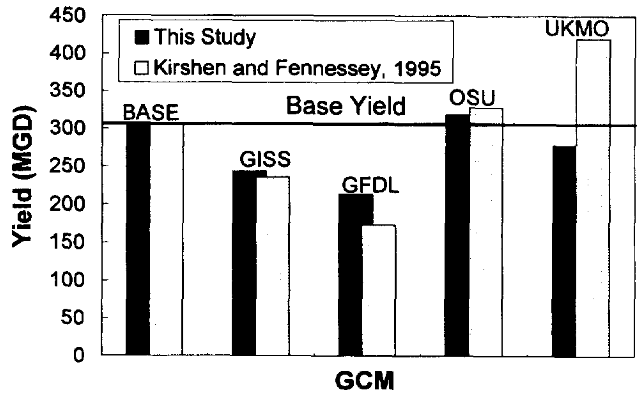

Fig. 6. Comparions of MWRA system yields predicted by this study and Kirshen and Fennessey (1995) for a 2 × CO_2 climate.

impact of changes in climatic regime (temperature) on streamflow. As discussed earlier, Eqs. (6) and (7) are based on a very broad range of values of P and T.

6.6. The general sensitivity of water supply yield to changes in climate

Since the hydroclimatologic and water supply behavior of river basins is described in analytic terms (Eqs. (6)–(11)) it is possible to derive expressions which describe the sensitivity of water supply system yield or resilience to such inputs as precipitation and temperature. This is accomplished using the chain rule. For example, the sensitivity of system yield Y to changes in annual average precipitation is obtained by deriving $\mathrm{d}Y/\mathrm{d}P$. Rearranging Eq. (8) to obtain $Y = \mu - m\sigma$, and applying the chain rule yields

$$\frac{\mathrm{d}Y}{\mathrm{d}P} = \left[\frac{\partial Y}{\partial m}\frac{\partial m}{\partial P}\right] + \left[\frac{\partial Y}{\partial \mu}\frac{\partial \mu}{\partial P}\right] + \left[\frac{\partial Y}{\partial \sigma}\frac{\partial \sigma}{\partial P}\right] \tag{12}$$

Noting that $\partial Y/\partial m = -\sigma$, $\partial Y/\partial \mu = 1$ and $\partial Y/\partial \sigma = -m$ leads to

$$\frac{dY}{dP} = \left[-\sigma\frac{\partial m}{\partial P}\right] + \left[\frac{\partial \mu}{\partial P}\right] - \left[m\frac{\partial \sigma}{\partial P}\right] \tag{13}$$

with $\partial\mu/\partial P$ and $\partial\sigma/\partial P$ easily derived from Eqs. (6) and (7), respectively. The challenge was to derive $\partial m/\partial P$ using the fact that

$$\frac{\partial m}{\partial P} = \frac{\partial m}{\partial S}\frac{\partial S}{\partial P} = \frac{1}{\frac{\partial S}{\partial m}}\frac{\partial S}{\partial P} \tag{14}$$

where the terms $\partial S/\partial m$ and $\partial S/\partial P$ are obtained from chain rule calculations similar to Eq. (12) using the analytic relations for S described by Eq. (9) and given in Vogel and Bolognese (1995) in their Appendix A. The details of these additional derivations are given in Bell (1995).

To reduce the number of dimensions in our evaluation of yield sensitivity $\mathrm{d}Y/\mathrm{d}P$, we derive the yield sensitivity, per unit drainage area, which we denote $d(Y/A)/dP$. Since the regression equations for μ (Eq. (6)) and σ (Eq. (7)) are both approximately linear in drainage area A, there is no loss of generality in examining $d(Y/A)/dP$ rather than dY/dP. Fig. 7 illustrates the sensitivity of water supply yield to changes in precipitation $d(Y/A)/dP$ for various values of precipitation and temperature. Fig. 7 assumes an annual reliability $R_a = 0.99$ and a serial correlation of annual flows $\rho = 0.2$. As expected, dY/dP is always positive, hence yield tends to increase as the level of development α increases and as precipitation increases. However, Fig. 7 illustrates that for a given level of development, increases in precipitation produce greater increases in yield at lower temperatures than at higher temperatures. This is due to the increased evaporation (E) and evapotranspiration (ET) which results from higher temperatures. Each curve in Fig. 7 represents a fixed climate or a fixed value of T or P. Fig. 7 illustrates that, for any given climate (curve), an increase in precipitation will lead to higher yields as the level of development increases because less reservoir spillage occurs as the level of development increases.

