Hydroclimatology of Meteorologic Floods

K.K. Hirschboeck
Laboratory of Tree-Ring Research
University of Arizona

L.L. Ely
Department of Geology
Central Washington University

R.A. Maddox
Cooperative Institute for Mesoscale Meteorological Studies
University of Oklahoma

Using the framework of flood hydroclimatology, we present an overview of precipitation systems that generate floods. We also discuss the role of antecedent climatic and hydrologic conditions and highlight factors that amplify the hazards associated with flood occurrence. Finally, we address flood variations over long-term time scales and examine the implications of hydroclimatic flood variations for the analysis and management of flood hazards.

Scale is a key component of flood causality because the way precipitation is delivered in space and time affects the type of flood and its accompanying hazards. Precipitation systems that occur at one scale are strongly interconnected with systems at other scales, and larger scale processes set the stage for activity at smaller scales. Across all scales, the *persistence* of a precipitation system is a key element for generating exceptionally large floods. Most major floods are characterized by a synergistic combination of atmospheric, hydrologic, and drainage-basin factors that intensify the event.

Over long-term time scales, regional flooding variability identified from historical records and paleoflood information may be linked to persistent synoptic to macroscale patterns of atmospheric circulation. Certain regional responses to teleconnections [e.g., El Niño/Southern Oscillation (ENSO)] that are detectable in precipitation records often are not as strong in flood records because of the complexities of flood causality. Emerging temporal and spatial relationships among flood hydroclimatic regions globally can augment information from specific regions to evaluate potential flood hazards under changing climatic conditions.

Introduction

Meteorologic floods arise when weather, climate, and hydrology work together in ways that produce greater-than-average amounts of runoff, exceeding the capacity of stream channels. The timing and spatial distribution of precipitation with respect to a river basin are key determinants of whether a flood occurs. Hence, to understand the origins of meteorologic floods, the temporal and spatial scales of the atmospheric processes that lead to floods must be addressed in conjunction with the temporal and spatial scales of the hydrologic processes that operate in specific drainage basins. This chapter focuses primarily on the atmospheric processes that generate floods, but we also address the issue of synergy between atmospheric and hydrologic processes in flood development.

The meteorologic and climatologic causes of different types of flooding have been discussed by many authors (Ward, 1978; Maddox et al., 1979, 1980; Hayden, 1988; Hirschboeck, 1987a, 1988, 1991, 1996; Doswell et al., 1996). We will not replicate either the comprehensive scope or specific concentration of these earlier works. The goal of this chapter is to present an overview of the meteorologic and climatologic processes that directly and indirectly cause floods, underscoring factors that amplify the hazards associated with flood occurrence. In addition, we highlight flood variations over long-term time scales (decadal-to-millennial) and examine the implications of long-term flood information for perception, analysis, and management of flood hazards.

Hirschboeck, K.K., Ely, L. and Maddox, R.A., 2000, Hydroclimatology of meteorologic floods, in Wohl, Ellen, ed, *Inland Flood Hazards: Human, Riparian and Aquatic Communities*, Cambridge University Press, p. 39-72

Flood Hydroclimatology: A Context for Understanding Flood Causation

The atmospheric causes of flooding can best be understood within the framework of *flood hydroclimatology.* Flood hydroclimatology is an approach to analyzing floods from the perspective of the temporal context of their history of development and variation and the spatial context of the local, regional, and global atmospheric processes and circulation patterns from which the floods develop (Hirschboeck, 1988). This approach is based on identifying the meteorologic causes of floods, but it also seeks to address this meteorologic-scale activity within a broader spatial and temporal, climatic perspective. In this way flood variability can also be examined in terms of antecedent conditions, regional relationships, large-scale anomaly patterns, global-scale controls, and long-term trends (Hirschboeck, 1988).

Meteorologic Processes That Directly Cause Flooding

Simply stated, meteorologic floods occur when excessive precipitation over a watershed cannot be fully accommodated by the basin's internal storage reservoirs and drainage network. The way this above-average precipitation is delivered – in space and time – affects both the type of flood that occurs and its accompanying hazards. The intensity of flood-causing precipitation and the temporal and spatial distribution of the precipitation with respect to a drainage basin depend on the kinds of storm systems that deliver the precipitation. Storms and storm systems can be classified in a number of ways: for example, Houze (1981) provides a comprehensive overview of the structures of atmospheric precipitation systems. We focus on storms and weather systems that deliver exceptionally large amounts of precipitation and describe different types of storm systems from the perspective of both their causative mechanisms and their spatial and temporal scales of influence, as represented by storm size and duration.

The scales of meteorological phenomena associated with heavy rainfall relate directly to the area affected by the precipitation and therefore to the type and extent of possible resultant flooding (Figure 2.1). Generally speaking, larger scale precipitation processes (i.e., those of **macroscale** and **synoptic scale**) tend to produce moderate to heavy rainfalls over fairly large regions (e.g., precipitation from one or more extratropical cyclones that affects a major river system, such as the Ohio River or Upper Mississippi River basin). The floods associated with macroscale and synoptic scale atmospheric processes tend to develop over tens of hours to days and affect large geographic regions. These large-scale floods have the potential to produce extreme property and crop damage and to cause tremendous economic losses. In regions of the world with efficient communication networks and public streamflow prediction and warning systems, floods on the larger scales pose only a small threat to human life. This is not true for less-developed countries where loss of lives may be significant.

In contrast to larger scale atmospheric processes, **mesoscale** and **storm scale** (Figure 2.1) processes are much shorter lived and can produce extreme amounts of rainfall over very localized areas (e.g., a portion of a state, a city, a small drainage basin) within a few hours or less. Floods caused by mesoscale and storm-scale rainfall tend to produce rapid runoff and

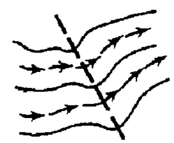
LENGTH SCALE KM	TIME SCALE (1 MONTH — 1 DAY — 1 HOUR)	SCALE TERMINOLOGY	EXAMPLE OF FEATURE	NATURE OF FLOODING
>10,000	Planetary long waves Teleconnections ITCZ & Monsoon circulations Macroscale waves & blocking patterns	MACRO-SCALE	 Planetary long waves	Floods develop over tens of hours to days Widespread floods affect major river basins and large geographic regions Antecedent soil moisture and/or snow cover from prior storms may have affected region due to large-scale steering or blocking patterns Flood effects may persist for weeks Extensive property and crop damage likely
10,000 to 2,000	Synoptic short waves e.g., ridges & troughs Extratropical cyclones & fronts	SYNOPTIC SCALE	Synoptic-scale ridges & troughs  Extratropical cyclone 	
2,000 to 50	Mesoscale short waves Tropical storms easterly waves, cloud clusters MCSs: MCCs, squall lines rainbands	MESOSCALE	Squall line   Tropical storm  Mesoscale short wave	Moderate-to-heavy rainfall, damaging winds or storm surges from severe weather may exacerbate flood conditions Rapid runoff and extreme flash flooding possible over fairly wide areas Heavy loss of human lives possible
50 to 5	Thunder-storms	STORM SCALE	 Convective thunderstorms	Localized flash flooding in small drainage basins; may take people by surprise Small-basin floods may be catastrophic with terrain effects and/or synergistic storm movement

Figure 2.1. A simple subdivision of atmospheric phenomena that influence the hydroclimatology of heavy rains and floods (modified from Orlanski, 1975; Hirschboeck, 1987a). The time scale refers to the typical duration or lifetime of specific phenomena. The length scale refers to the typical horizontal scale of specific phenomena.

extreme flash flooding. These events often cause major damage in conjunction with loss of human lives. Because of their short temporal and small spatial scales, prediction of these events is extremely difficult. Countries in which weather service agencies have been established may use radar and other observing technologies to detect the heavy rainfall associated with mesoscale and storm-scale events and to provide warnings of impending flash floods, but even this may not be enough to prevent deaths. Countries without such technologies are even more prone to disastrous flash-flooding scenarios and major loss of lives from these smaller-scale weather systems.

It should be noted that whenever precipitation systems of any scale develop over, or move into, regions of complex terrain, the character of the precipitation can be both increased and localized. The reasons for this are complicated but relate to two processes: (1) terrain causes increased upward motion whenever low-level winds flow upslope, and (2) mountains can influence the movement of precipitation systems, often causing them to slow or become nearly stationary. Orographic enhancement of precipitation is important in regions of complex terrain worldwide – from the intermontane western United States to the Himalayan foothills – and orographically enhanced precipitation events have resulted in many extreme floods (e.g., Matthai, 1969; Maddox et al., 1978; Smith et al., 1996).

In the following overview of specific flood-causing precipitation systems, we begin with smaller-scale systems and progress to larger-scale systems, arranging our discussion around atmospheric scaling terminology adapted from a framework advanced by Orlanski (1975) (see Figure 2.1). However, it is important to realize that precipitation systems that occur at one scale are strongly tied to systems and processes at other scales and that in most cases the larger-scale processes set the stage for atmospheric activity at smaller scales.

Smaller-Scale Precipitation Systems: Storm and Mesoscale

Storm-scale and mesoscale features that produce heavy rains are almost always convective in nature. Convective storms form within an atmosphere that is conditionally unstable. If an air parcel becomes saturated, the heat released when water vapor condenses causes the parcel to be warmer and less dense than its surrounding environment. The parcel then rises rapidly, much like a cork released under water. Sometimes deep convective clouds can reach heights well over 15 km above the ground. When lightning occurs within or from convective clouds, thunderstorms or thundershowers are said to be in progress. Intense convective storms can produce prodigious rainfall with amounts of tens to hundreds of millimeters occurring over small areas (e.g., tens of square kilometers) during periods of minutes to several hours. These heavy convective rains are usually, but not necessarily, accompanied by lightning.

Storm-Scale Systems. Isolated thunderstorms can deliver localized rainfall rates and amounts sufficient to produce local flash flooding. The most intense rain rates originate within isolated storms that occur in larger-scale synoptic environments with very high moisture contents in concert with very light winds aloft. These storms can produce rainfall rates of more than 100 mm/h and persist for up to an hour, although the most intense rain is likely only over areas of a few tens of square kilometers. Some isolated thunderstorms, called heavy precipitation supercells, can produce rain rates of this intensity for several hours (Weisman and Klemp, 1986). Flash flooding is likely if the intense rainfall produced by isolated thunderstorms is concentrated within a small drainage basin or if the storm moves slowly across the basin. When the rain from such storms falls into basins with a very impervious surface character, as is often the situation in sparsely vegetated, semiarid, desert, and mountainous regions of the world, the flood hazard increases dramatically. The intensity of the rain and rapid runoff can cause severe erosion and carry huge amounts of debris, clogging channels, culverts, and bridge openings and exacerbating the damage. Because these types of floods often occur in small or ephemeral watersheds in rural or remote areas, inhabitants, motorists, campers, and hikers may be taken by surprise. In canyons and mountainous regions, heavy thunderstorm rains in upstream areas may affect unsuspecting residents, hikers, and campers in downstream regions where rain has not occurred. This scenario took place in the tragic Antelope Canyon flash flood in northern Arizona during the summer of 1997 when 11 hikers died (Burstein, 1997).

Isolated thunderstorms can also be triggered by the interaction of winds near the ground and mountainous terrain. If winds aloft are weak, a series of very intense storm cells can form in the same location and then move slowly across the same region. This type of situation, although relatively rare, can produce prodigious rainfall totals and severe flash flooding. The floods at Rapid City, South Dakota, in 1972 and in the Big Thompson Canyon in Colorado during summer 1976 were of this type, and rain amounts approached 400 mm in less than 6 h (Maddox et al., 1978). A more recent example is the Rapidan flood in the central Appalachians, which produced 600 mm of rain in a 6-h period from a series of heavy convective showers (Smith et al., 1996). In some island locations (e.g., Hawaii, Taiwan), where winds from the ocean encounter abrupt mountain slopes, unusually moist conditions can cause nearly stationary convective showers that rain so heavily they produce significant flash flooding (Schroeder, 1977; Chen and Yu, 1988). Often these types of orographically forced convective storms occur at night, which further increases the flood hazard. Finally, even in small basins affected by isolated convective storms, under certain synergistic combinations of atmospheric and hydrologic processes, extremely large flash floods can occur. An example is the Eldorado Canyon, Nevada, flash flood of 1974, which was exceptionally large because the convective storm moved progressively downstream through the canyon, causing its intense rainfall to be superimposed on flood waters generated earlier and arriving from upstream (Glancy and Harmsen, 1975).

Mesoscale Systems. Convective clouds, particularly thunderstorms, do not typically evolve as isolated entities. Rather, they tend to organize into narrow lines or bands or into clusters or complexes of individual storms (Hobbs, 1978). These structures tend to be determined by the character of the atmospheric wind fields within which the convective clouds are occurring, by the degree of conditional instability present, and by the size and strength of the synoptic-scale upward motion field within which they are embedded. It is this organized aspect of convective storms that allows mesoscale precipitation systems to be characterized by lifetimes of a number of hours and to generate heavy rainfall over large areas. The heavy rain potential of these systems increases dramatically when they move slowly or become nearly stationary. In a very broad sense, mesoscale convective systems in the midlatitudes can be classified as *precipitation bands, squall lines*, and *mesoscale convective complexes* (MCCs). In tropical regions, mesoscale convective weather systems frequently occur as *convective cloud clusters* associated with easterly waves or tropical disturbances, and *tropical storms* (including hurricanes and tropical cyclones – see Gray, 1968; Pielke, 1990). In some subtropical and tropical regions squall lines and MCCs are also noted (Laing and Fritsch, 1997).

