

*Review Article***Riverine Flood Hazard: Part A. Types, Processes and Causative Factors**VIKRANT JAIN^{1,*}, MUSSIE BEYENE², L. SARDINE VARAY³, ROBERT JAMES WASSON⁴ and SHALEEN JAIN⁵¹*Discipline of Earth Sciences, Indian Institute of Technology Gandhinagar, Palaj Gandhinagar, Gujarat 382 355, India*²*Department of Civil and Environmental Engineering, University of Maine, Orono, Maine 04469-5711, USA*³*Department of Geology, Centre of Advanced Studies, University of Delhi, Delhi 110 007, India*⁴*Institute of Water Policy, Lee Kuan Yew School of Public Policy, 469A Bukit Timah Road, Singapore 259770*⁵*Department of Civil and Environmental Engineering, University of Maine, Orono, Maine 04469-5711, USA*

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Floods are one of the most severe natural hazards. Natural and anthropogenic climate change and a changing landscape have the potential to exacerbate flood impacts. An in-depth understanding of flood types, processes and causative factors is essential in planning effective flood management strategies. In-depth understanding of flood events needs analyses of their hydrological as well as geomorphic characteristics. The understanding of palaeofloods, often in conjunction with traditional analyses of flood frequency and magnitude, can also unravel changes related to flood hazard. The new flood management strategies need inclusion of non-structural measures on the basis of an understanding of basin to reach scale processes for sustainable management of flood hazard; however, to be effective it should be based on an understanding of flood dynamics. This study provides a review of the various aspects of flood hazard, with particular focus on the dynamical processes that underpin the flood phenomenon.

Keywords: Return Period; Flood Discharge; Climate Change; Sediment Transport; Flood Power; Palaeoflood; Flood Management; Statistical Hydrology

Introduction

Floods are one of the major natural hazards with manifold societal implications. Recurrent floods span nearly every part of the Indian sub-continent. In terms of frequency and severity, India is amongst the worst flood-affected countries in the world, only second to Bangladesh (Agarwal and Narain, 1996; NRSC, 2013). In particular, Himalayan rivers, that are the lifeblood of India's societal and economic well-being, represent some of the most hazardous river systems in the world. In India, approximately 40 million hectares of land is prone to flooding (NDMA, 2008), of which the Ganga River basin accounts for nearly half of flood affected region (GFCC, 2012). Flood causes

higher loss to lives, properties, livelihood systems, infrastructure and public utilities than any other natural hazard. Despite structural and non-structural measures to control and manage flood hazards in India, deleterious flood impacts continue to persist and accrue (Agarwal and Narain, 1996). As a nation faced with continued urbanization, a changing climate, and ever increasing demand for flood-proof built environment, the study of floods hazards is an exceedingly urgent priority. To this end, a multidisciplinary understanding of flooding phenomenon and processes is needed.

Floods occur when water flow (discharge) exceeds the channel capacity, resulting in the sub-

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mergence of adjacent lands which is usually the floodplain, thereby causing or threatening to cause damage and loss to life and property. In hydrology, floods are defined as flow events with magnitudes much higher than the mean or usual flow. This broadens the definition of floods to include events that do not necessarily overflow the channel, especially in deep channels.

Floods are also important geomorphic events, because the processes of erosion and transportation during a flood event and deposition (in the recession phase of a flood) have direct bearing on river morphology and sediment and nutrient fluxes between channels and floodplains. Morphological and flux variability in the channel are critical determinants of river ecosystems and their health.

Floods occur on multiple space and time scales. An understanding of the spatio-temporal distribution of flood hazard is a prerequisite for the characterization of magnitude-frequency relationships of a hydrological system in a river basin. While floods occur throughout the country, major flood prone areas in the Indian rivers basins are mostly concentrated in the Himalayan River basins (Fig. 1). These river basins are also characterised by high elevation of upstream region responsible for higher sediment supply at downstream reaches, high rainfall intensity and larger basin area. All these physical characteristics contribute towards higher water discharge and sediment supply in downstream reaches. This study provides a review of the hydrologic and geomorphic processes associated with the spatio-temporal variability of flood events. The hydrologic and geomorphic characteristics of flood hazards have been used to define the types, processes and various factors responsible for flood hazards with special reference to Indian rivers. While physical processes critically underpin the understanding of flood hazards and is the focus of the present study, it is worth noting that a holistic understanding of the flood hazard also requires continued attention to the human dimensions, including governance and capacity building.