The same approach was used to derive an expression for the sensitivity of yield to changes in temperature $d(Y/A)/dT$. Again applying the chain rule to $Y = \mu - m\sigma$ leads to

$$\frac{dY}{dT} = \left[\frac{\partial Y}{\partial m}\frac{\partial m}{\partial T}\right] + \left[\frac{\partial Y}{\partial \mu}\frac{\partial \mu}{\partial T}\right] + \left[\frac{\partial Y}{\partial \sigma}\frac{\partial \sigma}{\partial T}\right] \tag{15}$$

with

$$\frac{\partial m}{\partial T} = \frac{\partial m}{\partial S}\frac{\partial S}{\partial T} \tag{16}$$

From Eqs. (7) and (9), we obtain $\partial\sigma/\partial T = 0$ and $\partial S/\partial T = 0$, hence Eq. (15) reduces to

$$\frac{dY}{dT} = \left[\frac{\partial Y}{\partial \mu}\frac{\partial \mu}{\partial T}\right] = -1.005A\frac{P^{1.177}}{T^{2.005}} \tag{17}$$

Fig. 8 illustrates the sensitivity of water supply yield to changes in temperature $d(Y/A)/dT$ for various values of precipitation and temperature. In this case, the yield sensitivity to temperature is always negative because increases in temperature produce higher ET leading to decreases in yield. Fig. 8 illustrates that for a given climate, yield sensitivity to temperature is constant for all levels of development, as is also shown in Eq. (17). This is because temperature only influences the mean annual inflow and not the variability of the inflows (see Eqs. (6) and (7)) in this region.

6.7. The general sensitivity of water supply system resilience to changes in climate

In this and other studies, resilience, r (Eq. (11)), is defined as the probability that a reservoir system will deliver its stated yield in a year following a failure of the system to provide its stated yield. Resilience is a measure of the ability of a reservoir/watershed system to recover from a failure. Derivations similar to those outlined in the previous