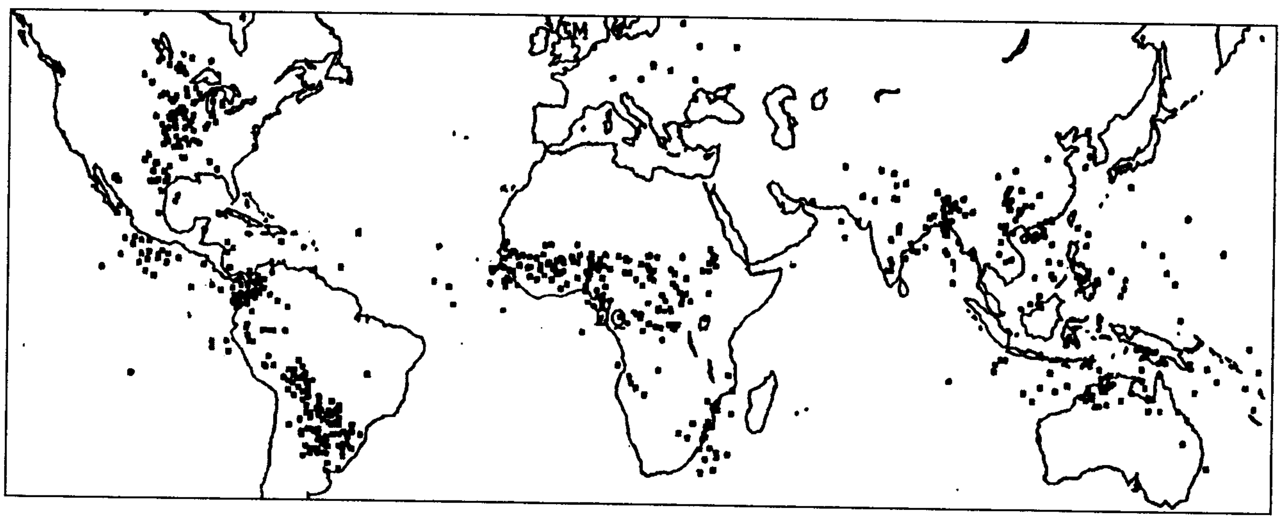
Precipitation Bands. The precipitation areas of synoptic-scale extratropical cyclones are often characterized by mesoscale rainbands that can assume a variety of orientations (Hobbs, 1978). These rain (or snow) bands contribute intensified amounts of precipitation to the overall weather event. The bands result when the circulations within the parent cyclone produce an unstable thermal stratification that will support convection. Mesoscale precipitation bands occur most frequently during the cool season over the oceans and adjacent coastal regions. The convection typically does not grow deep enough to produce lightning, unless the underlying ocean is much warmer than the air (e.g., over the warm Gulf Stream or Kuroshio currents). Occasionally, intense bands of snowfall develop over and downwind of large continental lakes, such as the North American Great Lakes. These bands can produce extreme snow accumulations – amounts greater than 500 mm in a matter of hours – and thus exacerbate snowmelt flooding later in the season. In general, however, mesoscale precipitation bands do not pose immediate local flooding threats; rather, they contribute to the overall synoptic-scale precipitation pattern.

Squall Lines. Long, narrow lines of intense thunderstorm cells are termed squall lines (see Hane, 1986). Squall lines are also usually associated with larger synoptic-scale cyclones and their attendant fronts (the boundary zones separating air masses having distinctly different temperature, wind, and moisture characteristics). The immediate local impact of squall lines may be great because of tornadoes, hail, or damaging wind gusts at the surface. They are most frequent over the continents and usually occur in the warm, moist air ahead of cold fronts. Thus, in the United States, they occur most often over portions of the country where there is a direct source of warm, moist, unstable air (e.g., the Gulf of Mexico) to the south of extratropical cyclone tracks. Squall lines sometimes form independently of synoptic cyclones (e.g., along the slopes of mountain ranges, along mesoscale air mass boundaries, and in tropical and subtropical latitudes).

The main weather threat of squall lines is the locally severe and damaging phenomena produced by intense thunderstorms. Squall lines tend to move rapidly with the winds aloft and thus affect any given location only briefly. Rain rates can be very intense (50–150 mm/h), but the duration of the rains usually are not long enough to cause serious local flooding. An exception occurs when squall lines stall and become nearly stationary (Chappell, 1986). This can happen if the line becomes oriented nearly parallel to the winds aloft. This situation allows many intense thunderstorm cells within the line to move across the same region, leading to very large rainfall amounts (sometimes greater than 300 mm) that accumulate over a few hours. Stationary squall lines therefore pose a great danger of local flash flooding and serious threats to life and property. The hazardous nature of these events is increased because squall lines, if they become stationary, tend to do so at night (Anderson and Arritt, 1998) when flood warnings may not be received.

Mesoscale Convective Complexes. MCCs are large, organized convective cloud systems that lack the distinct linear structure of squall lines (see Maddox et al., 1986). The precipitation structures embedded within MCCs can be very complicated, consisting of short squall lines, rainbands, clusters of thunderstorms, and individual intense thunderstorms as well as widespread areas of steady, light-to-moderate rainfall. Several of these structures are usually present simultaneously in the mature MCC. These systems tend to be less directly related to extratropical cyclones than the other small-scale systems discussed above. They usually occur during the warm season when extratropical cyclones and fronts are weak or have shifted poleward. MCCs occur over many areas of the globe, in both middle and low latitudes (Figure 2.2), where key ingredients are all present. These regions tend to be distant from synoptic-scale storm tracks downwind (with respect to prevailing middle-level winds) from significant mountain ranges and are usually affected frequently by low-level jets, which are strong inflows of near-surface winds and high humidity from lower latitudes that provide abundant water vapor to fuel the convection. MCCs frequently deliver rain rates and amounts similar to those produced by stationary squall lines. These features are often slow moving; hence, they pose a significant threat of heavy rains and local flash flooding (e.g., the damaging Johnstown flash flood of 1978) (Hoxit et al., 1982). They can also produce large hail, high winds, and extremely frequent lightning, which increases their hazardous character. If, over a period of days to weeks, a succession of MCCs develops in, and tracks across, roughly the same area, the regional flood hazard increases significantly (Fritsch et al., 1986). It was

Figure 2.2. Global distribution of mesoscale convective complexes (MCCs) based on regional samples of MCC locations obtained from satellite imagery over 1–3 years. (Source: Laing and Fritsch, 1997)



through this scenario that MCCs contributed to the Great Flood of 1993 in the Upper Mississippi River basin (Junker et al., 1995).

Tropical Systems. In the tropics, short-wave troughs in the easterly wind flow (i.e., easterly waves) are typically manifested as convective cloud clusters. Clusters pose little flooding threat unless they become nearly stationary or occur with unusual frequency over the same area. Some convective clusters interact with the environment in ways that lead to an increase in strength of the system. This local intensification can lead to tropical disturbances and more fully developed tropical storms and tropical cyclones (hurricanes, typhoons, etc.) that can amplify to great intensity and persist for days. The greatest hazards from tropical cyclones arise from the damaging winds and disastrous storm-surge flooding they produce in low-lying coastal areas. They can deliver huge amounts of rainfall (i.e., greater than 500 mm) if they stall or move slowly, causing severe flooding threats, particularly over the small drainages of mountainous ocean islands. Further, when storms move inland they sometimes become very slow moving and interact with local terrain features (e.g., Carcena and Fritsch, 1983). The principal threat becomes widespread heavy rains along the storm's path. Such events, for example hurricane Agnes in the eastern United States during summer 1972 (see Bailey et al., 1975), can produce widespread flooding including localized flash floods and large-basin floods. These events pose a severe risk to property and human life. Land-falling tropical storms also often produce numerous tornadoes, adding a severe weather threat to that of the strong winds and heavy rains. Finally, under certain circumstances, tropical storms can play a role in generating floods in areas distant from the storm itself or in areas affected by the dissipating stages of the storm (Figure 2.3c). This occurs when large amounts of water vapor associated with the decaying storm support development of other precipitation systems, such as one of the mesoscale convective systems described above.

Larger Scale Precipitation Systems: Synoptic and Macroscale

The larger-scale processes that produce heavy rains are often associated with anomalous and persistent middle-level atmospheric wave patterns, slow moving or stagnant features such as blocking anticyclones and cutoff lows, or nearly stationary synoptic frontal zones. When such features also import air, usually from lower latitudes, having very high absolute moisture contents, expansive regions of heavy precipitation can occur (Means, 1954). Rain amounts of 50 to 150 mm can be delivered over a period of 2 or 3 days over regions spanning thousands of square kilometers. Macroscale features of the atmosphere, such as large-scale wave patterns aloft, tend to help set the stage for the occurrence of heavy rain events on synoptic and smaller scales. In addition, large-scale interactions between the atmosphere and the earth's surface also occur at the macroscale and can affect extensive regions of the globe on seasonal or longer time scales by means of monsoonal circulations and ocean-atmosphere teleconnections, such as the El Niño/Southern Oscillation (ENSO).

Synoptic-Scale Systems. At the synoptic scale, ridge and trough short-wave patterns aloft coupled with slow-moving pressure centers (both highs

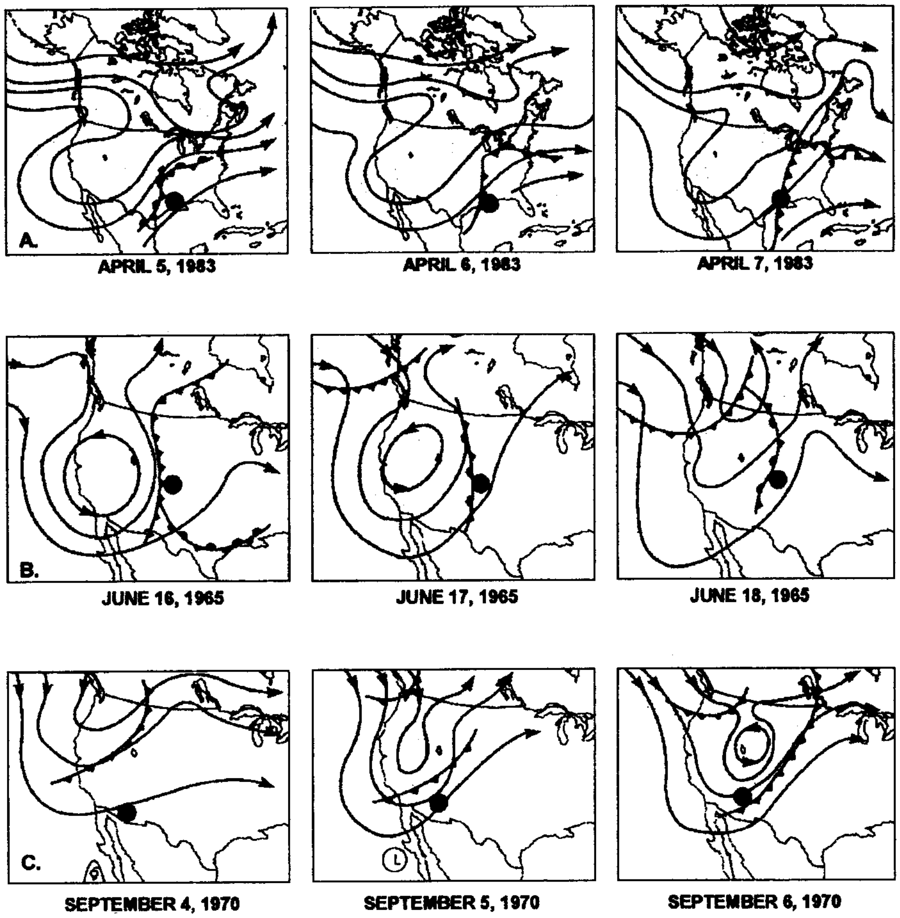


Figure 2.3. Three-day sequences of persistent or slowly changing large-scale circulation patterns associated with some severe floods in the United States. The direction of middle-level flow is depicted with arrows; surface frontal positions are also shown. Black circles indicate where flooding occurred. (A) Quasi-stationary trough-and-ridge pattern and surface frontal activity led to significant flooding in southern Louisiana. (B) Cutoff low within quasi-stationary large-scale trough over western United States steered warm unstable air northward into eastern Colorado, where it interacted with terrain. This, coupled with a shortwave perturbation on June 17, led to an episode of extreme flooding in east-central Colorado. (C) Interaction between a slow-moving large-scale trough, a surface front, and moisture from dissipating tropical storm Norma, led to severe flash flooding in central Arizona. (Modified from Hirschboeck, 1991)

and lows) and associated fronts can cause very heavy rains (see, for example, Matsumoto et al., 1971). This often occurs with continued inflow of very moist air from low-level jets, causing persistent rains along frontal zones. Synoptic-scale heavy rains can occur along and ahead of cold fronts or along and to the cool side of warm and stationary fronts. This type of situation can also indirectly contribute to heavy rains by setting up an environment very conducive to the development of smaller-scale precipitation systems. The key aspect of synoptic heavy rain events is that the synoptic pattern is changing slowly enough to allow winds from low latitudes to import large amounts of water vapor into the weather system (see Figure 2.3). The floods associated with synoptic weather features are usually widespread, so flooding occurs over large river systems during an extended period. However, this does not imply that very localized regions of enhanced rain, which generate local flash floods, cannot be embedded within synoptic events. It is also important to remember that intense synoptic systems in the winter can set the stage for future flooding by depositing very heavy snowfalls over large areas. The macroscale and synoptic scales often interact in the winter to establish a persistent, weeks-to-months storm track that allows many events to build large snow packs.

Perhaps the most important synoptic-scale features associated with heavy rains are *extratropical cyclones* (i.e., lows, low pressure systems, or depressions) and their associated waves (i.e., troughs) in the temperature and wind patterns aloft. These flow perturbations, generally called *short-wave troughs*, can occur on both the synoptic and mesoscale, depending on wavelength. They are distinctly different from macroscale waves in that they tend to move with the wind flow aloft and to evolve through a life cycle of formation, intensification, and decay during periods of several days to a week or so. It is these short-wave troughs that most often produce upward motions strong enough to generate regions of precipitation or to trigger formation of smaller-scale, more intense, weather features. Fronts associated with extratropical cyclones are also capable of producing heavy rains. It is important to note that atmospheric fronts tend to be both synoptic and mesoscale in nature. This is because fronts can extend horizontally for several thousand kilometers, whereas the temperature, humidity, and wind contrasts that define the front are confined most clearly in zones across the front that are only several hundred kilometers in extent. Synoptic cyclones and their attendant fronts can produce broad areas of general upward motion in the atmosphere or trigger the development of embedded smaller-scale weather features and storms. Often, the heaviest rains occur in extratropical cyclones in regions where both extensive precipitation, due to the cyclone's general lifting of the air, and embedded small-scale, intense rains occur (Maddox et al., 1979). On rare occasions, a synoptic scale short-wave trough can move over a region of surface high pressure without causing development of a surface cyclone. If such a feature imports moist, humid air aloft, widespread heavy rains and snows still occur, even though there is no distinct surface cyclone, or low pressure center, present. During the cold season extratropical cyclones often produce wide bands of very heavy snowfall.