Types of Floods

Flood events may be classified on the basis of physical mechanism that causes the flood event. The following are the major flood types:

Overbank Flooding

Overbank flooding is the most common type. It is caused by high discharge or low bankfull channel capacity. Increase in discharge is governed by excessive or intense rainfall during wet seasons or due to heavy rainfall during cyclonic activity related with low-pressure systems (rainstorm-floods) or due to snow-melt contribution. Channel bankfull capacity, and hence overbank flooding is governed by aggradation-degradation processes, which are controlled by available flow energy and its mode of expenditure (Jain *et al.*, 2018). These channel processes are governed by a threshold defined through the ratio of specific stream power (ω) and critical stream power (ω_c) (Bull, 1979). Specific stream power is defined as the liberation rate of kinetic energy from potential energy per unit of bed area (Bagnold, 1966), and represents the available power for geomorphic work while critical stream power represents the power needed to transport the available sediment load. The threshold of aggradation-degradation in a river channel is defined as (Bull, 1979)–

$$\text{Threshold} = \omega/\omega_c \quad (1)$$

where, ω the specific stream power and ω_c the critical stream power. The specific stream power ($\omega = \gamma \cdot Q \cdot s / w$), is a function of unit weight of water (γ), discharge (Q), bed slope (s) and channel width (w). Critical power (ω_c) depends on the volume, density and grain size of sediment load (Q_s). Hence, channel processes and bankfull capacity are governed by these fundamental parameters.

Channel bankfull capacity increases due to erosion process, when threshold will be positive, that is, unit stream power exceeds critical power. However, negative threshold value leads to channel siltation and reduction in bankfull capacity because stream loses its sediment load. In general, sudden change in elevation along long profile (i.e., slope variability) is responsible in creating such conditions (Fig. 2). For example, river reaches at downstream of mountain front are mostly affected by aggradation processes, where threshold of aggradation-degradation will be towards aggradation process (Fig. 2). These reaches are characterised by sudden decrease in channel slope, which reduces specific stream power, whereas higher and coarser sediment supply from mountains reaches is responsible for higher value of critical