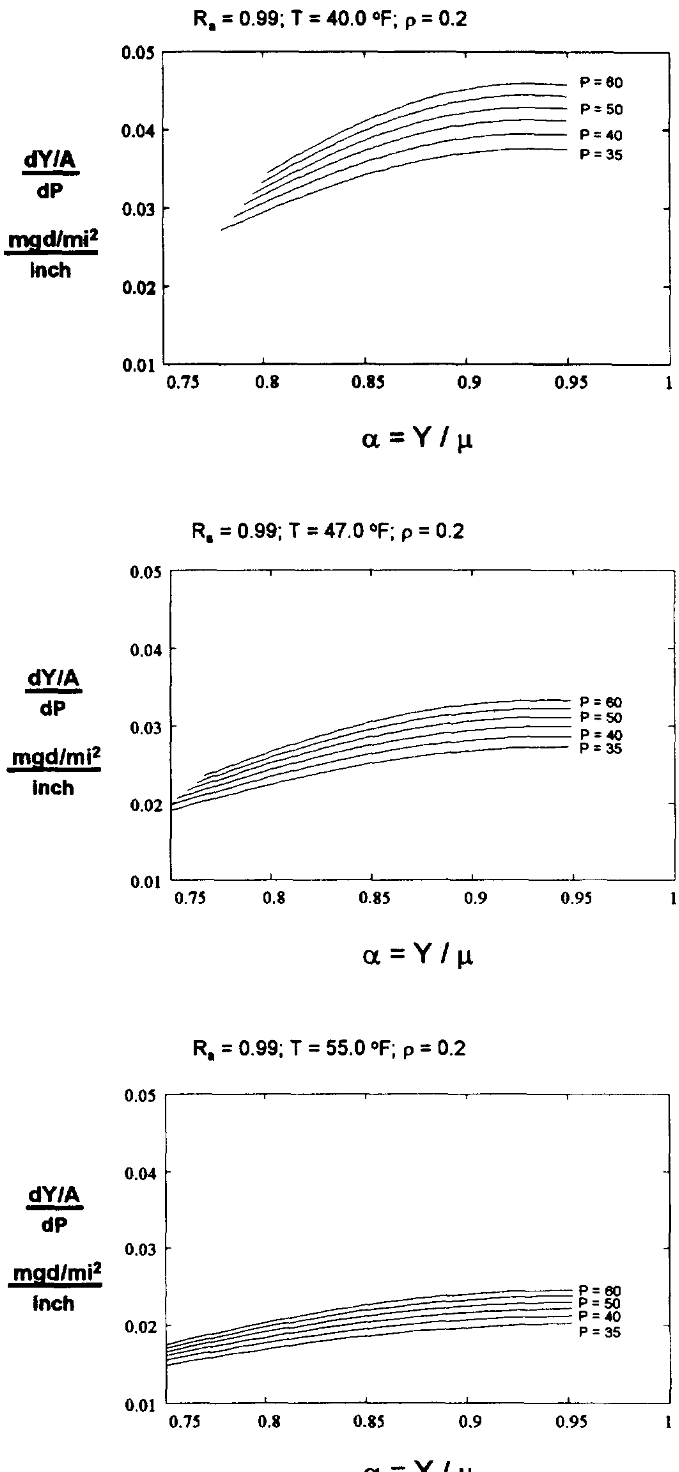

Fig. 7. The generalized sensitivity of overyear water supply yield, Y, to changes in precipitation P, in the northeastern US.