Macroscale Systems. Most major regional floods are associated with circulation anomalies that can be observed in the large-scale wave patterns of middle-level geopotential height fields (e.g., 700 or 500 mb level). These wave patterns aloft may range in scale from the synoptic to the macroscale depending on wavelength. Macroscale waves can support the development, persistence, or sequential recurrence of smaller-scale precipitation systems that produce floods. Moreover, the most extreme and widespread regional floods usually evolve from macroscale circulation anomalies that are characterized by exceptional persistence in their embedded synoptic-scale wave patterns (Figure 2.3). In particular, circulation anomalies involving quasi-stationary patterns such as blocking ridges and cutoff lows in the middle-level flow are especially likely to be associated with extremely large meteorologic floods (Hirschboeck, 1987a, 1991). Periods when the patterns of macroscale troughs and ridges in the middle-level atmospheric circulation remain blocked in approximately the same configuration for several days can lead to nearly stationary frontal patterns, allowing extended episodes of storm activity and precipitation (Figure 2.3a). Persistent wave patterns can also result in long periods of flow impinging on mountain ranges. If the winds are carrying high amounts of moisture, then heavy rainfall can occur along upwind slopes (Figure 2.3b). In general, however, heavy rains during periods with persistent circulation patterns occur in direct association with synoptic- and smaller-scale weather features embedded within the flow regime (Figure 2.3b), or interacting with it (Figure 2.3c).

In addition to macroscale wave patterns aloft, there are other macroscale systems or processes that can influence flood activity over extensive or distant regions of the globe. One such system is located in the tropical latitudes (~15°N to ~15°S), where convective rainfall is common throughout the year, within the *intertropical convergence zone.* This macroscale feature reflects the zone where the trade winds from the northern and southern hemispheres converge. It is characterized by extensive cloudiness and embedded easterly wave disturbances with clusters of deep convective clouds and rains (Gray, 1968). Tropical storms sometimes develop to the poleward side of this zone as well. Thus, the intertropical convergence zone essentially provides a globe-encircling moist and favorable environment for the occasional development of flood-producing heavy rains, which are usually delivered by smaller-scale features.

Other macroscale processes that affect flooding at continental or global scales derive from interactions between the atmosphere and the earth's surface (land or ocean) and occur as *monsoonal circulations.* Several regions of the world are affected by monsoon flow regimes. Simply stated, these are circulations characterized by cold continental air flowing off the continent during the winter and warm humid air from a tropical or subtropical ocean flowing onto the continent during the summer. Unusually strong summer monsoons can lead to very widespread and extreme amounts of rainfall. This is particularly true when a strong summer monsoon flow regime develops in regions with steep gradients of elevation, such as the Indian subcontinent, where the northeastward onshore flow encounters first the western Ghats and then the rapidly rising mountain massifs of the Himalayas. Monsoonal precipitation regimes are strongly seasonal. For example, in the Narmada River basin in central India (Figure 2.4), 90% of the 1250-mm mean annual rainfall falls during 4 months of the year (June through September) (Dhar et al., 1985). There also can be a great deal of interannual variability in the monsoon precipitation, as seen in the all-India monsoon rainfall record (Figure 2.5).

Although monsoon circulations can be viewed as macroscale features that are driven by large-scale atmosphere–ocean interactions, the monsoon rainfall itself is not delivered by a single large precipitation system but is composed of a series of discrete rainfall events that operate at smaller scales (Webster, 1987) Synoptic and mesoscale features embedded in the macroscale monsoon circulation play an important role in determining variations in interannual rainfall totals and in delivering heavy rainfall sufficient to cause floods. At times, MCCs develop within the general monsoon flow and affect monsoonal regions of Africa, India, Asia, Australia, and all the Americas (Laing and Fritsch, 1997) (see Figure 2.2), enhancing the likelihood of flooding.

Disturbances in the tropical easterly flow also can contribute to heavy rainfall during the monsoon season. An example of this type of feature can be seen in the low-pressure systems and monsoon depressions that form in the Bay of Bengal and move west-northwestward across India from June through September. These can lead to significant flooding in the interior of the subcontinent when their tracks move far inland along the axis of an east–west-oriented drainage basin, such as the Ganges or Narmada. For example, in the Narmada River basin, large floods can occur when several low-pressure systems track successively over the upper part of the basin or when a single system moves downstream with a trajectory

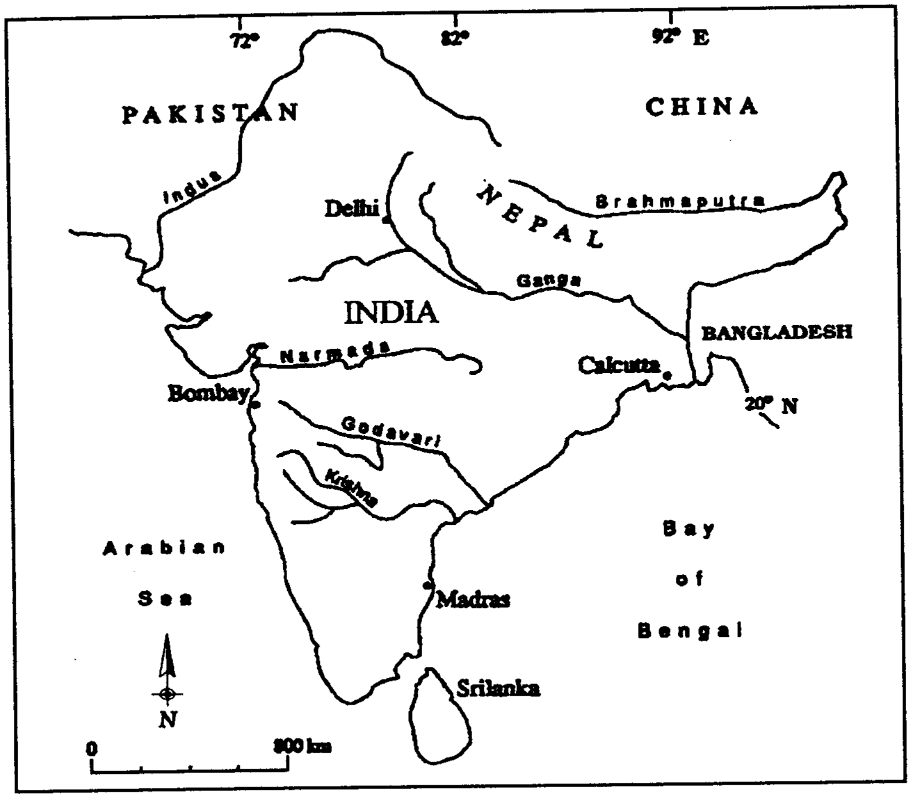


Figure 2.4. Map showing major rivers in India.

that is oriented along the axis of the basin, superimposing intense rainfall on the downstream-moving flood wave (Hirschboeck et al., 1996). In the postmonsoon period, intense tropical cyclones may also deliver heavy rains to the Indian subcontinent, further contributing to the flood hazard.

Monsoon rainfall also varies spatially – both intra- and interannually – due to shifts in larger scale features, such as the middle-level monsoon trough that forms over India. The position of this feature varies throughout the monsoon season, resulting in shifts in the location of heavy rainfall associated with monsoon disturbances embedded in the trough (Das, 1987; Bhalme and Mooley, 1980; Singh et al., 1988). Despite interannual variations in the strength of the Asian monsoon, some areas under its influence experience flooding on nearly an annual basis. In Bangladesh, for example, the discharges of the Brahmaputra and Ganges River basins (Figure 2.4) lead to a periodic inundation of the heavily cultivated and populated low-lying, delta regions of the combined basins. However, within the upstream portions of these basins, a strong spatial variation in the receipt of monsoon rainfall has been observed by some researchers. When the Ganges River basin to the west receives lower monsoon rainfall, the Brahmaputra River (and its lower tributary, the Meghna River) tend to receive above average rains and vice versa (Bandyopadhyay et al., 1997). Most of the severe flood years in Bangladesh (e.g., 1974, 1987, 1988) appear to be influenced by heavy rains and runoff from the Brahmaputra and Meghna basins, not the Ganges (Bandyopadhyay et al., 1997). Moreover, rainfall patterns in the upstream Himalayan parts of these basins appear to have less impact on severe Bangladesh flooding than does a synergism

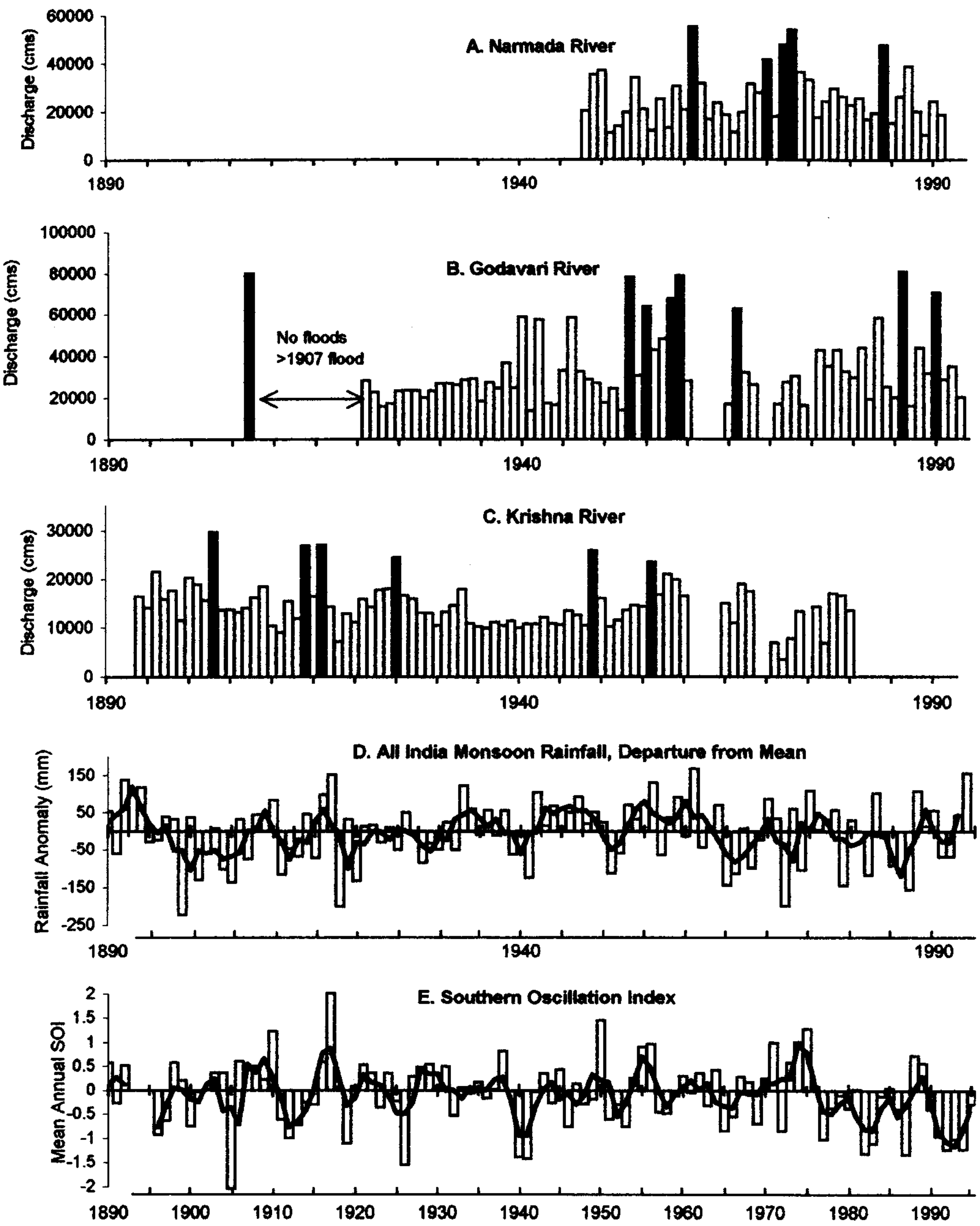


Figure 2.5. Peak annual gaged discharge records for the three largest rivers in central India compared with standardized all-India monsoon rainfall and the Southern Oscillation Index (SOI) over the last century. The period of above-average monsoon rainfall and moderate SOI in the middle of this century was one of few large floods on these rivers, as indicated by these gage records and paleoflood records (see text for more detail). (A–C) Dark bars highlight the largest five to seven floods on each river. Discharge data for the Narmada, Godavari, and Krishna Rivers compiled from UNESCO (1969–79, 1976), Rodier and Roche (1984), and Indian National Institute of Hydrology personal communication (1991, 1994). (D) All-India monsoon rainfall shows annual departure from 1874 to 1994 mean [data from Parthasarathy et al. (1993) and B. Parthasarathy, Indian Institute of Tropical Meteorology); dark line is 3-year running mean. (E) Annual variations in SOI, based on standardized data from NOAA Climate Prediction Center; dark line is 3-year running mean.

between more locally heavy monsoon rainfall (e.g., note the high concentration of MCCs in this area in Figure 2.2), combined with climatologic and hydrologic factors already in place in the lower reaches of the river systems. These flood-enhancing factors include high groundwater tables, low relief, infrastructure-related drainage impediments, and those occasions when a simultaneous inflow of above-normal discharge occurs from both the Ganges and Brahmaputra Rivers (Bandyopadhyay et al., 1997).

Because monsoons arise from large-scale seasonal land–ocean–atmosphere interactions, there is some evidence that they also may be linked globally through *teleconnections*. Teleconnections are statistically defined atmospheric and oceanic interactions between widely separated regions. For example, the Southern Oscillation Index (SOI) defines a teleconnection based on an inverse relationship in sea level pressure between two locations in the tropical Pacific Ocean (Darwin minus Tahiti pressure). Variations in this index are associated with the occurrence of anomalously warm (El Niño) or cold (La Niña) sea-surface temperatures in the Pacific. The combined atmosphere and ocean components of the phenomenon (ENSO) can be viewed spatially as macroscale phenomena, especially because various other responses throughout the globe have been statistically linked to ENSO variability. Furthermore, the influence of ENSO can be translated from tropical to extratropical latitudes and affect macroscale planetary long-wave patterns, which may in turn drive other teleconnections, such as the Pacific North American teleconnection pattern (Horel and Wallace, 1981). Temporally, teleconnections operate over a range of time periods. The ocean–atmosphere processes and interactions involved generally develop over a period of weeks to months and may persist for a year or longer. Hence, even though the influence on flooding of a teleconnection such as ENSO may be manifested at short time scales through meteorological processes that produce heavy rains in given areas of the world, teleconnections themselves operate on much longer climatic time scales and are discussed in more detail later.