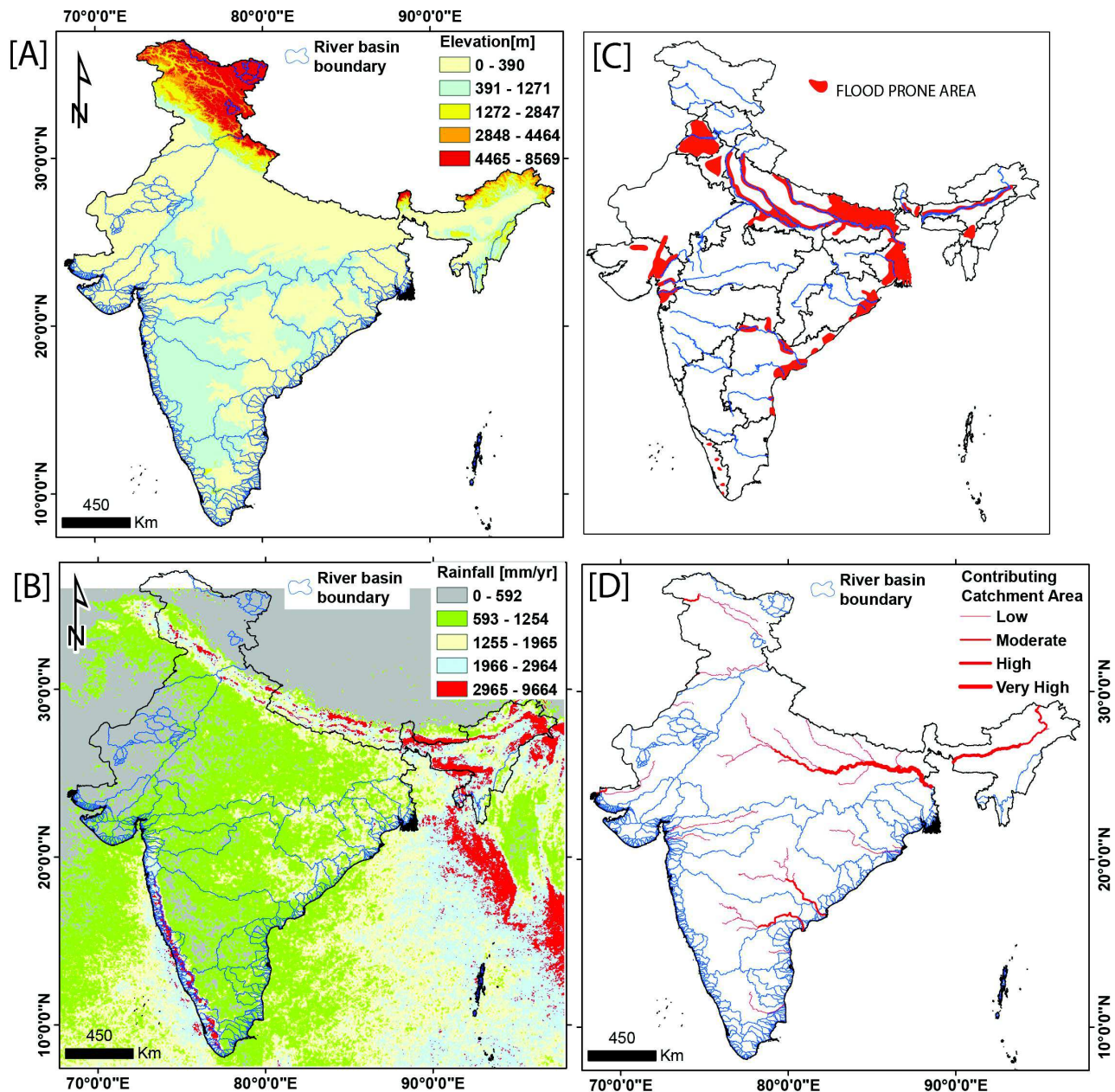


Fig. 1: [A] Elevation distribution in different river basin of India. The data used is Shuttle Radar Topography Mission (SRTM) Digital Elevation Model (DEM) of 90 m spatial resolution (data source: <https://hydrosheds.cr.usgs.gov/dataavail.php>); [B] Rainfall distribution over different river basins of India averaged over 12 years (1998 to 2009) (data source: <http://www.geog.ucsb.edu/~bodo/TRMM/>); Himalayan river basins are characterized by maximum rainfall [C] Flood prone areas of India (modified from NDMA guidelines on management of floods, 2008); The maximum flood prone area lie in Himalayan river basins [D] Variation of contributing catchment area in different river basins of India derived from SRTM DEM. Drainage boundaries were obtained from <https://hydrosheds.cr.usgs.gov/datadownload.php?reqdata=30bass>

stream power. Himalayan rivers draining north Bihar plains are typical example of such low stream power and high sediment supply channels, which are characterized by extensive siltation and frequent overbank flooding (Jain and Sinha, 2003a; Sinha *et*

al., 2005). In general, overbank flooding in a channel reach is a function of both climatological/hydrological and geomorphological processes in a river basin.

The ratio of flow energy to sediment transport is key to understanding channel processes and

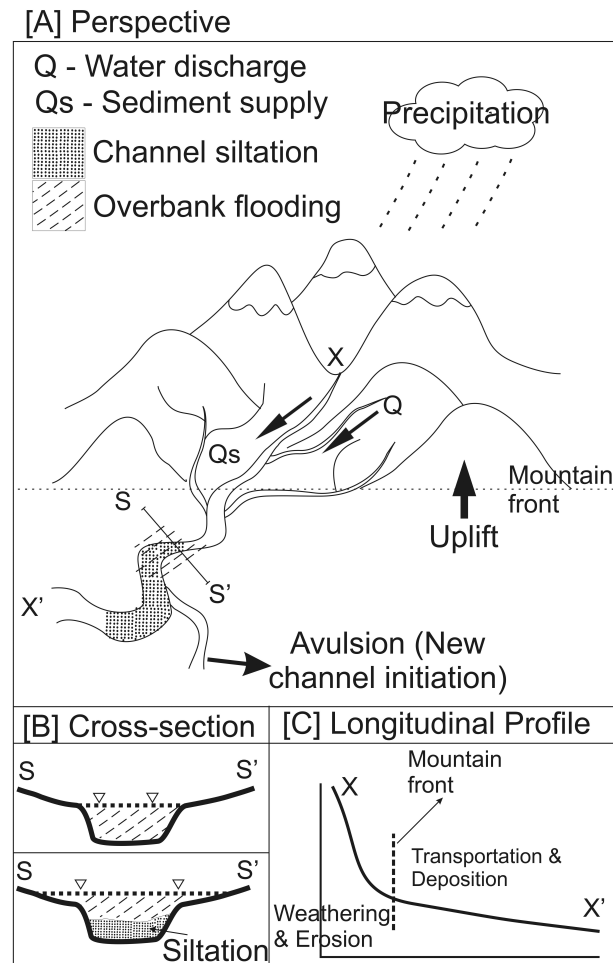


Fig. 2: [A] Important factors controlling the dynamics of riverine flood and [B] siltation reduces the water carrying capacity of river channel thereby causing overbank flooding; iii) Geomorphic processes along long profile exert significant control on spatial distribution of siltation process

