Hydroclimatic variability across the Susquehanna River Basin, USA, since the 17th century

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ABSTRACT: The Susquehanna River Basin (SRB) is an extensive river drainage network that covers a total of 71 228 km² in portions of New York, Pennsylvania, and Maryland, with more than 4 million inhabitants. It is the major source of fresh water to the USA’s largest estuary, the Chesapeake Bay. Therefore, the hydroclimatic conditions within the SRB affect a large human population in addition to the ecological health of one of the world’s most complicated and important estuarine environments.

This research project seeks to document the hydroclimatic history of the SRB and to understand the relationships between global-scale forcing mechanisms and the climatic variability of the Basin. Hydroclimatic variability since 1680 is documented through the reconstruction of temperature, precipitation, streamflow and a drought index using proxy and observational data. Results indicate that the hydroclimatic history of the SRB is very complex. Prior to the advent of European settlement, the SRB seems to have responded in concert with other landmasses within the North Atlantic Sector evidencing a regional manifestation of both the Medieval Warm Period and the Little Ice Age. Since the beginning of European settlement, the hydroclimate of the SRB has varied greatly on both inter-annual and decadal time scales. The climatic variability of the SRB is found to be only weakly associated with large-scale forcing mechanisms that are commonly assumed to influence the eastern USA. A notable exception is a strong temperature response associated with volcanic activity. Thus, regional forcings and land cover variations may play an important role in the hydroclimatic variability of this important river system. Copyright © 2008 Royal Meteorological Society

KEY WORDS climate variability; environmental change; hydroclimatology

Received 27 June 2007; Revised 16 November 2007; Accepted 17 November 2007

1. Introduction

The Susquehanna River Basin (SRB), situated in the mid-Atlantic region of the USA, is an extensive river drainage network covering 71 228 km² in portions of New York, Pennsylvania, and Maryland (Stranahan, 1995; SRBC, 2005a). Hundreds of tributaries feed the 714-km-long Susquehanna, notably the largest river in the country with an outlet to the Atlantic Ocean, and it is also the 16th largest river in the USA, overall. Major tributaries to the main stem of the river include the Susquehanna’s west branch and the Juniata River that flow for 367 and 138 km, respectively (SRBC, 2005a).

The SRB is comprised of roughly 60% forested land. The remaining area contains agricultural and urban environments that accommodate most of the 4.1 million people living in the SRB (SRBC, 2005b). The watershed includes six major sub-basins: Chemung (6744 km², 6% total pop.), Juniata (8822 km², 7% total pop.), lower Susquehanna (15 045 km², 47% total pop.), middle Susquehanna (9725 km², 16% total pop.), upper Susquehanna (12 805 km², 11% total pop.), and west branch Susquehanna (18 109 km², 13% total pop.; Figure 1(a); SRBC, 2005b).

A distinguishing feature of the SRB is its unique relationship to the largest estuary in the country, the Chesapeake Bay. The SRB encompasses 43% of the Chesapeake Bay drainage area, and contributes approximately 50% of the fresh water entering Chesapeake Bay. In the upper portion of the estuary, fresh water input from the Susquehanna is 90% of the total (Stranahan, 1995). These figures indicate that environmental changes in the SRB may have a profound effect on the health and productivity of the nation’s largest estuarine environment.

While many regions across the USA have undergone great environmental change since European settlement, few have withstood the repeated human-induced modifications, which the SRB has endured. This modification includes the present return to heavily forested land cover conditions after continuous deforestation in the 18th, 19th, and early 20th centuries associated with agriculture and lumbering activities. Consequently, the region represents a natural laboratory in which the effects of land-surface changes on climate and hydrologic systems can be better understood. In order to make adequate
future assessments of the hydroclimatic variability that this important watershed may experience, it is imperative that past climatic and hydrologic changes be understood. Added significance is given when considering the unique relationship of the SRB to the Chesapeake Bay.

The goal of this research is to document the hydroclimatological history of the SRB and to understand the relationship between large-scale forcing mechanisms and the hydroclimatic variability within the Basin. Specifically, climate reconstructions of basin-wide air temperature, precipitation, Palmer Drought Severity Index (PDSI) and streamflow will be developed for the past 300 years, and the variability depicted in each will be discussed. The association between several hemispheric to global-scale forcing mechanisms (i.e. El Nino/Southern Oscillation, North Atlantic Oscillation, Weighted Dust Veil Index, etc.) and hydroclimatic variables (i.e. stream flow, PDSI, etc.) within the SRB will also be investigated. It is important to note that the main emphasis of the study is to investigate relationships between global forcings and the hydroclimate of the SRB on inter-annual through decadal time scales.