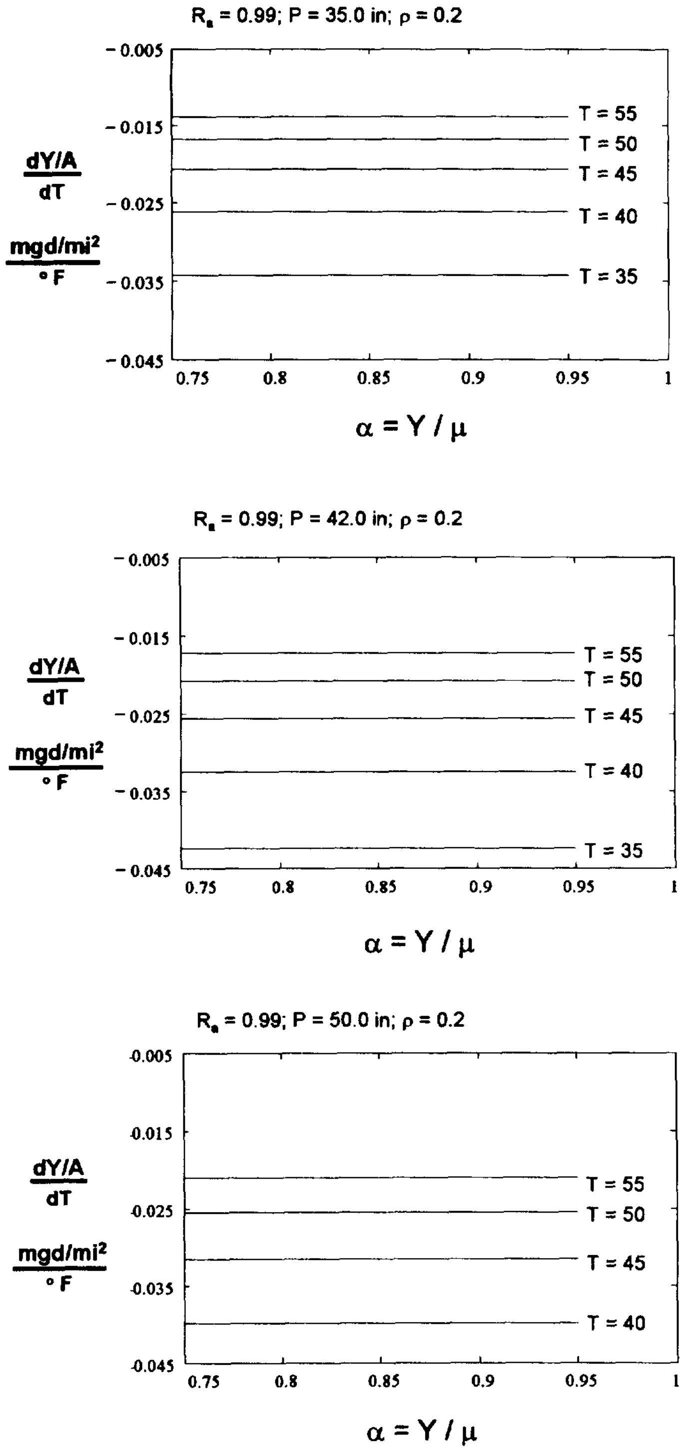

Fig. 8. The generalized sensitivity of overyear water supply yield Y, to changes in temperature T, in the northeastern US.