morphological evolution of alluvial rivers. Huang and Nanson (2000, 2002) defined maximum flow efficiency (MFE) as an equilibrium state when sediment transport for a given stream power condition will be at a maximum. They quantified this state as a function of stream power and stream discharge on the basis of hydraulic analysis. The stability of a channel (the dominance of aggradation or degradation processes) is governed by the distance of a channel reach from the MFE state, because all channel reaches try to achieve this state through changes in morphological processes. Recent work shows that the concept of MFE can be successfully used to explain variability in channel processes and morphological appearance in different kinds of rivers (Bawa *et al.*,

2014; Nanson and Huang, 2017).

Channel Shifting and Avulsion Process

Channel shifting is a geomorphic event driven by the avulsion process, which causes flooding in a new area of the river basin. Avulsion is defined as the sudden change in flow direction of a channel often with complete discharge into the new channel. Flooding due to avulsion generally causes extensive damage because a new area far from the main river is affected, where flood preparedness is absent or insufficient. Location of this new flood affected area is governed by connectivity structure around the river channel (Sinha *et al.*, 2013). Avulsion occurs when the slope of the pre-existing channel becomes less than the across-channel slope. Such processes are common in the rivers on the aggrading plains in front of the Himalaya, where channel sedimentation quickly fills pre-existing channels producing the conditions for channel avulsion. The floods along the Kosi River and its tributaries in the north Bihar plains are well-known examples of an avulsion-driven flood hazard, where the channel bed was at higher elevation than the floodplain level (Wells and Dorr, 1987; Jain and Sinha, 2004; Sinha *et al.*, 2014; Jain *et al.*, 2018) (Fig. 3).

Outburst Floods

These are the most severe flood events and include engineered dam failures, landslide lake outburst floods (LLOFs) and glacial lake outburst floods (GLOFs). O'Connor *et al.* (2013) gives two broad classes of impoundments, which are potential sources of outburst floods viz., a) valleys blocked by ice, landslides, constructed dams, and volcanic materials and b) closed basins such as tectonic depressions and calderas, and those rimmed by glaciers and moraines. An outburst can be triggered by seismic activity, snow avalanche, landslide, volcanic eruption, intense snowfall, calving at the terminus of a glacier, piping of the earthen dam, or overtopping of a constructed dam (Fig. 4).

Glacier lakes can be formed due to the interception of melt water of a glacier by another advancing/surging glacier or the terminal moraines of the same glacier. Increased temperature contributes to its formation by hastening the velocity of the damming/advancing glacier and increasing the melting