1.1. Land cover change in the SRB
Since the arrival of European settlers, beginning around the year 1700, the SRB has undergone extensive land cover changes. During the 18th century, the Europeans slowly cleared much of the land in the SRB, especially the lower Susquehanna sub-basin, for agricultural purposes (Stranahan, 1995). During the 19th century, deforestation accelerated as logging investors purchased extensive tracts of land in the West Branch and upper Susquehanna sub-basins, clear-cutting immense portions of this mountainous landscape. From approximately 1860 to 1920, forest stands were clear-cut for lumber resulting in the removal of nearly all hemlock (Tsuga Canadensis L.) and white pine (Pinus strobes L.) from the region (Taber, 1995). Over 400 million logs were cut and sent to saw mills through the lumbering center of Williamsport, Pennsylvania, via the Susquehanna lumber boom (a long containment structure) along the West Branch of the Susquehanna. It is estimated that over 8 billion board feet (one board foot equals the amount of wood contained in an unfinished board with dimensions of 1” thickness, 12” length, and 12” width) passed through the West Branch boom during that time, while additional lumber was transported out of the region by other means such as logging railroads (Taber, 1995). Photographic evidence from those times indicates that logging practices were performed with little to no regard for potential environmental consequences (Taber, 1995). In later years, parts of the SRB supported coal mining operations that coincided with society’s movement towards industrialization (Napton et al., 2003). In the last seven decades, much of the area clear-cut for lumber or for agriculture has experienced natural regeneration, though the species of second- and third-growth forests are different, with far more hardwoods (Napton et al., 2003).

Though not exact, a spatial and temporal record of land use change exists for the region. More difficult to determine are the impacts that land cover variability, attributed to human activity, have had on local and large-scale climate and hydrologic systems. This is in part due to the limited availability and reliability of climate data.
before 1895, and limited research on the environmental history of the SRB.

1.2. Previous research

Although the environmental history of the SRB is not well documented, many authors have attempted to record paleoclimatic changes that have taken place in the Chesapeake Bay estuary. Since the flow of fresh water from the Susquehanna greatly affects the characteristics of the Bay, it seems reasonable to assume that information gathered for the Chesapeake Bay may be applicable to at least the lower portions of the SRB.

Brush (2001) used paleoecological indicators to assess the impact of land cover changes and climatic variability on the Chesapeake Bay over the last millennium. Using sediment cores from the main stem, tributaries, and marshes of Chesapeake Bay, Brush (2001) was able to show evidence of at least two major climatic anomalies and the effects of large land cover changes on the Chesapeake Bay. Evidence of an extended dry period lasting from, approximately, the years 1000 to 1200 is found in sediments showing higher concentrations of fossil pollen and seeds associated with a dry landscape, and large amounts of charcoal indicating frequent fires. An increase in fossil pollen and seeds representing wet habitats and a lack of charcoal indicate a wetter climate for the period from about 1200 to 1600. Since significant European colonization began (approximately in 1700), Brush (2001) found that agriculture and deforestation have been the primary agents of change in the Chesapeake Bay ecosystem. These general results were subsequently confirmed by Saenger et al. (2006) who found evidence of similar dry and wet periods using benthic foraminiferal oxygen isotopic records in Chesapeake Bay sediment cores and by Willard et al. (2003) using pollen and dinoflagellate cysts.

Cronin et al. (2003) using Mg/Ca paleothermometry found large positive anomalies in estimated Chesapeake Bay spring temperatures for the period 400 through 1000, suggesting warmer SRB temperatures during this time. The lowest temperatures in the 2000-year-record were associated with the Atlantic Sector Little Ice Age, from about 1400 to 1800. These results are similar to those of Willard et al. (2005) who used pollen records from sediment cores to estimate Bay temperatures and Cronin et al. (2005) who utilized benthic foraminifera Elphidium. Since then, late 19th- and 20th-century Bay temperatures have generally exceeded those seen during the last millennium. The data also indicate that rapid transitions in Bay temperature regimes often occur on time scales of less than 100 years, indicating a possible association with systematic changes in regional teleconnection patterns (i.e. the North Atlantic Oscillation (NAO); Cronin et al., 2003).

Using sediment cores, Pederson et al. (2005) studied the environmental history of the lower Hudson River Valley of New York (a location just to the northeast of the SRB). Their studies confirmed the work of Brush (2001); Willard et al. (2003, 2005); Saenger et al. (2006) and Cronin et al. (2003, 2005) indicating a dry, warm period from the year 800 through 1300, and a return to cooler more moist conditions from roughly 1400 through 1700. After 1700, European settlement was the major environmental forcing in the lower Hudson Valley.

Growing season moisture variability in the southern portions of the mid-Atlantic region was investigated by Quiring (2004). Using 800-year tree-ring series, the Palmer Hydrological Drought Index (PHDI) was reconstructed for the period from about 1200 through the late 20th century. Results indicate a wet period during the 18th century and anomalously dry conditions, with long summer droughts during the 16th century. In general, the reconstruction showed large inter-annual variability in summer moisture throughout the period of record.