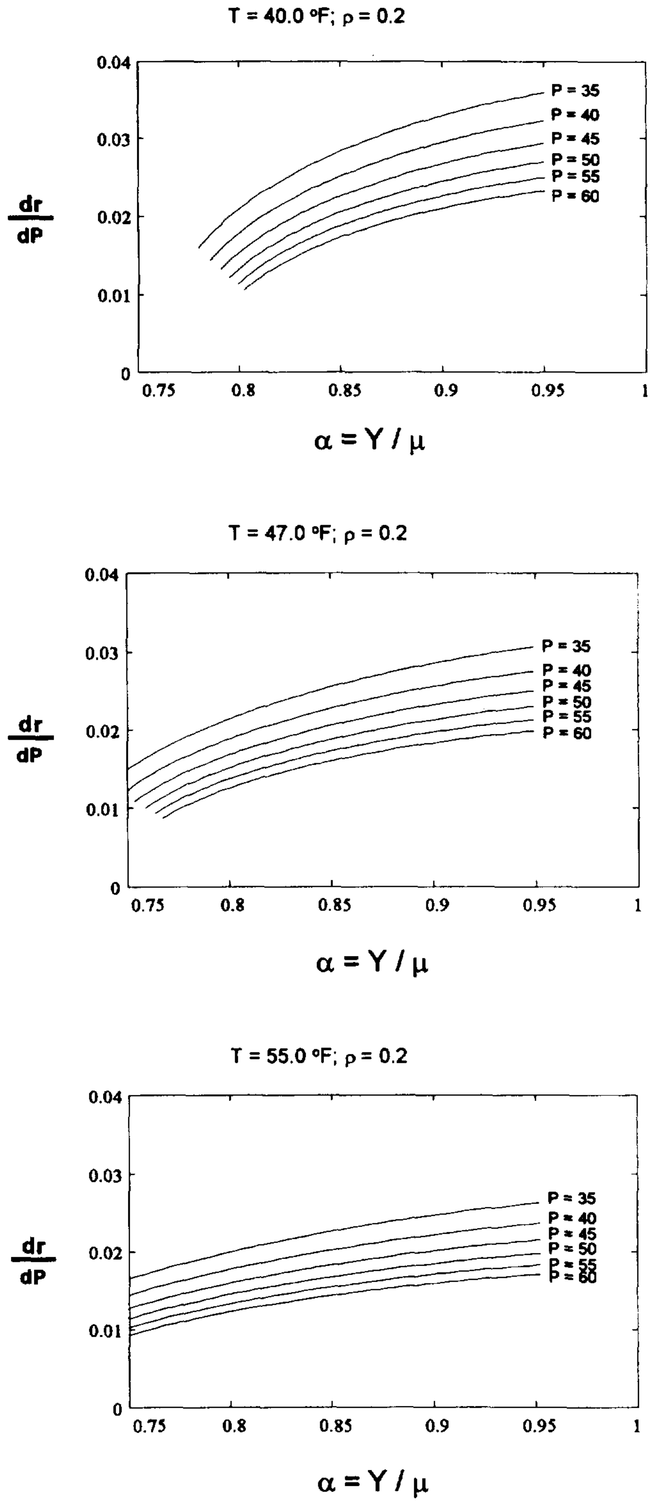

Fig. 9. The generalized sensitivity of water supply system resilience, r, to changes in precipitation P, in the northeastern US.

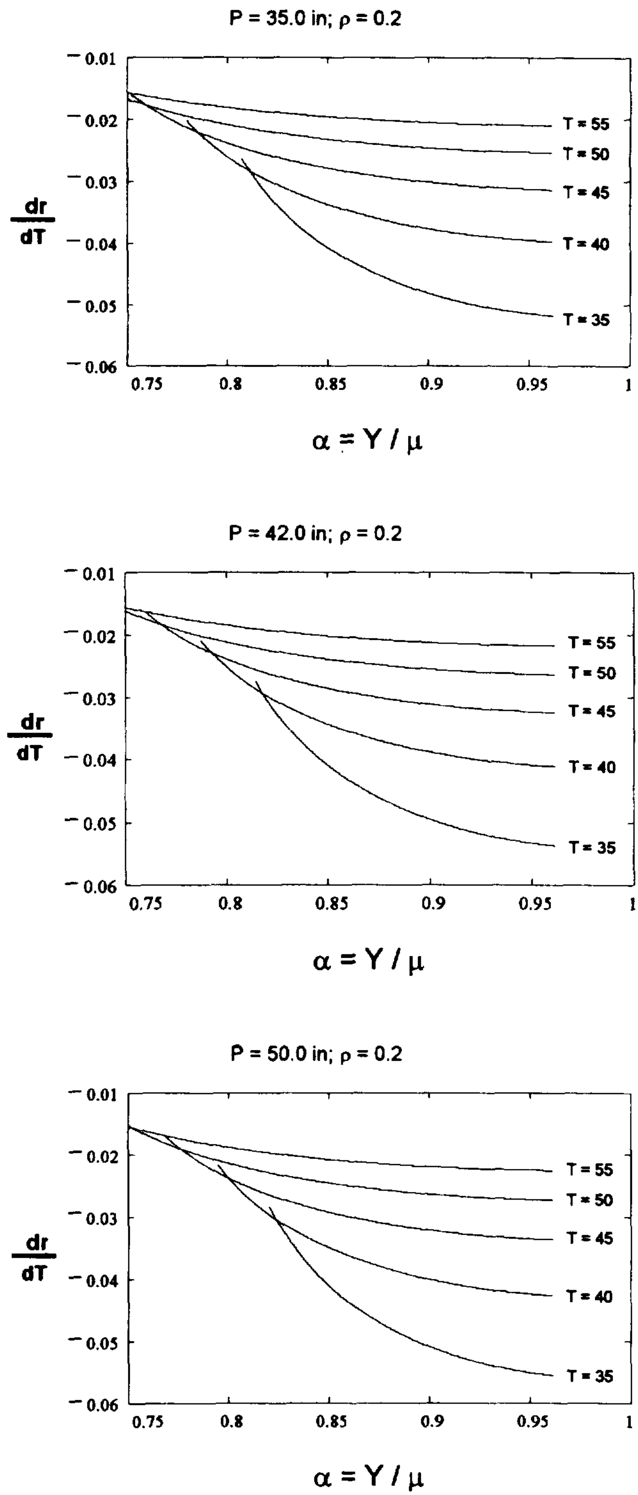

Fig. 10. The generalized sensitivity of water supply system resilience, r, to changes in temperature, T, in the northeastern US.

section were performed to obtain the sensitivity of system resilience to changes in precipitation (dr/dP) and to changes in temperature (dr/dT). Figs 9 and 10 illustrate the sensitivity of water supply system resilience to changes in precipitation and temperature, respectively.