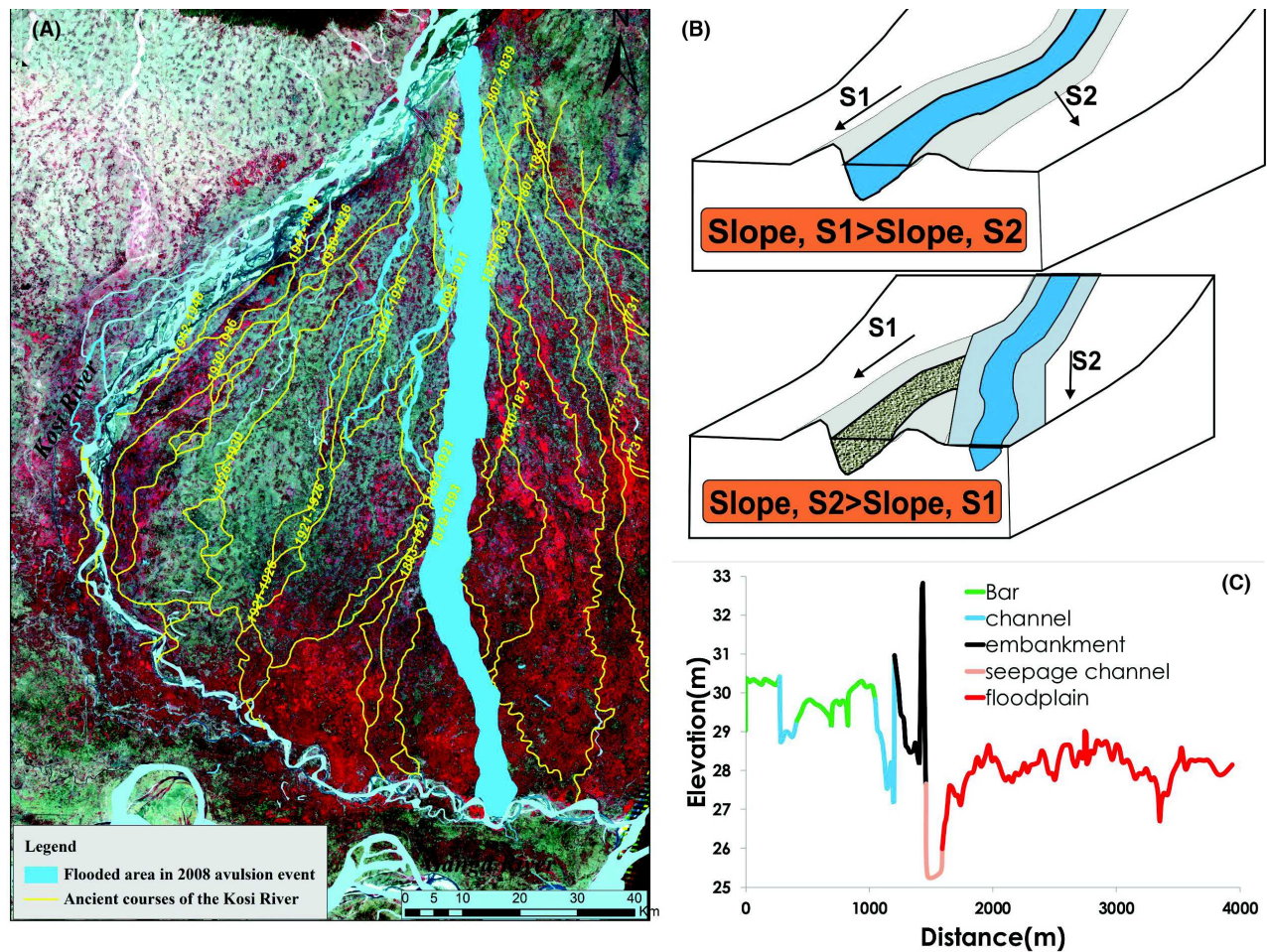


Fig. 3: (A) The Kosi River is a well-known example of channel shifting due to avulsion. A channel shifting event in 2008 caused major flooding hazard in an area which was far from the main channel. The River is dynamic because of frequent avulsions in the past. These avulsions are also responsible for around 110 km westward shifting of the Kosi River in the last 250 years. Palaeochannels were drawn from Gole and Chitale (1966). (B) A site becomes susceptible to channel avulsion when the slope of the potential channel (S_2) is more than the existing channel slope (S_1). (C) Channel cross section across the Kosi River near the avulsion site. The channel bed is at a higher elevation than the floodplain due to extensive sedimentation of the channel at the avulsion site. This super-elevated condition of the channel bed is responsible for a steeper gradient across the embankment and hence makes it susceptible to avulsion (modified after Sinha, 2009; Sinha *et al.*, 2014; Jain *et al.*, 2018)

rate of the glacier that contributes melt water to the lake. Himalayan rivers are also characterised by damming of lakes due to major landslides (Wasson *et al.*, 2008; Wasson *et al.*, 2013). Climate warming and monsoon intensification will enhance the likelihood of outburst floods in near future by increasing the number of landslide dammed lakes and glacial lakes (Wasson *et al.*, 2013; Westoby *et al.*, 2014). Such outburst floods bring about long lasting effects on landscapes, because these events generate flood discharge of several orders of magnitude than the channel capacity. Stream power of an outburst flood can attain

a magnitude much greater than those of large meteorological floods, can erode bedrock channels and transport large clasts with diameters upto 10's of meters. The outburst flood parameters such as peak discharge, duration, and volume of flood water depend on breach geometry, rate of its development and on the size and geometry of the impounded water body (Clague and O'Connor, 2015).