Stahle et al. (1998) used bald cypress tree-ring chronologies to investigate summer moisture conditions during the early American colonial period in southern Virginia. They found that the Lost Colony of Roanoke Island disappeared during the most extreme summer drought in the period (about 1200 to the present), and that the large mortality rates of the Jamestown Colony during its early years (1606–1612) occurred during the most severe 7-year drought of the same period. These results suggest a possible role for climatic anomalies in these important historic events.

An integration of the studies above suggest that the SRB likely experienced an interval of warmth (Cronin et al., 2003, 2005; Pederson et al., 2005) and dryness (Brush, 2001; Willard et al., 2003; Pederson et al., 2005; Saenger et al., 2006) during the period from about 800 to 1200, a time synchronous with the Medieval Warm Period in northern Europe. The region experienced generally cooler conditions during the time of the Atlantic Sector Little Ice Age (1400–1700), with highly variable summer moisture conditions throughout the last four centuries (Stahle et al., 1998; Quiring, 2004).

2. Data and methods

2.1. Paleoclimatic data

Instrumental records of temperature and precipitation for the SRB are extant only since the mid-19th century. Estimates of hydroclimatic variables before this time are obtained using proxy indicators. In the SRB, the most common sources of climatic data proxies are from lake and estuary sediment cores and tree-ring widths. As previously discussed, sediment cores from the Chesapeake Bay have been used in a number of studies to estimate the gross magnitude of bay temperatures and basin moisture conditions to 2000 years before the present, while marsh cores were utilized by Pederson et al. (2005) in the Hudson River Valley. For the purposes of this research, it is assumed that these estimates are a good representation of the general hydroclimatic conditions of the SRB.

Tree-ring data are also available for the SRB region and are obtained for this study. Eight separate tree-ring
width chronologies are acquired from the International Tree-Ring Data Base (NOAA/NCDC WDC, 2006), representing standardized ring widths, mainly from eastern hemlock. Ring-width chronologies are gathered for trees within, or on the periphery of the SRB. Initial examination showed the ring-width chronologies to be poorly correlated with both SRB temperature and precipitation values and are, therefore, not included in any further analyses.

2.2. Instrumental records

Though instrumentally derived climatic records for the SRB start in the middle of the 19th century, climatic data were used from two nearby sites where their records started decades earlier. Mean annual temperature anomaly data exist for New Haven, CT, from 1781 (Landsberg, 1949) and mean annual temperature data for Philadelphia, PA, from 1825. A record of annual average precipitation is also available for Philadelphia since 1829 (Landsberg et al., 1968). These two stations are in close enough proximity to the SRB to be useful in reconstruction of climate across the region and, therefore, the records are selected for this analysis. United States Historical Climate Network (USHCN 2006; Williams et al., 2005) monthly temperature data are available across the SRB as early as 1854 for a single station (Cooperstown, NY) and generally during the 1870s for most locations. These data have been quality controlled and corrected for various biases (time of observation, change of instrumentation, etc.) and are considered suitable for climatological studies (NOAA/NCDC, 2006). The spatial distribution of USHCN stations across the SRB during the late 19th century is not sufficient to allow extensive use of these data in the present research. However, the USHCN data are utilized to validate the United States Climate Division data discussed below.

Beginning in 1895, United States Climate Division data are available for the entire SRB (NOAA ESRL, 2006). These data consist of monthly mean temperature and total monthly precipitation for 344 US climate divisions. Nine climate divisions have at least 40% of their land area within the confines of the SRB, and are included in the areally averaged temperature and precipitation estimates for the SRB (Figure 1(b)). These include climate divisions 1, 3, 4, 5, 6, 7, and 8 in Pennsylvania, and climate divisions 1 and 2 in New York (Figure 1(b)).

Streamflow data are available from the USGS for much of the SRB beginning in the late 19th century (USGS, 2006). The gauging station with the longest period of record within each sub-basin (Chemung, Upper, West Branch, Middle, Lower and Juniata) is obtained for analysis. Table I gives information on the streamflow data used in this study.

2.3. Large-scale forcings

A further goal of this study is to understand the role that large-scale climatic forcing mechanisms play in the variability of SRB climate. To investigate this question, monthly values for several atmospheric teleconnection indices are obtained from the United States Climate Prediction Center CPC (NOAA/NWS CPC, 2006). Monthly indices include the NAO, the Pacific/North American (PNA) teleconnection pattern, the Southern Oscillation (SOI), the Arctic Oscillation (AO) and sea surface temperature (SST) anomalies across the Nino 3.4 region. In addition, monthly Northern Hemisphere temperature anomalies are obtained from the Climate Research Unit (CRU) at the University of East Anglia (CRU UEA, 2006) and the Weighted Dust Veil Index (WDVI) is taken from Mann et al. (2000). All teleconnection data are acquired for the period 1950 through 2003, while the Northern Hemisphere Temperature anomalies and the WDVI are collected for their entire recorded periods.