In all cases, dr/dP is positive, hence system resilience always tends to increase as precipitation increases. Yet for a given level of development, dr/dP is inversely related to P. This is because systems with larger values of P tend to be more resilient systems than systems with small values of P and it is more difficult to increase the resilience of an already resilient system. Similarly, systems with high levels of development tend to have larger values of dr/dP than systems with lower levels of development because systems with low levels of development are already quite resilient and it is difficult to improve the resilience of an already resilient system.

In all cases, dr/dT is negative, because increases in temperature lead to increases in ET resulting in decreases in inflow, hence lowering system resilience. Again, systems which are already resilient, such as systems with low levels of development, tend to be less sensitive (dr/dT closer to zero) to changes in temperature than systems with high levels of development. For a given level of development, temperature tends to have a greater (negative) impact on resilience for systems located in cooler climates than those located in warmer climates.

Figs 7–10 document the complex effects of both temperature and precipitation on water supply system behavior emphasizing that systems with high levels of development are generally more sensitive to changes in climate than systems with low levels of development in terms of both yield and resilience. Figs 7–10 only apply to over-year water supply systems located within the northeastern U.S.

7. Conclusions

This study has sought to improve our understanding of the general relationships among climate, streamflow and water supply for the northeastern United States. Although the results of this study only apply to this region, the methodology introduced can be extended to other regions. Our approach involved the development of empirical regional relationships between annual streamflow, annual climate and geomorphic basin variables using streamflow and climatic data for the northeastern region of the U.S. shown in Figs 1 and 2. The resulting regional hydroclimatologic model of annual streamflow was shown to be remarkably precise over this broad geographic region. Ongoing research (R.M. Vogel and I. Wilson, 1997, in preparation) confirms that similar regional hydroclimatologic relations could be developed for most other regions of the U.S. Four case-studies for basins in New York revealed that a regional annual hydroclimatologic model (Eq. (6)) can reproduce the results of the more detailed daily watershed modelling approach implemented by Tung and Haith (1995).

Another goal was to combine a regional annual hydroclimatologic streamflow model with analytic relationships among storage, reliability, resilience and yield (SRRY) for the purpose of deriving general relationships among climate, streamflow and water supply system behavior for the northeastern U.S. For this purpose, we exploit the SRRY relationships recently introduced by Vogel and Bolognese (1995). Our approach was compared with the detailed monthly hydroclimatologic modelling approach summarized by Kirshen and Fennessey (1995) for the water supply system which services the Boston metropolitan area. That comparison documents that our regional annual modelling approach compares favorably with a much more detailed site-specific monthly hydroclimatologic modelling approach for the purpose of evaluating the impact of climatic change on the behavior of an over-year water supply system.

Since our modelling approach is analytic and rather general, we were able to derive generalized sensitivity curves which describe the impact of changes in climate on water supply system yield and resilience. Figs 7–10 illustrate the impact of changes in temperature and precipitation on system yield and resilience for water supply systems in northeastern U.S. which are characterized by carry-over storage or over-year behavior. These curves may be used to approximate the impact of climatic change on the yield and resilience of other existing systems within this region. It is hoped that the methodology and figures introduced here will allow municipalities and other regional authorities the opportunity to approximate the influence of potential climatic change on their water supply operations. Hopefully future studies will consider extending this methodology to other regions with possible extensions to within-year reservoir operations.

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