Antecedent Climatic and Hydrologic Factors That Indirectly Cause Flooding

Floods are relatively rare and each one tends to arise from a unique set of ingredients. For example, the key atmospheric ingredients that are needed for flash-flood-producing storms are sustained high rainfall rates generated by forced ascent of air containing substantial water vapor (Doswell et al., 1996). However, even when sustained heavy rainfall occurs from some of the meteorologic systems described earlier, it may not always be enough to cause flooding without a synergistic combination of factors related to climatic and hydrologic conditions already in place in a drainage basin. To predict floods effectively, meteorologic information must be coupled with a knowledge of the rainfall-runoff processes, surface and subsurface hydrologic processes (Ramirez, Chapter 11, this volume), and land-use properties of the basin of interest. In addition, antecedent conditions must be considered to determine the likelihood of flooding from a given rainfall amount. Antecedent climatic and hydrologic factors that can influence runoff and subsequent flood occurrence are the degree of soil saturation and the level of shallow groundwater tables, which are affected by earlier precipitation events; the amount of snow accumulation from previous storms; the rate of snow melt, which depends on the length and degree of warm intervals; and the depth of frozen soil, which is related to the timing and severity of prior cold spells. In large basins with dams and reservoirs, the amount of available reservoir storage is another important factor that can influence future flooding.

Soil Moisture and Subsurface Water. Soil moisture, soil saturation, and groundwater levels have a variable influence as antecedent determinants of flooding when coupled with a heavy rainfall event. Saturated soils are not a required precursor to severe flooding. For example, flash floods commonly occur when runoff is conveyed to stream channels rapidly as *Hortonian overland flow,* regardless of soil moisture content (Dunne, 1983). Hortonian flow occurs when rainfall intensity exceeds the infiltration capacity and is associated with arid regions; urbanized areas; shallow, impermeable or disturbed soils; bedrock exposures; and sparsely vegetated areas. Meteorologic factors also play a role in whether Hortonian flow takes place. Heavy, convectively driven rainfall from thunderstorms can be so intense that it falls faster than it infiltrates, thereby generating overland flow, even where soils are unsaturated or the overall rainfall totals are not very high. This is common in arid and semiarid regions where investigators have found that thunderstorm-generated runoff is not very sensitive to antecedent soil moisture (Goodrich et al., 1994; Michaud and Sorooshian, 1994). It also has been noted that significant flash floods in arid regions of the western United States can be generated with much less rainfall than is required to generate a similar flash flood in wetter parts of the country (Maddox et al., 1979). In more humid regions, Hortonian flow can occur when soils are wet from previous storms (but not yet saturated) because the wet soils can decrease the infiltration capacity. In addition to Hortonian overland flow, runoff may rapidly enter stream channels as *subsurface storm flow,* which occurs during storm events when water infiltrates and very quickly moves laterally through the soil to enter the stream channel (Dunne, 1983). Because of the rapidity of both the Hortonian and subsurface storm flow processes during intense rains from convectively driven meteorologic processes, floods are possible when soils are not saturated or even wet. Hence, in some environments, antecedent soil moisture conditions may or may not significantly affect flood occurrence.

In certain hydroclimatic settings, saturated soils and shallow groundwater tables do play a critical role in causing and exacerbating floods. When rain falls onto saturated areas adjacent to streams, runoff can be generated rapidly at the surface as *saturation overland flow* (Dunne, 1983). During sustained periods of rainfall, the areas contributing saturation overland flow within a drainage basin grow larger as shallow water tables rise to the surface. Saturation overland flow is most likely to occur in watersheds that contain thin soils, concave hillslopes adjacent to wide valley bottoms, and perennial flow. It is also likely to be limited to humid areas and occur only after a lengthy episode of antecedent precipitation or when soil moisture is at a maximum at the end of the snowmelt season. The role of soil moisture, saturated soils, and shallow groundwater tables is most evident when storms have repeatedly rained on a watershed so that nearly all natural and artificial surface and subsurface storage reservoirs are full. Under such conditions, even a small amount of rainfall can result in a major flood hazard.

Snow and Snowmelt. In cold climates and at high elevations, where snow accounts for a significant portion of the total annual precipitation, the magnitude of the seasonal snow accumulation and its rate of melting are important antecedent factors that influence flooding. The stored snowpack can be released as runoff gradually or episodically throughout the winter and spring seasons, but when it is released rapidly during an abrupt thaw the flood hazard is increased. When portions of rivers have been frozen, ice breakup and ice jams (Cenderelli, Chapter 3, this volume) also present a severe backwater flood hazard, as occurred during the 1997 floods on the Red River of North Dakota. It is during and right after the spring snowmelt season that soil moisture reservoirs are usually filled to capacity, compounding the flooding problem from saturation overland flow. At such times, and during intermittent thaws throughout the winter, the presence of frozen ground, which prevents infiltration, is another flood-enhancing factor. Rapid, flood-producing snowmelt will take place over a span of several exceptionally warm and sunny days, especially when a persistent high-pressure anomaly, often associated with a blocking wave pattern, has established itself over an area previously visited by a sequence of cold low-pressure systems that precipitated large volumes of snow.

Some of the largest snowmelt-related floods occur when rain falls on an antecedent snow cover. The effect of rain on snow depends on the characteristics of the snowpack, including its temperature and the temperature of the rain itself (Harr, 1981). Early in a snow season before the snow compacts, rain falling on snow can be absorbed by a snowpack, adding to its total water content but not producing snowmelt. Later in the season when the snow is more dense, rainfall is less able to be absorbed. If the rain falling on the snow is warm, it is believed to accelerate the melt rate; however, the heat transferred to the snowpack during condensation of water vapor on the snow's surface may be another source of heat for the snowmelt process (Harr, 1981). In either case, it is during a prolonged, warm rainfall event that snowmelt can be a major contributor to flood runoff. During such an event, a large amount of water from both the current rainfall and the accumulated water content from the prior season's snowfall events can be rapidly introduced into a drainage system. The threat of severe flooding is compounded when the basin also contains areas of both frozen and saturated soils.

Flood Hydroclimatology and the Long-Term Perspective

It is possible to forecast or monitor a developing flood situation over meteorologic and short climatic time scales, but other approaches are needed to address flood hazards over longer time scales. Hydrologists have long used time series of gaged flood-peak data to evaluate the probability of occurrence of a flood of a given magnitude. Flood frequency analysis (i.e., the statistical analysis of observed flood events to estimate flood magnitudes at different recurrence intervals) constitutes a basic component of flood hazard zone determination in the United States and many other countries

(Stedinger, Chapter 12, this volume). A basic underlying assumption in this statistical approach to evaluating flood variability is that the data values in a flood time series represent an array of flood information that "is a reliable and representative time sample of random homogeneous events" (U.S. Water Resources Council, 1981, p. 6). Implicit in this assumption is the notion that flood-flow values are stationary and not affected by climatic trends over time and that the record represents a sample of time-homogeneous events, even though the floods may have originated from different types of hydrometeorologic causes. From a physical-process perspective it has been acknowledged that climatic variability may indeed affect flood-flow values in a nonrandom way and that flood series often are not homogeneous but composed of mixed distributions of events originating from different causal mechanisms (U.S. Water Resources Council, 1981; Hirschboeck, 1985, 1987b, 1988). However, despite the widespread recognition of these possible violations of the random homogeneous events assumption, the practical implications of these concerns for flood hazard perception, assessment, prevention, and readiness are only beginning to be explored. From a statistical test standpoint, violations in the randomness and time-homogeneity assumptions often are not evident; yet from the perspective of flood hydroclimatology, important and useful information for flood hazard management can be obtained from both a hydrometeorologic and hydroclimatic appraisal of a gaged flood record. In addition, flood evidence prior to the observed record (Baker, Chapter 13, this volume) can be used to extend our hydroclimatic understanding even further back in time. In the following sections we address how the flood hydroclimatology framework can be used to explore some of the underlying meteorologic and climatic causes of long-term flooding variability.

Climatic Information in Gaged Records: Multiple Mechanisms for Generating Floods

In many areas of the world, especially in the middle latitudes, floods in a single drainage basin can be generated by different types of precipitation events. The type of event that occurs depends in part on the season but also may depend on factors related to the large-scale circulation environment. In the middle latitudes of North America, for example, individual convective storms and mesoscale convective systems are predominant summer flood producers, whereas extratropical cyclones and their associated fronts are more likely to generate winter, spring, or fall floods. Tropical storm flooding is generally limited to summer and fall. In lower latitudes and warmer climates with sufficient moisture, convective storm systems may prevail year-round. Because the intensity and duration of heavy rainfall varies in different types of precipitation systems, the flood peaks in a given stream may reflect this, resulting in hydroclimatically defined mixed distributions in the overall probability density function of flood peaks (Hirschboeck, 1985, 1987b). Figure 2.6 illustrates this for two rivers in the Arizona portion of the Lower Colorado River basin, both of which experience flooding from a variety of precipitation systems (Hirschboeck, 1985, 1987b), which have been grouped into three general categories: storm-scale or mesoscale convective precipitation systems (e.g., mostly summer thunderstorms), tropical storm-related precipitation systems,

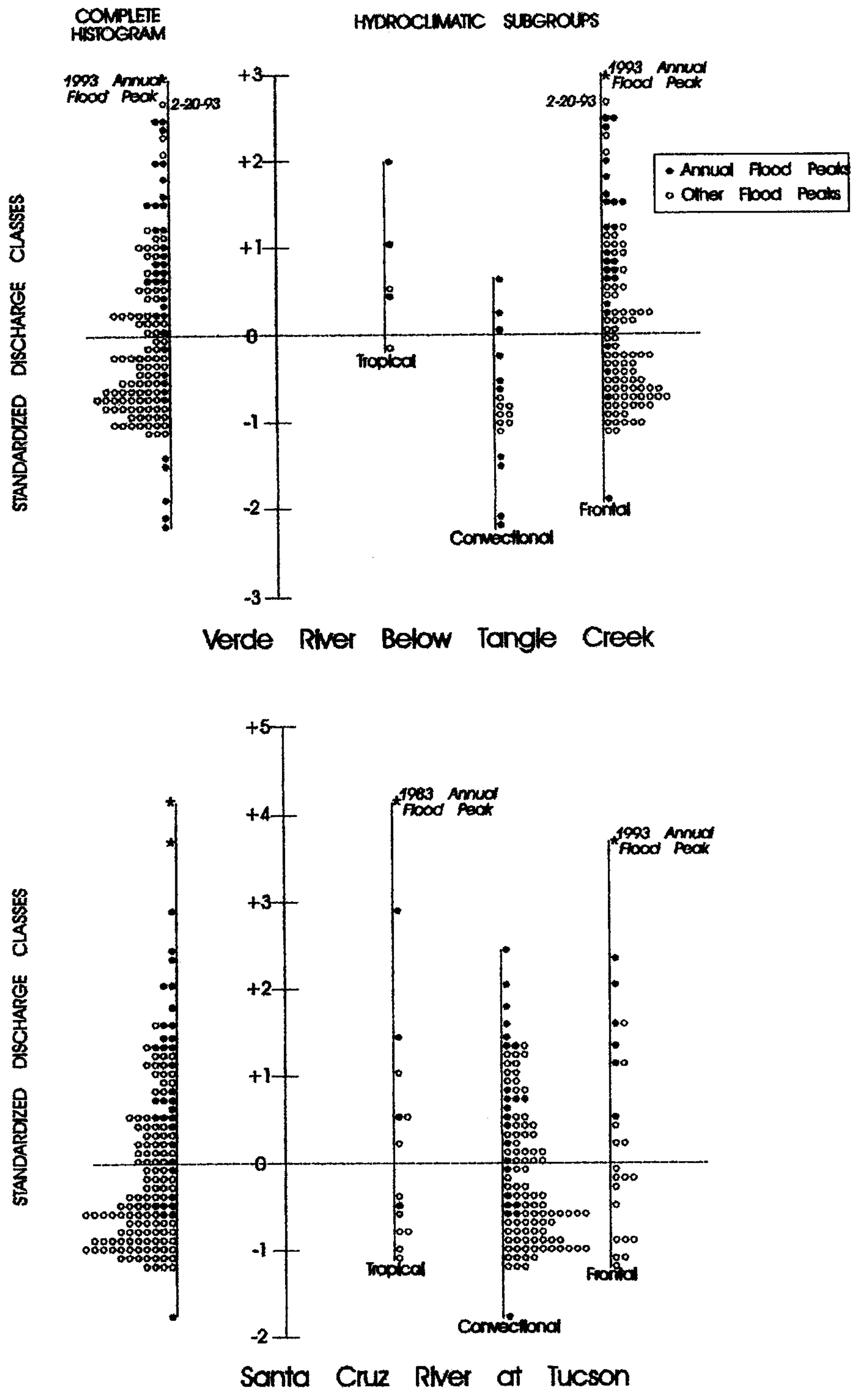


Figure 2.6. Decomposition of two flood records from the lower Colorado River basin according to the type of precipitation system that produced each flood. Peak discharges are displayed in the form of standardized dimensionless *z* scores for comparison. Annual flood peaks are shown as solid circles and all other peaks above base are shown as open circles. The period of record is 1950–1985; the extreme flood peaks of winter 1993 are also plotted. Verde River data are for the Verde below Tangle Creek (USGS gage no. 09508500). Santa Cruz River data are for the Santa Cruz at Tucson (USGS gage no. 09482500). (Source: House and Hirschboeck, 1997)

and synoptic-scale extratropical cyclone–frontal precipitation systems (House and Hirschboeck, 1997).