2.4. Methodologies

In both the reconstruction of past SRB climate from proxy data and in the analysis of climatic forcing mechanisms, standard regression techniques are utilized. Mean annual temperature anomalies from New Haven and mean annual temperature from Philadelphia are used to reconstruct SRB temperatures from 1894 back to 1781. For the reconstruction from 1894 back to 1825, temperature data from both sites are available for the reconstruction. An overlapping period of record from 1895 through 1970 is available for the predictand (SRB areally weighted divisional temperature) and the predictors (New Haven and Philadelphia temperatures). This period of overlap is divided into a calibration period, 1940 through 1970, and a validation period, 1895 through 1939. A multiple linear regression model is developed for the calibration period. This model is applied to the validation period with New Haven and Philadelphia temperatures explaining 89% of the variance in the SRB divisional temperatures during the validation period (Figure 2(a)). This model is subsequently applied to the New Haven and Philadelphia data for the period 1825 through 1894 to complete the reconstruction.

For the period from 1824 back to 1781, only temperature anomalies from New Haven are available for the reconstruction. The same periods for calibration and testing are available, and a simple univariate linear regression model is developed for the New Haven data only. In this case, New Haven temperature anomalies explain 81% of the variation in SRB divisional temperatures during

<table>
<thead>
<tr>
<th>Sub-Basin</th>
<th>Station</th>
<th>Period of record</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chemung</td>
<td>Chemung, NY</td>
<td>1916–2002</td>
</tr>
<tr>
<td>Upper</td>
<td>Conklin, NY</td>
<td>1913–2002</td>
</tr>
<tr>
<td>West Branch</td>
<td>Williamsport, PA</td>
<td>1896–2003</td>
</tr>
<tr>
<td>Middle</td>
<td>Danville, PA</td>
<td>1906–2003</td>
</tr>
<tr>
<td>Juniata</td>
<td>Newport, PA</td>
<td>1900–2003</td>
</tr>
<tr>
<td>Lower</td>
<td>Harrisburg, PA</td>
<td>1896–2003</td>
</tr>
</tbody>
</table>

Table I. Location and period of record for USGS gauging stations used in the study.
the validation period, however, the mean value of reconstructed temperatures is 0.59 °C below the observed mean (Figure 2(b)). This model is used with the New Haven temperature anomalies to reconstruct SRB temperatures from 1824 back to 1781. A positive adjustment of 0.59 °C is made to the reconstructed values to account for the model bias.

Precipitation data from Philadelphia are available back to 1829. A methodology similar to that for temperature is used to reconstruct SRB precipitation from 1894 back to 1829 (the same calibration and validation periods are utilized). The univariate linear regression model indicated that Philadelphia annual precipitation explains 53% of the variation in SRB areally weighted total precipitation over the validation period (Figure 3). The scatter plot shows a definite bias in the model with mean predicted values more than 9.0 cm greater than observed values during the validation period. An adjustment of −9.6 cm is applied to the reconstructed values to account for the bias. The model is then used with the pre-1895 Philadelphia data to reconstruct SRB precipitation back to 1829.

Cook et al. (1999) developed a gridded summer drought dataset for the USA estimated from a dense nationwide network of annual tree-ring chronologies covering the period from the late 1600s through 1978. The drought metric used is the Palmer Drought Severity Index (PDSI; Palmer, 1965). The PDSI is a meteorological drought index that is calculated using precipitation and temperature data from a given location, as well as information on the amount of water present in the soils. Grid point 143 in the Cook et al. (1999) data lies near the center of the SRB. The data from this grid point is obtained for the period 1680 through 1980. Observed summer PDSI values for the SRB, derived from climate division data, are available for the period 1895 through 2005. During the period of overlap between datasets (1895 through 1980) a calibration (1938 through 1980) and validation period (1895 through 1937) are used to develop a univariate regression model for subsequent reconstruction of SRB summer PDSI values. Figure 4 shows a scatter plot of the observed SRB PDSI and the reconstructed values for the validation period ($R^2 = 0.41$).

Susquehanna River discharge values at Harrisburg, PA, (furthest downstream gauging station) were reconstructed using precipitation data from Philadelphia and summer PDSI values from grid point 143 of Cook et al. (1999). Both data types are available for the reconstruction of

Figure 2. Scatter plot of reconstructed temperature versus observed SRB temperature for the validation period (1895–1939) using (a) New Haven and Philadelphia temperatures ($R^2 = 0.89$), and (b) using only New Haven temperature data ($R^2 = 0.81$).

Figure 3. Scatter plot of reconstructed precipitation versus observed SRB divisional precipitation for the validation period (1895–1939). Reconstructed values calculated using Philadelphia precipitation ($R^2 = 0.53$).