The distribution of annual and peak-above-base floods for the Verde River in central Arizona shows the dominance of synoptic–frontal floods for shaping the river's overall flood probability distribution. The Verde basin is situated at a latitude that is regularly influenced by winter storms and frontal passages and these cold-season storms yield seasonal precipitation totals that are larger than those received by the basin in summer from convective events. Hence, in the Verde, floods produced by synoptic–frontal events are more frequent than those produced by either of the other types of systems. These winter synoptic–frontal floods are also responsible for the largest peaks of record. Furthermore, when floods in the Verde are produced by rare tropical storm-related precipitation systems, the size of these events tends to be larger than the smaller-scale convective storm events, larger than the mean flood of the series, and comparable to some of the higher-magnitude frontal floods. Compared with the Verde, the Santa Cruz River is located in southern Arizona – south of the typical winter storm track – and the most frequent flood-producing precipitation systems are small-scale summer convective systems, which are especially active in southeastern Arizona. Note, however, that the largest floods in the Santa Cruz record were produced by tropical storm-related systems or winter synoptic–frontal systems, even though these types of events occurred less frequently.

The separation of these two relatively short flood records (~35 years) into their hydroclimatically defined components adds process-based meteorological and geographical information, which can supplement a purely statistical analysis of the flood peaks in these rivers by revealing the kinds of weather systems that produce different magnitudes of floods in each. Such an approach can also be used to evaluate the causes of flood variations over time – by examining changes in the decade-to-decade frequencies of floods produced by different types of storms in several streams throughout a region (Hirschboeck, 1985, 1987b) or by evaluating a long flood record of a single stream in detail (Webb and Betancourt, 1992).

On a larger scale, regional floods are often repeatedly associated with characteristic synoptic and macroscale atmospheric circulation patterns (Hirschboeck, 1985, 1991). For example, composite maps of daily 700-mb pressure heights over the eastern North Pacific Ocean and western North America show that severe winter floods in six different subregions of southern California, Arizona, Utah, and Nevada are all associated with a low 700-mb height anomaly off the California coast and a high 700-mb height anomaly near Alaska. Relatively minor shifts in the position of these anomalies appear to control which subregions experience floods (Ely et al., 1994). Hence, an analysis of variations in the position, intensity, or frequency of these large-scale atmospheric circulation patterns can improve our understanding of the underlying hydroclimatic causes for the spatial distribution of flooding.

Climatic Information in Gaged Records: Teleconnection Signals

The effect of ENSO on interannual variations in regional precipitation and streamflow has been a focus of a variety of research efforts across the globe (e.g., Ropelewski and Halpert, 1986, 1987; Cayan and Peterson, 1989; Gregory, 1989; Redmond and Koch, 1991; Cayan and Webb, 1992; Lins, 1997). However, the potential connection between these ENSO-related interannual variations in climatic conditions and the frequency or severity of extreme flood events is more difficult to document. In at least one area of the world the ENSO–precipitation–flood relationship is straightforward: along arid northern coastal Peru all but one recent strong El Niño event – that of 1976 – corresponded to both high rainfall and flood peaks (Waylen and Caviedes, 1986; Wells, 1990). But in other regions of the world that exhibit an ENSO link with precipitation, a corresponding relationship to floods may not be evident. High seasonal precipitation alone is not enough to cause a severe flood because of the synergism needed for many specific hydrologic and meteorologic factors to occur in just the right combination and timing to create the flood peak. In one extremely long gaged record from 1824 to 1973, low annual peak discharges on the Nile River resulting from below-average Indian–African monsoon precipitation showed a correlation with strong El Niño years (Parthasarathy and Pant, 1984; Quinn, 1992). However, the relationship between high peak annual floods and interannual climatic variations is rarely as straightforward as that between years of low discharge and drought. As noted above, there are only a few areas of the world where floods and precipitation are integrally linked to a single climatic variable or causal mechanism to the extent that a strong, consistent correlation appears on an interannual scale. Most regions are affected by multiple hydroclimatic mechanisms for generating floods and this complicates the relationship of floods with the interannual variations of a single regional climate component.

In the Arizona portion of the lower Colorado River basin a positive correlation exists between El Niño conditions and precipitation from winter extratropical cyclones and fronts and early fall tropical storms (Andrade and Sellers, 1988; Redmond and Koch, 1991). Because of the demonstrated effect of El Niño on the types of precipitation systems that produce the largest floods in Arizona (see Figure 2.6), one would logically expect a correlation between large floods and El Niño years. The single largest floods on some of the rivers in the lower Colorado River basin have indeed occurred during strong El Niño events (Cayan and Webb, 1992), although the record-breaking floods in Arizona in winter 1993 were associated with only a moderate El Niño (House and Hirschboeck, 1997). In general, there appears to be some connection between floods and El Niño in the lower Colorado River basin, even though the sample size may prohibit a statistically significant correlation (Webb and Betancourt, 1992; Ely, 1997). Still, not all El Niño years exhibit identical regional responses in precipitation and flooding (Schonher and Nicholson, 1989). Large floods have occurred during non-El Niño years, and some strong El Niño years (e.g., winter 1997–98) have not been characterized by significant flooding in the lower Colorado River region.

In another ENSO-teleconnected region, central India, summer-monsoon precipitation shows a positive correlation with the positive SOI (non-El Niño) phase of ENSO (Parthasarathy and Pant, 1984). However, the largest historic floods on the major rivers in central India (Narmada, Tapi, Godavari, Krishna) do not occur exclusively during the years of high monsoon precipitation or strong positive SOI (Ramaswamy, 1987; Dhar and

Nandargi, 1993). In fact, several of the largest floods in the region were actually associated with El Niño conditions and years of rainfall deficiency (Ely et al., 1996; Kale et al., 1997). As noted earlier, major inland floods in this region are caused by low-pressure systems and monsoon depressions that move west-northwestward from the Bay of Bengal and over the Indian subcontinent (Mooley and Shukla, 1987). Although these systems occur during the monsoon season, there does not appear to be a strong link between their flood-producing capabilities and ENSO.

These examples show that a relationship between ENSO and flooding is geographically variable, and, in those areas where such a relationship can be detected, it may not be consistent or statistically robust. Nevertheless, because there is an established connection between ENSO variability and seasonal precipitation in several areas of the world, there is considerable interest in using this information to issue climatic and hydrologic forecasts up to a season in advance. As in the case of the Nile River discharge, such forecasts may be far more reliable for low-discharge years than for extreme flooding years (Quinn, 1992); yet, some applications for flood forecasting are possible. For example, in river basins of the western United States, where annual peak flooding depends on the previous winter's accumulated snowpack at high elevations and where a strong ENSO sensitivity to winter precipitation has been identified, long-range hydrologic forecasts for the upcoming flood season can be issued. However, in any given year, the circulation patterns that regulate the rate of snowmelt will ultimately determine the severity of flooding that occurs in such basins. If we are to improve our understanding of the linkages between regional flooding and teleconnections such as ENSO, we need to understand the full range of variability that can exist within regional ENSO–precipitation–flood relationships. One means of doing this is to examine these relationships over much longer time periods than gaged flood records provide.

Bridging from Gaged Records to Longer Periods

Projected anthropogenic climatic changes as well as natural shifts in climatic conditions are potentially capable of affecting hydrologic systems to a degree that exceeds the range of conditions experienced within the relatively short observational records of most rivers (Houghton et al., 1996). Flood hazards are no exception to this prediction. Past or future changes in climatic conditions can be described in terms of changes in long-term means of precipitation and temperature, but the heterogeneity of the weather processes that compose climate dictates that there will be significant variance from these means, including changes in the frequency or intensity of extreme flood-generating storms or droughts. Floods have more severe effects when superimposed on already high streamflows, and thus a change in average seasonal climate and streamflow conditions could be associated with increased flood hazards. Paleohydrologic research has indicated that large changes in flood frequency and magnitude on river systems in several different hydroclimatic regions have been contemporaneous with small changes in mean temperature and precipitation over the past few thousand years (Chatters and Hoover, 1986; Knox, 1993; Ely et al., 1993; Ely, 1997). Traditional statistical approaches to estimating flood magnitudes and frequencies may break down over long time periods when the circulation patterns and processes that drive flooding variability change or shift over time, rendering assumptions of stationarity and random homogeneous time series invalid. In their place, a hydroclimatic understanding of the sources of flood variability may provide alternative or supplementary ways to estimate the likelihood of extreme floods. For example, in the lower Colorado River basin, where research has been conducted on links between circulation patterns and floods in the gaged record, additional research on the long-term spatial and temporal patterns in the occurrence of large floods in this region shows that a consistent relationship between floods and climatic variations appears to hold true for at least the past 5000 years (Ely, 1997).

Decadal-Scale Variations over the Twentieth Century. The period of historical climatic records, generally the last century at most, has been characterized by regional variations in temperature and precipitation persisting for periods up to several decades (Diaz and Quayle, 1980; Barry et al., 1981; Balling and Lawson, 1982; Webb and Betancourt, 1992). Many of these decadal-scale climatic fluctuations have coincided with shifts between two general states of macroscale wave patterns – one dominated by more frequent occurrences of zonal flow and the other dominated by more frequent meridional flow patterns, manifested as either recurring or persistent ridges and troughs. From the 1930s through the 1950s in the northern hemisphere, zonal flow occurred somewhat more often, whereas the periods before and after experienced a tendency toward more frequent periods of meridional flow (Dzerdzeevski, 1969; Kalnicky, 1974; Hirschboeck, 1988; Webb and Betancourt, 1992). Changes in climatic and hydrologic systems have been observed at similar times in several parts of the globe, including a shift from warmer to cooler North Atlantic and North Pacific sea-surface temperatures around 1960 (Folland et al., 1986; Namias et al., 1988; Gordon et al., 1992; Slowey and Crowley, 1995); generally low variability in SOI and the frequency and intensity of El Niño events from the 1920s to the mid-1950s (Elliott and Angell, 1988; Webb and Betancourt, 1992; Whetton et al., 1990); and increased summer monsoon rainfall in India from 1930 to 1964 (Parthasarathy et al., 1987). Other decadal-scale climate variations have also been observed. In the early part of the century, southern California and the lower Colorado River basin in the southwestern United States were wetter than average, and the northwestern part of the country was drier than average during the same time period (McGuirk, 1982; Fritts, 1991). In contrast, the middle of this century was distinguished by low winter precipitation in the lower Colorado River basin (Balling and Lawson, 1982). This well-documented tendency for opposite patterns of precipitation and streamflow in the southwestern and northwestern United States correlates with variations in ENSO and has been linked to persistent large-scale atmospheric circulation states (Bradley et al., 1987a, 1987b; Meko and Stockton, 1984; Cayan and Peterson, 1989; Redmond and Koch, 1991; Lins, 1997).

These prominent decadal-scale climatic fluctuations over the twentieth century affect seasonal precipitation and therefore could have an impact on flooding as well. In the southwestern United States, if floods varied directly with patterns in decadal precipitation and streamflow, it is possible that differences in flood frequency in individual rivers might be detected when comparing climatically different subperiods of the twentieth century. Webb and Betancourt (1992) divided the annual flood series of the Santa Cruz River in southern Arizona into three periods: 1915–29, 1930–59, and 1960–86. Although the mean discharges for the three periods were not significantly different, the variance for 1960–86 was significantly greater than for the preceding periods and the variance for 1930–59 was the lowest of the three. Webb and Betancourt attributed this pattern to the increased magnitude and frequency of floods from winter frontal storms and tropical cyclones and the decreased frequency of summer floods after 1960. In the same region, floods that exceeded the 10-year recurrence interval discharge on 20 rivers in the lower Colorado River basin were tabulated to determine whether any regional-scale variations existed in the temporal distribution of floods (Ely, 1992). No significant change was found in the frequency of winter floods from 1900 to 1988, although the peak discharges of winter floods were generally higher in the periods 1905–41 and 1965–88 than in the intervening decades. Over the period 1920–88, the frequency of floods from tropical storms also remained fairly constant, except for a prominent absence of floods from this storm type from 1940 to 1950. In contrast, the number of summer floods in this region was slightly greater during the middle part of the century, from 1919 to 1955, than after 1955. Although the patterns in the occurrence of 10-year floods are not particularly strong for all three types of precipitation systems, they do indicate that the findings from the Santa Cruz River (Webb and Betancourt, 1992) are valid across a larger hydroclimatic region and they support the expected response to the decadal variations in climate and atmospheric circulation over this century.

Many regional and global oceanic and atmospheric phenomena that influence the potential for floods on an interannual scale, such as ENSO or the North Atlantic oscillation, also hold the potential to influence longer-term temporal patterns in the occurrence of floods over decades to centuries through links with the atmospheric circulation patterns discussed earlier. Examining the likelihood of clusters of floods within multiple-year periods dominated by a particular set of climatic conditions conducive to flooding is more successful than attempting to assess the occurrence of floods on an annual scale. For example, although the effect of individual El Niño events on flood magnitude may vary among El Niño years, winter floods and tropical storms in the lower Colorado River basin show a much more consistent relationship with multiple-year periods dominated by negative SOI values (Ely et al., 1993).

Long-term flooding variations have also been observed in some rivers in India during the twentieth century. Few to no large floods occurred on the three largest rivers in central India during the middle third of the century from about 1926 to 1949 (Figures 2.4 and 2.5). Interestingly, this decrease in extreme floods overlaps with an extended period of above-average monsoon precipitation from the 1930s through the 1950s (Figure 2.5). The number of large floods increased after 1950 and continued into a period of greater variability in both monsoon precipitation and ENSO activity that began around 1960 and that has lasted to the present (Elliott and Angell, 1988; Whetton et al., 1990; Ely et al., 1996). This heightened variability may have enhanced the conditions conducive to the occurrence of extreme floods through changes in atmospheric circulation or other unidentified factors. With the scarcity of long discharge records for other rivers in

India, it is impossible to determine with certainty whether the decadal-scale patterns observed on these three rivers are related to regional climatic factors or are random. In addition, construction of two dams on the lower Krishna River has probably diminished the flood peaks in the latter part of the record. However, evidence from paleoflood records on these rivers suggests that the recent cluster of large floods is unusual in the long-term experience of the rivers. Floods equivalent in magnitude to those in recent decades have not occurred in at least the past 300–400 years on the Narmada River and in the past 650 years on the Godavari River. The frequency of extreme floods since 1960 is unprecedented in the 1700-year paleoflood record on the Narmada River (Ely et al., 1996). The relationship between the recent floods and paleofloods on the Krishna River is unclear because of the presence of dams on the river.