Figure 4. Scatter plot showing relationship between observed and reconstructed summer PDSI during validation period ($R^2 = 0.41$).
streamflow from 1895 back to 1829. For the reconstruction from 1828 back to 1680, only PDSI data are available. During the period of overlap between data sets (1896 through 1980) a calibration (1945 through 1980) and validation period (1896 through 1944) are used for subsequent reconstruction of discharge values using the same methodologies as discussed above. Figure 5 shows scatter plots of the predicted versus observed discharge values for the validation period (1896 through 1944) using Philadelphia precipitation and summer PDSI in a multiple regression model developed over the calibration period (Figure 5(a)) and in a univariate regression model using only PDSI values (Figure 5(b)). Philadelphia precipitation and PDSI values explain 36% of the variation in discharge for the validation period, while PDSI values alone explain 22% of the variation in discharge over the same period. The multiple regression model (precipitation and PDSI) is used to reconstruct discharge values for the period 1895 back to 1829 and the univariate model (PDSI) is used for the reconstruction from 1828 back to 1680.

To ascertain the degree to which SRB climate responds to large-scale forcing mechanisms, stepwise, multiple linear regression is utilized, regressing the predictor variables (large-scale forcing indices) against the predictand (SRB hydroclimatic variable). This analysis is conducted using annual or seasonal data. Only large-scale forcings that have been shown to affect climate in the eastern USA are used in the analysis.

3. Results

3.1. Temperature

Mean annual temperature areally weighted across the SRB from climate division data average 8.8°C for the period from 1896 through 2003. No long-term trends are apparent in the observed data for this period with instrumental records (Figure 6). In general, temperatures in the early portion of the record are below the long-term mean, but increased steadily through the early portion of the 20th century to the early 1950s. After this, a cooling period took place that persisted through the 1980s. Since that time, temperatures have been rising, reaching levels last seen in the 1930s and 1950s. Mean annual temperatures during the entire period range from a high of 10.6°C in 1998 to a low of 7.2°C in 1904. Thus, the range of annual temperature variability, detected over this period with extensive observational data in the SRB, is 3.4°C.

Mean annual temperatures for the SRB are reconstructed according to the method outlined above for the period 1825 through 1895 (Figure 6). Data from Philadelphia and New Haven are used to reconstruct this record. Mean annual temperatures average 8.2°C during this period, more than 0.5°C cooler than the later period (Figure 6). This average includes the two coldest years in either the observational or reconstructed periods; 1836 and 1837 (with mean annual temperatures of 6.2°C and 6.8°C respectively). After a relatively warm period in the late 1820s, temperatures dipped sharply in the 1830s and then slowly warmed during the decades leading up to 1895.

Only data from New Haven is available for the reconstruction of SRB mean annual temperature prior to 1825. A significant warm interval dominated during the first decade of the 1800s followed by an intense cold period (second only to the 1830s) during the 1810s. The period from 1781 to 1800 is generally near the long-term mean (Figure 6).

Taken as a whole, the observed and reconstructed mean annual temperatures for the SRB from 1781 through 2003 indicate a period with conditions near the long-term mean early in the record, followed by two intervals of extreme cold, one in the 1810s and a second, more intense cooling in the 1830s (Figure 6). A nearly century-long warming trend began about 1840 and continued through the 1940s, before conditions once again cooled from approximately 1950 through the early 1990s. Since that time temperatures have risen to levels equivalent to the warmest on record. The average mean annual temperature for the entire period of record is 8.5°C, ranging from a low of 6.2°C in 1836 to a high of 10.6°C in 1998, a 4.4°C difference. A statistically significant ($p = 0.01$) positive trend in temperature of 0.04°C per decade is found for the entire period of record (1781 through 2003).
3.2. Precipitation

Areally weighted total annual precipitation is calculated using the climate division data for the period 1896 through 2003. A general increasing trend in precipitation is apparent beginning in the mid-1920s, continuing through the end of the record (Figure 7). Superimposed on this trend are two distinct decadal-scale variations with lower precipitation values during the 1960s and generally higher values in the 1970s. The 1960s are generally considered the ‘drought of record’ for many locations within the SRB. It is interesting to note, however, that the dry episode of the 1960s is actually a return to pre-1920 precipitation levels, and represents a dry interval only when compared to the post-1920 time period. Average annual precipitation values for the entire period with extensive observations is 99.9 cm, ranging from 64.0 cm in 1930 to a high of 143.5 cm in 1996. It should be noted that the increasing trend in precipitation since the mid-1920s was found in both the USHCN and the climate division data.

Using the method outlined above, annual precipitation for the SRB is reconstructed using data from Philadelphia for the period 1829 through 1895. The reconstructed average annual precipitation over this interval is 95.2 cm, matching very closely the observed values from the early portion of the instrumental record. Throughout most of this time, annual precipitation values hovered near the long-term mean. A general increase in precipitation totals is evident in the 1860s and a general decrease in the 1880s, followed by near normal values until the beginning of the upward trend in the 1920s. A statistically significant ($p = 0.01$) positive trend of 0.73 cm per decade is found for the interval 1829 through 2003.

3.3. Streamflow and the PDSI

Streamflow data are available from the USGS for much of the SRB beginning in the late 19th century...
The gauging station with the longest period of record within each sub-basin (Chemung, upper, West Branch, middle, lower and Juniata) is obtained for analysis (Table I). Periods of record range from 108 years (1896–2003) in both Harrisburg (lower Basin) and Williamsport (West Branch Basin) to 88 years (1916–2003) in Chemung (Chemung Basin). An inspection of the time series of mean annual streamflow for each station (Figure 8) indicates a consistent signal across all basins of low streamflow values during the 1970s and again in the 1990s. It is important to note that although the 1960s appear as low-flow years across the entire SRB, several basins have had drier periods, even when considering the smoothed data. Thus, the 1960s should be considered the drought of record only since the 1930s, which had streamflow values equally as low as those of the 1960s in many of the sub-basins.

Mean annual discharge for the Susquehanna River at Harrisburg is reconstructed using precipitation data from Philadelphia and summer PDSI values from Cook et al. (1999) for the period 1829 through 1895 and PDSI values alone for the period 1828 back to 1680 (Figure 9). The average annual discharge over the reconstruction period is 972 cms, matching very closely the observed value of 963 cms from the observational record. The reconstruction period is characterized by values fluctuating around the long-term mean from 1680 through approximately 1780. After 1780, larger decadal-scale variability is evident in the data. Especially evident are periods with large discharge values in the 1830s and 1870s and an interval with relatively low discharge values in the 1880s. No statistically significant trend in discharge values is found for the period from 1680 through 2003.

Figure 10 shows the time series of reconstructed summer PDSI values produced using the regression methodology discussed above. No long-term trend in summer PDSI values is found across the period of record ($p = 0.4$). However, three decadal-scale episodes of persistently dry conditions with low inter-annual variability are identified. The first occurs during the 1730s and 1740s, prior to extensive European settlement across most of the SRB. Another longer-lasting dry interval stretches from the late 1840s through the 1880s, coinciding with the time of most intense deforestation across the SRB. The third episode is present during the 1960s, a time of reforestation across much of the SRB. Three periods with persistently moist summer conditions are evident in the 1810s, the 1830s, and again in the 1970s. However, the length of these moist intervals appears to be shorter than those of dry periods. In general, large inter-annual variability of summer PDSI is apparent throughout the remainder of the record.

Streamflow data and the PDSI point to a SRB hydrologic system that varies greatly on an inter-annual basis. The rhythm of inter-annual variability is punctuated...
Figure 9. Reconstructed Harrisburg discharge. Dashed gray line is observed discharge, solid dark gray line is reconstructed discharge using Philadelphia precipitation and summer PDSI, and dashed black line is using only summer PDSI values. Solid black line represents smoothed values.

Figure 10. Reconstructed summer PDSI values for the Susquehanna River Basin. Gray line is observed summer PDSI, and black line is reconstructed values.

by decadal-scale intervals of persistent dry and moist periods, with dry episodes tending to persist longer.

3.4. Large-scale forcing of SRB hydroclimatology
The final goal of this study is to better understand the role of large-scale climate forcing mechanisms in the hydroclimatic variability of the SRB. To this end, values of several atmospheric and oceanic teleconnections shown to be related to the climate of eastern North America are obtained for analysis. Annual and seasonal indices used for the analysis include the NAO, the PNA, the AO, the SOI, SST anomalies in the Nino 3.4 region, and Northern Hemisphere temperature (NHT) anomalies.

A stepwise, multiple linear regression analysis is used to ascertain the association between the independent variables (teleconnections; NAO, PNA, AO, SOI, SST, NHT) and the dependent variables (seasonal and annual values of SRB temperature, precipitation, and PDSI). The analysis is carried out for the 54-year period from 1950 through 2003. An F-probability of 0.05 is used for variable entry while an F-probability of 0.10 is used for removal of a variable. Results of the analysis are given in Table II.

Simultaneous large-scale forcing mechanisms explain only a fraction of the annual and seasonal variability in SRB temperature, precipitation, and PDSI. The largest association is found between SRB winter temperatures and the Arctic Oscillation and Pacific/North American pattern during the winter season, indicating an important role for these teleconnections in influencing SRB winter temperatures (Table II). Weaker relationships exist between SRB spring, summer, and annual temperatures...
Table II. Results of multiple linear regression between SRB climate variables (dependent) and hemispheric forcing mechanisms (independent). Analysis is conducted for the 54-year period 1950 through 2003.