As mentioned earlier, neither abundant seasonal monsoon precipitation nor a positive SOI phase is correlated with the pattern of severe floods in central India. The floods are directly caused by synoptic-scale low-pressure systems and monsoon depressions from the Bay of Bengal and late-season tropical cyclones. A decrease in the frequency of tropical cyclones in the Indian Ocean from 1951 to 1986 (Raper, 1993) suggests that cyclone frequency alone is not sufficient to bring about this type of change in the magnitude and frequency of extreme floods. Much debate has arisen about the possible effects of future global warming on the frequency, intensity, and geographic range of tropical cyclones (Henderson-Sellers et al., 1998). However, evidence from areas such as India and the United States, where many extreme floods are associated with tropical cyclones, indicates that factors other than tropical storm frequency may be more important for predicting the impact of future climate variations on regional flood hazards. A synergism between the tropical cyclones and their pathways with respect to drainage basins is a key factor that leads to severe floods in tropical-storm regions and this relationship can evolve independently of any change in cyclone frequency. However, in some areas, more frequent storm occurrence may increase the likelihood of tropical storms taking pathways that produce major floods. A good example is the increased occurrence of floods from tropical cyclones during El Niño years in the southwestern United States. Typically, the greatest number of eastern North Pacific tropical storms are generated in midsummer and the number tapers off toward fall. However, it has been observed that many of the floods from tropical storms in the Southwest occur during El Niño years in the late summer and fall seasons. One explanation is that the higher overall frequency of tropical storms during El Niño years increases the probability of a few being caught in late-season meridional flow and steered northward.

Lessons from the Paleorecord: Flood Variations on Centennial-to-Millennial Time Scales

The rare, catastrophic floods that characterize the extremes of the hydrological system are the least understood and the most difficult to study by traditional methods of direct observation (Baker, 1987). Paleohydrologic data provide the best means to capture the long-term range of variability exhibited by these rare events over time scales of centuries or greater. A variety of paleohydrologic techniques exist for examining different aspects of the hydrologic system (Costa, 1987; Jarrett, 1991; Wohl and Enzel, 1995; Baker, Chapter 13, this volume). By combining the paleoflood data with independent paleoclimatic evidence of mean precipitation and temperature conditions over the same period, one can begin to examine the relationship between floods and climate change over long time scales (Knox, 1993; Ely, 1997; Enzel and Wells, 1997). This link is best established in regions where only one or two major storm types are associated with the largest floods in the modern record and where these flood-generating storms are consistently associated with a specific large-scale atmospheric circulation pattern or global phenomenon such as ENSO.

A growing body of paleoflood evidence from diverse hydrologic and climatic systems around the world indicates that many rivers have experienced variations in the magnitude or frequency of large floods at centennial to millennial time scales. Comparisons of the temporal and spatial patterns in paleoflood and paleoclimatic data have revealed that the magnitude and frequency of large floods in several regions around the world are associated with Holocene climatic variations (Hassan, 1981; Knox, 1983, 1993; Chatters and Hoover, 1986; Enzel et al., 1989; Smith, 1992; Ely et al., 1993; Rumsby and Macklin, 1994; Benito et al., 1996). It must be stressed that long-term changes in mean temperature or mean precipitation do not directly cause changes in flood frequency; rather, a change in mean conditions signals a change in large-scale circulation patterns, air mass boundaries, or storm trajectories that may have influenced the occurrence of flood-generating conditions (Knox and Kundzewicz, 1997).

A regional chronology of paleoflood records covering the past 5800 years on 19 rivers in the lower Colorado River basin of Arizona and southern Utah shows that clusters of large floods in this region coincide with intervals of cool wet climate, global neoglacial advances, high regional lake levels, and an increased frequency of strong El Niño events (Ely, 1992, 1997; Ely et al., 1993). The frequency of extreme floods has varied in response to these shifting climatic conditions and is clearly nonrandom in time. Large floods were relatively frequent in this region from 3800 to 2200 B.C. and after 400 B.C., with particularly prominent peaks in magnitude and frequency from 3600 to 3200 B.C., from A.D. 900 to 1100, and after A.D. 1400. In sharp contrast, the periods from 2200 to 400 B.C. and A.D. 1200 to 1400 were marked by significant decreases in the number of large floods on virtually all the rivers in the study. The intervals of extreme floods coincided with cool, wet, climatic conditions in the region, whereas the sharp decrease in large floods from A.D. 1200 to 1400 directly coincided with the warm, dry conditions during the Medieval Warm Period. Over at least the past 1000 years a positive relationship has existed between paleoflood frequency and variations in the frequency of strong El Niño events (Anderson, 1992). The same relationships between floods and climatic conditions that are seen at interannual or decadal scales in the Colorado River basin are apparent over these longer time scales. The studies listed above show that extreme floods are not random in time but that the climatic factors that influence the occurrence of extreme floods in the short term also exert control over the response of floods to longer-term climatic variations.

Comparison of paleoflood chronologies from different regions can highlight consistent, predictable, long-term similarities or differences in the

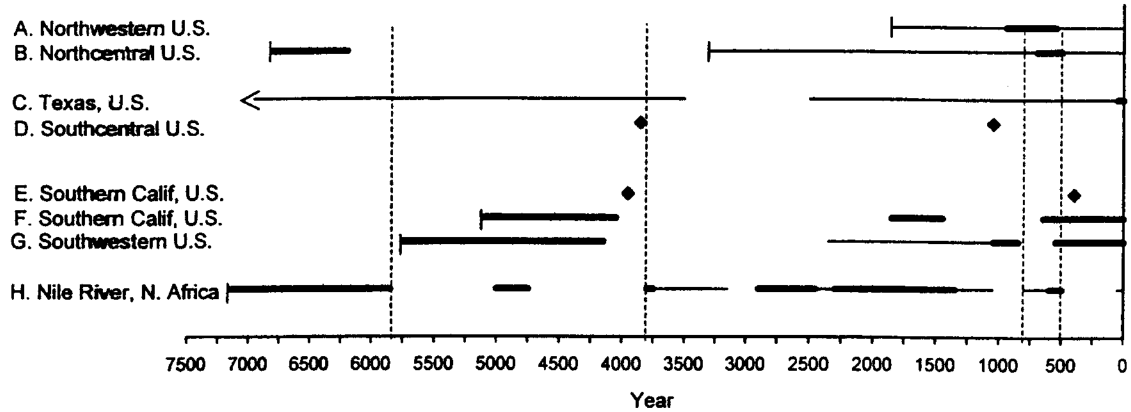


Figure 2.7. Comparison of historical and paleoflood chronologies from Nile River and several regions of North America. The flood record on the Nile River (H) bears a negative correlation with strong El Niño years (Quinn, 1992). Periods with an increased (decreased) frequency of large floods on the Nile River correspond to periods of decreased (increased) floods in the southwestern United States (E–G) and, to a lesser extent, increased (decreased) floods in the northwestern and northcentral United States (A and B). The timing of floods in the southcentral United States (C and D) appears similar to that of the southwestern United States, but the results are inconclusive because of the sample size and resolution of the data set. Solid heavy lines, periods of frequent large floods; solid thin lines, periods of moderate-sized floods; blank spaces, periods of few to no large floods; diamonds, single paleoflood; vertical dashed lines, times of major changes in Nile River record. Years are calibrated to calendar years before 1950 (Stuiver and Reimer, 1993). Data sources: A (Chatters and Hoover, 1986); B (Knox, 1993); C (Kochel et al., 1982; Patton and Dibble, 1982); D (McQueen et al., 1993); E (Enzel and Wells, 1997); F (McGill and Rockwell, 1998); G (Ely, 1997); H (Hassan, 1981; Said, 1993).

occurrence of floods among distinct hydroclimatic regions. As discussed earlier, individual floods can result from multiple causes that may or may not be related to the same set of persistent, large-scale atmospheric or oceanic conditions. Therefore, in attempting to identify climate-related patterns in paleoflood records from different areas, analysis must begin with regions that show very straightforward connections to present climatic phenomena, such as ENSO, or that exhibit strong positive or negative correlations with seasonal streamflow or precipitation in another region. The paleoflood record from the lower Colorado River basin provides examples of several different interregional hydroclimatic links (Figure 2.7). Discharge records in the lower Colorado River basin and the Nile River demonstrate an opposite response to El Niño years and this is reflected in centennial-scale variations in flood characteristics over the past 1000 years. Periods of exceptionally low Nile River floods from 3900 to 1840 B.C. [5850 to 3790 before present (BP)], A.D. 930 to 1090 (1020 to 860 BP), and in the past 300 years (Hassan, 1981; Said, 1993) correspond with prominent peaks in flood magnitude and frequency in the lower Colorado River basin (Ely et al., 1993), whereas the period with very few years of low floods from A.D. 1090 to 1470 (860 to 480 BP) overlaps with the sharp decline in large floods in the lower Colorado River basin during the Medieval Warm Period.

In another example, one of the dominant spatial patterns of hydroclimatic variation in North America is the out-of-phase relationship in precipitation and streamflow between the northwestern and southwestern parts of the United States (Meko and Stockton, 1984; Redmond and Koch, 1991). This pattern is partially attributed to the strong opposite response of these two regions to variations in the intensity of large-scale atmospheric and oceanic phenomena in the Pacific, particularly ENSO and the Pacific North American pattern (Cayan and Peterson, 1989). The highest frequency of large floods in a 1900-year record from the Columbia River in the northwestern United States was from A.D. 1020 to 1390 (Chatters and Hoover, 1986), which is the interval of least frequent floods in the lower Colorado River basin over the same period (Figure 2.6). A similar asynchronous pattern occurs between these southwestern United States paleofloods and a 7000-year flood record from the upper Mississippi River basin in the north-central United States (Knox, 1993; Knox and Kundzewicz, 1997). In southwestern Texas, on the other hand, frequent, moderate-sized floods during a humid climate from ~3400 to 2600 calendar years BP coincided with the virtual lack of large floods on the southwestern rivers, whereas the drier periods before and after this time experienced sporadic, but larger floods (Patton and Dibble, 1982). Thus, the clusters of the largest floods coincide in time, although the climatic conditions associated with the floods differ. The relationships between all these areas suggest that persistent large-scale atmospheric circulation patterns that increase the potential flood hazards in the southwestern and possibly south-central United States have decreased the potential for floods in the northwestern and north-central parts of the country over the same time periods (Ely, 1992; Knox and Kundzewicz, 1997).

In several studies, clusters of particularly frequent severe floods have been attributed to enhanced variability during periods of climatic transition, such as the change from warmer to cooler climate in many parts of North America and Europe sometime between A.D. 1250 and 1500 (Ely, 1992; Knox, 1993; Benito et al., 1996). The frequency of floods in the upper Mississippi River valley in the north-central United States increased between A.D. 1250 and 1400 (Knox, 1993), whereas floods in the southwestern United States and Spain increased after A.D. 1400 to 1500 (Ely, 1992; Benito et al., 1996). Although the periods of increased flood frequency are out of phase between the north-central United States and the other regions, they still may be responding to different aspects of this transitional period, either the end of the warm period or the beginning of the cool period. This sort of intriguing pattern can be examined and verified as the database of long-term records grows across the globe.

Contradictions in the expected patterns of floods and climate change between regions introduce an interesting point. Obvious hydroclimatic links among separate paleoflood records may follow modern examples during certain periods in the past, whereas at other times floods in the same or different regions appear to respond to climatic conditions in different ways. For example, several periods of high and low floods occurred on the Nile River (Said, 1993) during the long interval with no depositional record of large floods in the southwestern United States from about 4100 to 2300 years ago (Figure 2.6). Very long paleoflood records from relict flood levees at tropical latitudes in Australia suggest that the magnitude and frequency of extreme floods could increase under two dramatically different climate conditions: the much cooler, wetter conditions prior to the last glacial maximum and the warmer, wetter conditions of the mid-Holocene (Nott et al., 1996). The challenge therefore is to develop enough data sets in a variety of climatic regions around the globe that these periods of coincidence and divergence in the hydrologic responses to climate begin to emerge. In this way, the flood records can actually begin to point toward specific shifts in large-scale circulation patterns, airmass boundaries, and storm tracks that might not be apparent from changes in mean climatic conditions, providing insight into how the global climate system has operated at different periods in the past and how hydrologic systems in different regions could respond to future climate changes.

Concluding Remarks

Understanding the hydroclimatic controls on variations in the magnitude and frequency of extreme floods is critical for flood-frequency forecasting.

As shown in the examples in this chapter, the spatial and temporal distribution of large floods in a variety of regions around the world are influenced to a large extent by persistent, anomalous patterns in hemispheric to global-scale atmospheric and oceanic circulation that have an influence on flood-generating precipitation systems. However, it is these smaller-scale precipitation systems, operating at relatively short time scales, that directly cause floods. Moreover, in many cases, floods tend to develop from a unique set of ingredients that involve a synergism between meteorologic, climatic, hydrologic, and drainage basin factors. To understand the causes of floods fully, we attest that both short-term meteorologic processes and longer-term climatic processes must be addressed with a flood hydroclimatology perspective.

Extreme floods are by nature rare, and the gaged record from a single river provides insufficient data to characterize the frequency of these events accurately or to evaluate the maximum flood potential in the context of the long-term history of the river. Geologic evidence of paleofloods from the past several centuries to millennia reveals patterns in the magnitude and frequency of the largest floods that are not apparent from the relatively short stream-gage records in the western United States. Examining the linkages between climate and floods across various temporal and spatial scales will improve our ability to anticipate the local and regional flood hazards that can be expected under a range of climatic conditions, an important component of long-term planning in the face of uncertain future climate scenarios. The understanding gained through establishing these connections provides practical input for floodplain development and management; evaluation of the safety, design, and economic feasibility of dams and other structures; and the understanding and management of riparian ecosystems.

One main conclusion that can be drawn from our overview of flood-causing meteorologic and antecedent climatologic processes is that the manner in which above-average precipitation is delivered in space and time affects both the type of flood that occurs and its accompanying hazards. Hence, the scale of a precipitation system is an important component of flood causality. However, precipitation systems that occur at one scale are strongly interconnected with systems at other scales, and larger-scale processes tend to set the stage for activity at smaller scales. Across all scales of activity, persistence – or the slow-moving nature of a precipitation system – appears to be a key element for producing exceptionally large floods.

Over the long term, differences in paleoflood patterns within and between regions have enormous implications for flood-frequency forecasting and river management. Long-term variations in climate create changes in the overall flooding regime and cannot be used to predict the occurrence of individual floods in any given year or even in a span of multiple years. Understanding the extent of the impacts of past climatic variations on flooding can aid in assessing the potential economic or societal implications of future change. For example, past changes in climate and flood regimes may have contributed to the demise of ancient societies such as the Anasazi and Hohokam of the southwestern United States through an increased frequency of severe floods that caused deep incision of alluvial channels and/or destruction of floodplain irrigation systems (Huckleberry, 1993). Could this type of change have a substantial impact on modern societies? Our river management is designed around expectations of a predictable range of flood hazards centered largely on statistical extrapolations based on the historical record of a river over a period of decades. A change in the long-term flood regime may render aspects of this infrastructure no longer applicable, leading to economic impacts at the least, and possibly extensive loss of life and property. Stationarity in a flood series means that the probability distribution of any random variable, such as the mean discharge or variability, is constant through time. Nonstationarity, or nonrandom changes in the flood distribution, create significant problems for traditional statistical methods of flood frequency analysis. Examination of the past relationships between floods and climate places the current flood hydroclimatic conditions into perspective and reveals any evidence of nonstationarity in the record. For example, is a region currently undergoing a period of frequent or infrequent flooding compared with the long-term paleorecords? How would the addition of the paleoflood information affect statistical flood-frequency analyses and risk assessments based solely on the modern flood record? On some rivers, paleoflood information has revealed pronounced nonstationarity in the flood series over long time periods (Baker, 1987). Since the inception of the gaged record in 1948, the Narmada River in India has experienced several high-magnitude floods (Figure 2.5). Both the magnitude and frequency of these floods stand out as an anomaly compared with the 1700-year paleoflood record (Ely et al., 1996). The design of the Narmada Sagar Dam currently under construction was based on the gaged record, which by chance in this case incorporates the largest floods in the long-term history of the river and is probably an adequate representation of the maximum floods that are likely to occur in the future. On rivers where past floods were larger or more frequent than present, the gaged records could grossly underestimate the floods that could be reasonably expected with a moderate shift in future climatic conditions. Knowledge of the paleoflood history could avert the potential disaster of underdesigning major structures on the river.

The flood–climate relationship on a river is also an effective means of evaluating the feasibility of the theoretical models of worst-case scenario floods commonly used for design purposes. On the Verde and Salt Rivers in central Arizona the largest paleofloods in at least the past 1000 years were slightly less than twice the magnitude of the largest floods in the gaged records before 1993 (Ely and Baker, 1985; Partridge and Baker, 1987). The theoretical probable maximum flood, calculated by the Bureau of Reclamation for dam design on these rivers, was four to five times greater than the maximum paleoflood on each river (Baker et al., 1987). Armed with this knowledge, one can better evaluate the utility of the enormous public expenditure in designing for a theoretical flood that is significantly greater than the worst floods that have occurred under the range of climatic conditions over the last millennium. In other cases, such as the Colorado River itself, the probable maximum flood was only slightly larger than the largest paleoflood (O'Connor et al., 1994).

A possible solution to these problems of choosing the appropriate design flood for structures or floodplain development is to identify past climatic periods characterized by the highest flood frequencies or magnitudes and use them as a basis for future flood hazard assessment in a region. Incorporating the range of past natural hydroclimatic variability into a long-range flood-frequency forecast model is critical for an accurate assessment of the range of effects future climatic change might have on flood hazards for that specific region. The emerging patterns of similarities and differences among flood hydroclimatic regions globally can aid tremendously in augmenting evidence from any single region in evaluating the potential flood hazards under scenarios of changing climate, from annual to millennial scales.

Acknowledgments

We wish to thank our reviewers, E.E. Wohl, J.M. Friedman, and M.W. Douglas for their insightful comments and suggestions. This work was funded by NASA Earth Science Enterprise Grant NAGW-3498 and NOAA Paleoclimatology Grant GC-94-504 (Hirschboeck), and NSF Grants EAR-9202498 and EAR-9104489 and the Faculty Research Fund at Central Washington University (Ely).

References

Anderson, R.Y. (1992). Long-term changes in the frequency of occurrence of El Niño events. In *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, ed. H.F. Diaz and V. Markgraf, pp. 193–200. Cambridge: Cambridge University Press.

Anderson, C.J. and Arritt, R.W. (1998). Mesoscale convective complexes and persistent elongated convective systems over the United States during 1992 and 1993. *Monthly Weather Review*, **126**, 578–599.

Andrade, E. and Sellers, W. (1988). El Niño and its effect of precipitation in Arizona. *Journal of Climatology*, **8**, 403–410.

Bailey, J.F., Patterson, J.L., and Paulhus, J.L.H. (1975). Hurricane Agnes rainfall and floods, June–July 1972. U.S.G.S. Professional Paper 924. Washington, DC: U.S. Government Printing Office.

Baker, V.R. (1987). Paleoflood hydrology and extraordinary flood events. *Journal of Hydrology*, **96**, 79–99.

Baker, V.R., Ely, L.L., O'Connor, J.E., and Partridge, J.B. (1987). Paleoflood hydrology and design applications. In *Regional Flood Frequency Analysis*, ed. V. Singh, pp. 325–338. Boston: D. Reidel.

Balling, R. and Lawson, M. (1982). Twentieth century changes in winter climatic regions. *Climatic Change*, **4**, 57–69.

Bandyopadhyay, J., Rodda, J.C., Kattelmann, R., Kundzewicz, and Kraemer, D. (1997). Highland waters – a resource of global significance. In *Mountains of the World – a Global Priority*, eds. B. Messerli and J.D. Ives, pp. 131–155. London: The Parthenon Publishing Group.

Barry, R.G., Kiladis, G., and Bradley, R.S. (1981). Synoptic climatology of the western United States in relation to climatic fluctuations during the twentieth century. *Journal of Climatology*, **1**, 97–113.

Benito, G., Machado, M.J., and Pérez-González, A. (1996). Climate change and flood sensitivity in Spain. In *Global Continental Changes: The Context of Palaeohydrology*, ed. J. Branson, A.G. Brown, and K.J. Gregory, pp. 85–98. Geological Society of London Special Publication No. 115.

Bhalme, H.N. and Mooley, D.A. (1980). Large-scale droughts/floods and monsoon circulation. *Monthly Weather Review*, **108**, 1197–1211.

Bradley, R., Diaz, H., Eischeid, J., Jones, P., Kelly, P., and Goodess, C. (1987a). Precipitation fluctuations over northern hemisphere land areas since the mid-19th century. *Science*, **237**, 171–75.

Bradley, R., Diaz, H., Kiladis, G., and Eischeid, J. (1987b). ENSO Signal in continental temperature and precipitation records. *Nature*, **327**, 497–501.

Burstein, J. (1997). "1 dead, 10 missing in canyon flood: Guide found at bottom is only survivor," Article in *The Arizona Daily Star*, Tucson, Arizona, Thur. 14 Aug. 1997.

Carcena, F. and Fritsch, J.M. (1983). Forcing mechanisms in the Texas Hill Country flash floods of 1978. *Monthly Weather Review*, **111**, 2319–2332.

Cayan, D. and Peterson, D. (1989). The influence of North Pacific atmospheric circulation on streamflow in the West. In *Aspects of Climate Variability in the Pacific and the Western Americas*, ed. D.H. Peterson, pp. 375–397. Washington, DC: American Geophysical Union Geophysical Monograph 55. Washington, DC.

Cayan, D. and Webb, R.H. (1992). El Niño/Southern Oscillation and streamflow in the western United States.

In *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, ed. H. Diaz and V. Markgraf, pp. 29–68. Cambridge: Cambridge University Press.

Chappell, C.F. (1986). Quasi-stationary convective events. In *Mesoscale Meteorology and Forecasting*, ed. P. Ray, pp. 289–310. Boston: American Meteorological Society.

Chatters, J.C. and Hoover, K.A. (1986). Changing Late Holocene flooding frequencies on the Columbia River, Washington. *Quaternary Research*, **26**, 309–320.

Chen, G.T.-J. and Yu, C.-C. (1988). Study of low-level jet and extremely heavy rainfall over northern Taiwan in the Mei-Yu season. *Monthly Weather Review*, **116**, 884–891.

Costa, J.E. (1987). A history of paleoflood hydrology in the United States, 1800–1970. In *History of Hydrology*, eds. E.R. Landa and S. Ince, pp. 49–67. History of Geophysics, vol. 3. Washington, DC: American Geophysical Union.

Das, P.K. (1987). Short- and long-range monsoon prediction in India. In *Monsoons*, eds. J.S. Fein and P.L. Stephens, pp. 549–578. New York: John Wiley and Sons.

Dhar, O.N., Mandal, B.N., and Mulye, S.S. (1985). Some aspects of rainfall distribution over Narmada basin up to Sardar Sarovar Damsite in Gujarat. Proceedings from National Seminar-cum-Workshop on Atmospheric Sciences and Engineering, Jadavpur University, 186–193.

Dhar, O.N. and Nandargi, S. (1993). The zones of severe rainstorm activity over India. *International Journal of Climatology*, **13**, 301–311.

Diaz, H. and Quayle, R. (1980). The climate of the United States since 1895, spatial and temporal changes. *Monthly Weather Review*, **108**, 249–266.

Doswell, C.A. III, Brooks, H.E., and Maddox, R.A. (1996). Flash flood forecasting: an ingredients-based methodology. *Weather and Forecasting*, **11**, 560–581.

Dunne, T. (1983). Relation of field studies and modeling in the prediction of storm runoff. *Journal of Hydrology*, **65**, 25–48.

Dzerdzeevski, B. (1969). Climate epochs in the Twentieth Century and some comments on the analysis of past climates. In *Quaternary Geology and Climate, Pub. No. 1701*, ed. H. Wright, pp. 49–60. Washington, DC: National Academy of Sciences.

Elliott, W. and Angell, J. (1988). Evidence for changes in Southern Oscillation relationships during the last 100 years. *Journal of Climate*, **1**, 729–737.

Ely, L.L. (1992). *Large Floods in the Southwestern United States in Relation to Late- Holocene Climatic Variations*. Ph.D. dissertation, University of Arizona, Tucson. 326 pp.

Ely, L.L. (1997). Response of extreme floods in the southwestern United States to climatic variations in the Late Holocene. *Geomorphology*, **19**, 175–201.

Ely, L.L. and Baker, V.R. (1985). Reconstructing paleoflood hydrology with slackwater deposits: Verde River, Arizona. *Physical Geography*, **6**, 103–126.

Ely, L.L., Enzel, Y., Baker, V.R., and Cayan, D.R. (1993). A 5000-year record of extreme floods and climate change in the southwestern United States. *Science*, **262**, 410–412.

Ely, L.L., Enzel, Y., Baker, V.R., Kale, V.S., and Mishra, S. (1996). Changes in the magnitude and frequency of Late Holocene monsoon floods on the Narmada River, Central India. *Geological Society of America Bulletin*, **108**, 1134–1148.

Ely, L.L., Enzel, Y., and Cayan, D.R. (1994). Anomalous North Pacific atmospheric circulation and large winter floods in the southwestern United States. *Journal of Climate*, **7**, 977–987.

Enzel, Y., Cayan, D.R., Anderson, R.Y., and Wells, S.G. (1989). Atmospheric circulation during Holocene lake stands in the Mojave desert: evidence of regional climate change. *Nature*, **341**, 44–48.

Enzel, Y. and Wells, S.G. (1997). Extracting Holocene paleohydrology and paleoclimatology information from modern extreme flood events: an example from Southern California. *Geomorphology*, **19**, 203–226.

Folland, C.K., Palmer, T.N., and Parker, D. (1986). Sahel rainfall and worldwide sea temperatures, 1901–1985. *Nature*, **320**, 602–607.

Fritsch, J.M., Kane, R.J., and Chelius, C.H. (1986). The contribution of mesoscale convective weather systems to the warm season precipitation in the United States. *Journal of Climate and Applied Meteorology*, **25**, 1333–1345.

Fritts, H.C. (1991). *Reconstructing Large-Scale Climatic Patterns from Tree-Ring Data*. Tucson: University of Arizona Press.

Glancy, P. and Harmsen, L. (1975). A hydrologic assessment of the September 14, 1974 flood in Eldorado Canyon, Nevada. *USGS Professional Paper 930*. Washington, DC: U.S. Government Printing Office.

Goodrich, D.C., Schmugge, T.J., Jackson, T.J., Unkrich, C.L., Keefer, T.O., Parry, R., Bach, L.B., and Amer, S.A. (1994). Runoff simulation sensitivity to remotely-sensed initial soil water content. *Water Resources Research* **30**, 1393–1405.

Gordon, H., Whetton, P., Pittock, A., Fowler, A., and Haylock, M. (1992). Simulated changes in daily rainfall intensity due to the enhanced greenhouse effect: implications for extreme rainfall events. *Climate Dynamics*, **8**, 83–102.

Gray, W.M. (1968). Global view of the origin of tropical disturbances and storms. *Monthly Weather Review*, **96**, 669–700.

Gregory, S. (1989). Macro-regional definition and characteristics of Indian summer monsoon rainfall, 1871–1985. *International Journal of Climatology*, **9**, 465–483.

Hane, C.E. (1986). Extratropical squall lines and rainbands. In *Mesoscale Meteorology and Forecasting*, ed. P. Ray, pp. 359–89. Boston: American Meteorological Society.

Harr, R.D. (1981). Some characteristics and consequences of snowmelt during rainfall in western Oregon. *Journal of Hydrology*, **53**, 277–304.

Hassan, F.A. (1981). Historical Nile floods and their implications for climatic change. *Science*, **212**, 1142–1145.

Hayden, B.P. (1988). Flood climates, In *Flood Geomorphology*, eds. V.R. Baker, R.C. Kochel, and P.C. Patton, pp. 13–26. New York: John Wiley and Sons.

Henderson-Sellers, A., Zhang, H., Berz, G., Emanuel, K., Gray, W., Landsea, C., Holland, G., Lighthill, J., Shieh, S., Webster, P., and McGuffie, K. (1998). Tropical cyclones and global climate change: a post-IPCC assessment. *Bulletin of the American Meteorological Society*, **79**, 19–38.