<table>
<thead>
<tr>
<th>SRB Variable (dependent)</th>
<th>Forcing(s) entering equation (independent)</th>
<th>Adjusted $R^2$ value</th>
</tr>
</thead>
<tbody>
<tr>
<td>SRB Temp–winter</td>
<td>Arctic Oscillation, N.H.T., PNA</td>
<td>0.220**</td>
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<tr>
<td>SRB Temp–spring</td>
<td>N.H.T.</td>
<td>0.059**</td>
</tr>
<tr>
<td>SRB Temp–summer</td>
<td>N.H.T.</td>
<td>0.178**</td>
</tr>
<tr>
<td>SRB Temp–autumn</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>SRB Temp–annual</td>
<td>N.H.T.</td>
<td>0.162**</td>
</tr>
<tr>
<td>SRB Precipitation–winter</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>SRB Precipitation–spring</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>SRB Precipitation–summer</td>
<td>NAO</td>
<td>0.058**</td>
</tr>
<tr>
<td>SRB Precipitation–autumn</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>SRB Precipitation–annual</td>
<td>NAO</td>
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<td>SRB PDSI–winter</td>
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<td>–</td>
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<tr>
<td>SRB PDSI–spring</td>
<td>PNA</td>
<td>0.088**</td>
</tr>
<tr>
<td>SRB PDSI–summer</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>SRB PDSI–autumn</td>
<td>PNA</td>
<td>0.072**</td>
</tr>
<tr>
<td>SRB PDSI–annual</td>
<td>–</td>
<td>–</td>
</tr>
</tbody>
</table>

** indicates significance at the 95% level

and Northern Hemisphere temperature anomalies suggesting that SRB temperatures during these seasons tend to be in phase with hemispheric variations. SRB precipitation is poorly specified by the large-scale forcings, although the NAO shows weak associations with precipitation during the summer season and annually. Finally, the PDSI is weakly related to the PNA teleconnection during the spring and fall seasons (Table II).

During the course of this analysis, it is observed that many of the coldest periods across the SRB are coincident with times of increased volcanic activity as measured by the WDVI (Lamb, 1971; Mann et al., 2000). To further investigate this potential relationship annual values of the WDVI are regressed against the reconstructed SRB temperatures for the period 1781 through 1995. Pearson correlation coefficients suggest a strong relationship between the WDVI and SRB mean annual temperatures. A correlation value of −0.405 ($p = 0.01$) is found for simultaneous associations while a value of −0.442 ($p = 0.01$) is found for 1-year lag associations (WDVI leading SRB mean annual temperature by 1 year).

To further investigate the strength of the relationship, an Independent Samples T-test for the difference between means was conducted. The SRB temperature data was stratified by the WDVI. All years with WDVI values greater than 30 were placed in one sample while all years with values less than 30 were placed in another. The results of the T-test are shown in Table III. The T-test indicates that mean annual SRB temperatures are statistically different when stratified by values of the WDVI.

An inspection of Figure 11 confirms a strong correspondence between WDVI values and mean annual temperature across the SRB. Especially evident is the relationship between the two coldest periods in the record (1810s and 1830s) and the WDVI. Moreover, the longest prolonged period of warmth across the SRB (1920 to 1960) is coincident with a period in which WDVI values are very low (Figure 11).

In summary, these results indicate that the SRB does not respond in a strong manner to any of the large-scale atmospheric or SST forcings evaluated in this study. However, the analysis does indicate a strong relationship between SRB mean annual temperatures and the

Table III. Results of Independent Samples T-test.

<table>
<thead>
<tr>
<th></th>
<th>N</th>
<th>Mean</th>
<th>S.D.</th>
<th>T-value</th>
<th>Sig.</th>
</tr>
</thead>
<tbody>
<tr>
<td>WDVI $&gt; = 30$</td>
<td>109</td>
<td>8.23 °C</td>
<td>0.66 °C</td>
<td>−5.624</td>
<td>0.000</td>
</tr>
<tr>
<td>WDVI $&lt; 30$</td>
<td>106</td>
<td>8.73 °C</td>
<td>0.65 °C</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 11. Observed and reconstructed SRB mean annual temperature (line graph) and annual values of the Dust Veil Index (DVI) gray bars.
annual value of the WDVI. During times with significant volcanic activity, SRB mean annual temperatures drop substantially, presumably in response to this global-scale forcing.

4. Summary and conclusions

The Susquehanna River Basin is home to more than 4 million people and is the major source of fresh water to the USA’s largest estuary, the Chesapeake Bay. Therefore, the hydroclimatic conditions within the SRB affect a large population and the ecological health of one of the world’s most complicated and important estuarine environments. The SRB has undergone extensive land use and land cover changes since the arrival of Europeans in the early 18th century. These changes have affected both the climate and the hydrology of the region, but these effects are difficult to quantify given the lack of conventional hydroclimatic data prior to the late 1800s. Moreover, the sensitivity of the SRB to hemispheric-scale forcing mechanisms (atmospheric teleconnections, anomalous SSTs, and volcanic activity) is not well understood.