Hirschboeck, K.K. (1985). *Hydroclimatology of Flow Events in the Gila River Basin, Central and Southern Arizona*. Ph.D. dissertation, University of Arizona, Tucson. 335 pp.

Hirschboeck, K.K. (1987a). Catastrophic flooding and atmospheric circulation anomalies. In *Catastrophic Flooding*, eds. L. Mayer and D.B. Nash, pp. 23–56. Boston: Allen and Unwin.

Hirschboeck, K.K. (1987b). Hydroclimatically-defined mixed distributions in partial duration flood series. In *Hydrologic Frequency Modeling*, ed. V.P. Singh, pp. 199–212. Dordrechet: D. Reidel Publishing Company.

Hirschboeck, K.K. (1988). Flood hydroclimatology. In *Flood Geomorphology*, eds. V.R. Baker, R.C. Kochel, and P.C. Patton, pp. 27–49. New York: John Wiley and Sons.

Hirschboeck, K.K. (1991). Climate and floods. In *National Water Summary 1988–89 – Floods and Droughts: Hydrologic Perspectives on Water Issues*, pp. 67–88. U.S. Geological Survey Water-Supply Paper 2375.

Hirschboeck, K.K. (1996). Floods. In *Encyclopedia of Climate and Weather*, ed. S.H. Schneider, pp. 308–311. Oxford, UK: Oxford University Press.

Hobbs, P.V. (1978). Organization and structure of clouds and precipitation on the mesoscale and microscale in cyclonic storms. *Reviews of Geophysics and Space Physics*, **16**, 741–755.

Horel, J. and Wallace, J. (1981). Planetary-scale atmospheric phenomena associated with the Southern Oscillation. *Monthly Weather Review*, **109**, 813–829.

Houghton, J.T., Meira Filho, L.G., Callander, B.A., Harris, N., Kattenberg, A., and Maskell, K. (eds.) (1996). *Climate Change 1995: The Science of Climate Change IPCC Report*. New York: Cambridge University Press, 572 pp.

House, P.K. and Hirschboeck, K.K. (1997). Hydroclimatological and paleohydrological context of extreme winter flooding in Arizona, 1993. In *Storm-Induced Geological Hazards: Case Histories from the 1992–1993 Winter Storm in Southern California and Arizona*, eds. R.A. Larson and J.E. Slosson, pp. 1–24. Boulder, Colorado: Geological Society of America Reviews in Engineering Geology, v. XI.

Houze, R.A. Jr. (1981). Structures of atmospheric precipitation systems: a global survey. *Radio Science*, **16**, 671–689.

Hoxit, L.R., Maddox, R.A., Chappell, C.F., and Brua, S.A. (1982). Johnstown – Western Pennsylvania storm and floods of July 19–29, 1977. U.S. Geological Survey Professional Paper 1211. Washington, DC: U.S. Government Printing Office.

Huckleberry, G.A. (1993). *Late-Holocene Stream Dynamics on the Middle Gila River, Pinal County, Arizona*. Ph.D. dissertation, University of Arizona, Tucson. 135 pp.

Jarrett, R.D. (1991). Paleohydrology and its value in analyzing floods and droughts. In *National Water Summary 1988–89 – Floods and Droughts: Hydrologic Perspectives on Water Issues*, pp. 105–16. U.S. Geological Survey Water-Supply Paper 2375. Washington, DC.

Junker, N.W., Schneider, R.S., and Scofield, R.A. (1995). The meteorological conditions associated with the Great Midwest Flood of 1993. In *Preprints, 14th Conf. On Weather Analysis and Forecasting*, pp. (J4) 13–17. Washington DC: American Meteorological Society.

Kale, V.S., Mishra, S., and Baker, V.R. (1997). A 2000-year palaeoflood record from Sakarghat on Narmada, Central India. *Journal Geological Society of India*, **50**, 283–288.

Kalnicky, R. (1974). Climatic change since 1950. *Annals of the Association of American Geographers*, **64**, 100–112.

Knox, J.C. (1983). Responses of river systems to Holocene climates. In *Late Quaternary Environments of the United States*. Vol. 2, *The Holocene*, ed. H. Wright, pp. 26–41. Minneapolis: University of Minnesota Press.

Knox, J.C. (1993). Large increases in flood magnitude in response to modest changes in climate. *Nature*, **361**, 430–432.

Knox, J.C. and Kundzewicz, Z.W. (1997). Extreme hydrological events, palaeo-information and climate change. *Hydrological Sciences Journal*, **42**, 765–779.

Laing, A.G. and Fritsch, M.J. (1997). The global population of mesoscale convective complexes. *Quarterly Journal of the Royal Meteorological Society*, **123**, 389–405.

Lins, H.F. (1997). Regional streamflow regimes and hydroclimatology of the United States. *Water Resources Research*, **33**, 1655–1667.

Maddox, R.A., Canova, F., and Hoxit, L.R. (1980). Meteorological characteristics of flash flood events over the western United States. *Monthly Weather Review*, **108**, 1866–1877.

Maddox, R.A., Chappell, C.F., and Hoxit, L.R. (1979). Synoptic and meso-α scale aspects of flash flood events. *Bulletin of the American Meteorological Society*, **60**, 115–123.

Maddox, R.A., Hoxit, L.R., Chappell, C.F., and Caracena, F. (1978). Comparison of meteorological aspects of the Big Thompson and Rapid City flash floods. *Monthly Weather Review*, **106**, 375–389.

Maddox, R.A., Howard, K.A., Bartels, D.L., and Rodgers, D.M. (1986). Mesoscale convective complexes in the middle latitudes. In *Mesoscale Meteorology and Forecasting*, ed. P. Ray, pp. 390–413. Boston: American Meteorological Society.

Matsumoto, S., Ninomiya, K., and Yoshizumi, S. (1971). Characteristic features of the Baiu front associated with heavy rainfall. *Journal of the Meteorological Society of Japan*, **49**, 267–281.

Matthai, H.F. (1969). *Floods of June 1965 in South Platte River Basin, Colorado*. U.S. Geological Survey Water Supply Paper 1850-B. Washington, DC: U.S. Government Printing Office.

McGill, S. and Rockwell, T. (1998). Ages of Late Holocene earthquakes on the Central Garlock Fault near El Paso Peaks, California. *Journal of Geophysical Research*, **103(B4)**, 7265–7279.

McGuirk, J. (1982). A century of precipitation variability along the Pacific Coast of North America and its impact. *Climatic Change*, **4**, 41–56.

McQueen, K., Vitek, J., and Carter, B. (1993). Paleoflood analysis of an alluvial channel in the south-central Great Plains: Black Bear Creek, Oklahoma. *Geomorphology*, **8**, 131–146.

Means, L.L. (1954). A study of the mean southerly wind-maximum in low levels associated with a period of summer precipitation in the middle west. *Bulletin of the American Meteorological Society*, **35**, 166–170.

Meko, D.M. and Stockton, C.W. (1984). Secular variations in streamflow in the Western United States. *Journal of Climate and Applied Meteorology*, **23**, 889–897.

Michaud, J. and Sorooshian, S. (1994). Effect of rainfall sampling errors on simulations of desert flash floods, *Water Resources Research*, **30**, 2765–2775.

Mooley, D. and Shukla, J. (1987). *Characteristics of the Westward-Moving Summer Monsoon Low Pressure Systems Over the Indian Region and Their Relationship with the Monsoon Rainfall*. College Park, Maryland: Center for Ocean-Land-Atmosphere Interactions, University of Maryland.

Namias, J., Yuan, X.C., and Daniel, R. (1988). Persistence of North Pacific sea surface temperature and atmospheric flow patterns. *Journal of Climate*, **1**, 682–703.

Nott, J.F., Price, D.M., and Bryant, E.A. (1996). A 30,000 year record of extreme floods in tropical Australia from relict plunge-pool deposits: implications for future climate. *Geophysical Research Letters*, **23**, 379–382.

O'Connor, J.E., Ely, L.L., Wohl, E.E., Stevens, L.E., Melis, T.S., Kale, V.S., and Baker, V.R. (1994). 4000-year record of large floods on the Colorado River in the Grand Canyon. *Journal of Geology*, **102**, 1–9.

Orlanski, I. (1975). A rational subdivision of scales for atmospheric processes. *Bulletin of the American Meteorological Society*, **56**, 527–530.

Parthasarathy, B., Kumar, K.R., and Munot, A.A. (1993). Homogeneous Indian monsoon rainfall: variability and prediction. *Proceedings of the Indian Academy of Sciences*, **102** (March), 121–155.

Parthasarathy, B. and Pant, G.B. (1984). The spatial and temporal relationships between the Indian summer monsoon rainfall and the Southern Oscillation. *Tellus*, **36A**, 269–277.

Parthasarathy, B., Sontakke, N.A., Monot, A.A., and Kothawale, D.R. (1987). Droughts/floods in the summer monsoon season over different meteorological subdivisions of India for the period 1871–1984. *Journal of Climatology*, **7**, 57–70.

Partridge, J.B. and Baker, V.R. (1987). Paleoflood hydrology of the Salt River, Arizona. *Earth Surface Processes and Landforms*, **12**, 109–125.

Patton, P.C. and Dibble, D. (1982). Archeologic and geomorphic evidence for the paleohydrologic record of the Pecos River in west Texas. *American Journal of Science*, **282**, 97–121.

Pielke, R.A. (1990). *The Hurricane*. New York: Routledge.

Quinn, W.H. (1992). A study of SO-related climatic activity for A.D. 622-1990 incorporating Nile River flood data. In *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, eds. H.F. Diaz and V. Markgraf, pp. 119–149. Cambridge: Cambridge University Press.

Ramaswamy, C. (1987). *Meteorological Aspects of Severe Floods in India, 1923–1979*. Meteorological Monograph Hydrology, No. 10. New Delhi: Indian Meteorological Department.

Raper, S.C. (1993). Observational data on the relationships between climatic change and the frequency and magnitude of severe tropical storms. In *Climate and Sea-Level Change: Observations, Projections,* *and Implications*, eds. R.A. Warrick, E.M. Barrow, and T.M.L. Wigley, pp. 192–212. New York: Cambridge University Press.

Redmond, K. and Koch, R. (1991). ENSO vs. surface climate variability in the western United States. *Water Resources Research*, **27**, 2381–2399.

Rodier, J. and Roche, M. (1984). *World Catalog of Maximum Observed Floods*. Wallingford, UK: International Association of Hydrological Sciences Publication 143. 354 pp.

Ropelewski, C. and Halpert, M. (1986). North American precipitation and temperature patterns associated with El Niño-Southern Oscillation (ENSO). *Monthly Weather Review*, **114**, 2352–2362.

Ropelewski, C. and Halpert, M. (1987). Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. *Monthly Weather Review*, **115**, 1606–1626.

Rumsby, B. and Macklin, M.G. (1994). Channel and floodplain response to recent abrupt climate change: the Tyne Basin, Northern England. *Earth Surface Processes and Landforms*, **19**, 499–515.

Said, R. (1993). *The River Nile: Geology, Hydrology and Utilization*. Oxford: Pergamon Press.

Schonher, T. and Nicholson, S. (1989). The relationship between California rainfall and ENSO events. *Journal of Climate*, **2**, 1258–1269.

Schroeder, T.A. (1977). Meteorological analysis of an Oahu flood. *Monthly Weather Review*, **105**, 458–468.

Singh, N., Soman, M.K., and Krishna Kumar, K. (1988). Hydroclimatic fluctuations of the upper Narmada catchment and its association with break-monsoon days over India. *Proceedings of the Indian Academy of Sciences (Earth and Planetary Sciences)*, **97**, 87–105.

Slowey, N.C. and Crowley, T.J. (1995). Interdecadal variability of Northern Hemisphere circulation recorded by Gulf of Mexico corals. *Geophysical Research Letters*, **22**, 2345–2348.

Smith, A. (1992). Holocene palaeoclimatic trends from palaeoflood analysis. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **97**, 235–240.

Smith, J.A., Baeck, M.L., and Steiner, M. (1996). Catastrophic rainfall from an upslope thunderstorm in the central Appalachians: the Rapidan storm. *Water Resources Research*, **32**, 3099–3113.

Stuiver, M. and Reimer, P. (1993). Extended ^{14}C data base and revised CALIB 3.0 age calibration program. *Radiocarbon*, **35**, 215–230.

UNESCO (1969–79). *Discharge of Selected Rivers in the World: A Contribution to the International Hydrological Decade*, v. 1–3. Paris: Unesco Press.

UNESCO (1976). *World Catalogue of Very Large Floods*. Paris: Unesco Press.

U.S. Water Resources Council (1981). *Guidelines for Determining Flood Flow Frequency*. Bulletin 17B, Washington, DC: U.S. Water Resources Council.

Ward, R. (1978). Floods, *A Geographical Perspective*. New York: John Wiley. and Sons

Waylen, P.R. and Caviedes, C.N. (1986). El Niño and annual floods on the North Peruvian littoral. *Journal of Hydrology*, **89**, 141–156.

Webb, R.H. and Betancourt, J.L. (1992). *Climatic Variability and Flood Frequency of the Santa Cruz River, Pima County, Arizona*. U.S. Geological Survey Water-Supply Paper 2379. Washington, DC: U.S. Government Printing Office.

Webster, P.J. (1987). The variable and interactive monsoon. In *Monsoons*, eds. J.S. Fein and P.L. Stephens, pp. 269–330. New York: John Wiley and Sons.

Weisman, M.L. and Klemp, J.B. (1986). Characteristics of isolated convective storms. In *Mesoscale Meteorology and Forecasting*, ed. P. Ray, pp. 331–358. Boston: American Meteorological Society.

Wells, L.E. (1990). Holocene history of the El Niño phenomenon as recorded in flood sediments of northern coastal Peru. *Geology*, **18**, 1134–1137.

Whetton, P., Adamson, D., and Williams, M. (1990). Rainfall and river flow variability in Africa, Australia and East Asia linked to El Niño – Southern Oscillation events. *Geological Society of America Symposium Proceedings*, 1, 71-82*ENSO-Related Floods and Droughts*, vol. 1.

Wohl, E.E. and Enzel, Y. (1995). Data for Palaeohydrology. In *Global Continental Palaeohydrology*, eds. K.J. Gregory and L. Starkel, pp. 23–59. New York: John Wiley and Sons.