This research project has sought to document the hydroclimatic history of the SRB and to understand the relationships between hemispheric-scale forcing mechanisms and that history. The gross temperature of the SRB during the previous two millennia can be inferred from the Chesapeake Bay paleothermometry work of Cronin et al. (2003). Their data indicate the existence of warmer spring season Chesapeake Bay temperatures during the period from about 400 to 1000. In addition, a period of colder temperatures was identified from approximately 1400 through 1800, coinciding with the Atlantic sector Little Ice Age. This work is supported by both Cronin et al. (2005) and Willard et al. (2005). SRB temperature was reconstructed for the period from 1781 through 1894 using data available from two stations just outside the confines of the SRB; New Haven and Philadelphia. These reconstructions show conditions near the long-term mean at the beginning of the nineteenth century, followed by two intervals of intense cold in the late 1810s and the late 1830s. Temperatures slowly rose after 1840 through the end of the reconstruction period in 1895. Observations since 1895 indicate a continuation of the warming trend through the 1950s, followed by a period of cooling from the 1960s through the 1980s. Since then, temperatures have generally been increasing through the end of the period of record.

Basin moisture conditions can be inferred from the sediment core work of Brush (2001); Willard et al. (2003); Pederson et al. (2005) and Saenger et al. (2006) for the period 1000 through 1800. Sediments indicate a dry period (lower precipitation) in the SRB from about 1000 through 1200, followed immediately by wet conditions (higher precipitation) for a 400-year interval from approximately 1200 through 1600. After this time, human modification of the environment dominates the characteristics of the sediment cores in both the Chesapeake Bay and the lower Hudson River. Reconstructed precipitation for the SRB was developed in this research using data from Philadelphia for the interval 1829 through 1895. Reconstructed SRB precipitation values indicate conditions were near the long-term mean throughout the reconstruction period. Precipitation observations are available across the SRB beginning in 1896. These data show conditions continuing near the mean through the 1920s then trending upward through the end of the period, except for the dry decade of the 1960s. This long-term trend in precipitation was confirmed using the USHCN records available across the SRB. Streamflow data from gauging stations in each of the SRB’s six sub-basins show large inter-annual variability and decadal-scale variations throughout the period of record (1900–2003). The drought of the 1960s is clearly evident in each sub-basin, as is a tendency for greater year-to-year variability since 1970. Reconstructed discharge values for the SRB indicate conditions near the long-term mean during most of the 18th century, with decadal-scale fluctuations in streamflow throughout the 19th century. Especially apparent are positive discharge anomalies from 1860 through the 1880s. Reconstructed and observed values of the PDSI show no long-term trend over the period 1680 through 2003. However, as with streamflow there is large inter-annual and decadal-scale variability imbedded within the time series.

No hemispheric-scale forcing mechanisms are found to play a consistently important role in the hydroclimatic variability of the SRB over the last half century. This result seems to contradict several studies that have shown significant relationships between large-scale teleconnections and climate across the northeastern USA (Leathers et al., 1991; Miller et al., 2006; Notaro et al., 2006). However, a close inspection of these studies indicates that the relationships between the teleconnections and Northeast US climate varies greatly over space and timing within the annual cycle. Specifically, Miller et al. (2006) choose to use a synoptic climatological typing procedure instead of large-scale indices of climate variability (i.e. ENSO, NAO, PNA, etc.) in their study of spring SRB discharge because the large-scale indices have not been shown to be good predictors of streamflow within the SRB.

Many of the coldest periods in the last 200 years have been associated with large values of the WDVI (Lamb, 1971; Mann et al., 2000). An analysis of the association between SRB temperatures and the WDVI indicate strong connections between volcanic activity and temperatures across the SRB.

Taken as a whole, the results indicate that the hydroclimatic history of the SRB is quite complex. Prior to the advent of European settlement (approximately 1700), the SRB seems to have responded in concert with other land masses within the North Atlantic sector evidencing a regional manifestation of the Medieval Warm Period (Hughes and Diaz, 1994; Cronin et al., 2003; Pederson et al., 2005) and a similar manifestation of the Atlantic sector Little Ice Age. Since the beginning of European settlement, the hydroclimate of the
region has varied greatly on an inter-annual basis, but this variability is not directly associated with any large-scale forcing mechanisms that are commonly assumed to influence the climate of this region. One notable exception is the significant temperature response associated with volcanic eruptions. The lack of any strong relationship with hemispheric-scale forcings points to a role for more regional-scale mechanisms in the variability of the SRB climate. In addition, changes in land cover and land use (especially deforestation and reforestation) are also likely contributors to hydroclimatic variability (Brush, 2001; Pederson et al., 2005). Isolating and understanding regional mechanisms, and better specifying the role of land cover change in the climatic variability of the SRB is the subject of future, more detailed studies.

Acknowledgements

The authors would like to acknowledge the valuable assistance of Kevin Brinson, Jason Butke, Nicholas Klingaman, and Elsa Nickl in the preparation of this manuscript. We also thank Dr Delphis Levia for his assistance throughout the course of the research, and the anonymous reviewers for their helpful suggestions.

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