Other Types of Floods

In general, flood events are characterised by high velocity water flow and rapid inundation of an area

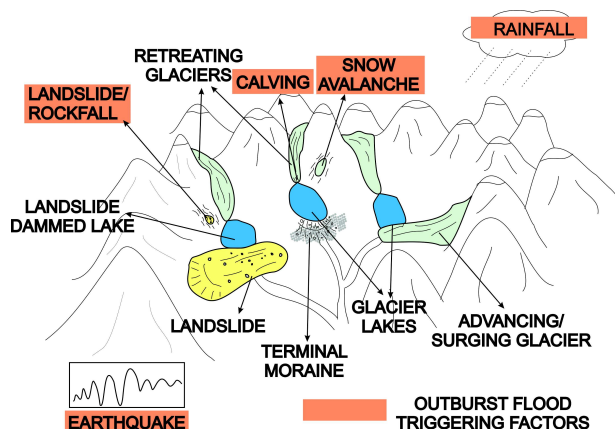


Fig. 4: A sketch of Glacial and Landslide Outburst Flood showing the causes and triggering factors of flash floods (highlighted in orange)

adjacent to river channels. However, in downstream reaches, where gradients are lower, and in deltaic reaches, floods can be slow moving with low velocities and long periods of inundation, presenting a different scenario of flood hazard. Masood and Takeuchi (2015) suggested an application of Flood Duration Curve (FDC) and Drought Duration Curve (DDC) to estimate the flood persistence in a river basin. In a comparative study of the Ganges-Brahmaputra-Meghna river basin, shape of FDC and DDC suggested that the floods in the Brahmaputra River would be long lasting in comparison to the Ganga River (Masood and Takeuchi, 2015). Further, one such slow moving flood was observed in the Chao Phraya river upstream of and in near Bangkok in Thailand (Liew *et al.*, 2016), where the floodplain is only around 3.5 to 1 m higher than MSL with a slope between 0.00001 and 0.00002 (Liew *et al.*, 2016). Flooding in 2011-12 caused inundation of a large area, where water remained on the floodplain for several months. The sediment-laden water also causes extensive sediment deposition on floodplains. In Thailand the slow moving large flood in a relatively flat region was exacerbated by overbank flooding both overbank and via levee breaching (Liew *et al.*, 2016).

Recently, flooding in urban areas (or 'urban flooding') has been enhanced by a number of anthropogenic and natural processes (Jha *et al.*, 2012). Flooding in urban areas causes considerable economic damage because of the exposure and vulnerability of a large number of people, buildings and infrastructure. Urban flooding is mostly governed by anthropogenic

factors related to rapid growth of urbanisation associated with poor planning and management of urban areas. Increased impervious areas, drainage blocking, reduction in the area of natural water bodies that can store floodwater, combined with high intensity rainfall causes extensive water logging and flooding (Barroca *et al.*, 2006; Jha *et al.*, 2012; Schmitt *et al.*, 2004). Increase in the frequency and magnitude of rainfall events in response to climate change may further increase the urban flooding (Schneider *et al.*, 2000; Ali *et al.*, 2014).

Causative Factors

Climatic Cause

Temporal and spatial distribution of precipitation determines runoff in a given basin. Using data from 93 catchments located in most parts of India, Garde and Kothiyari (1990) showed that the coefficient of variation of monthly rainfall along with catchment area accounts for 92% of the variance in the coefficient of variation of the annual peak flow series. Therefore, rainfall and catchment area are the dominant controls on the variance of floods India-wide, additionally prolonged precipitation and/or intense heavy precipitation from thundershower/cloud bursts over a short period of time increase river discharge. Furthermore, the direction of storm movement, intensity of a low pressure system, enhanced melting of snow and ice (e.g. rain falling on melting snow) can affect the amount of runoff which in turn characterizes the nature of a flood. Changed patterns and intensity of precipitation of rainfall caused by climate change will affect the magnitude and frequency of flooding.

Recently, the impact of climate change on flooding in the near future has been projected through different modelling approaches. In general, these studies suggest increased flood magnitude and frequency in the near future for most Indian rivers. Mirza *et al.* (2003) suggested enhanced flooding in the Ganga and Brahmaputra Rivers in response to climate change, though peak discharge increase will be more in the Ganga River basin in comparison to the Brahmaputra River. Gosain *et al.* (2006) provided projections for Indian rivers by integrating HadRM2 daily weather data with the distributed hydrological model SWAT (Soil and Water Assessment Tool). Their work on 12 large rivers highlighted that a few

rivers (namely Godavari, Brahmani and Mahanadi) will face severe floods, however other rivers may witness a reduction in water availability in response to climate change. Another SWAT-based study by Mishra and Lilhare (2016) projected future stream flow in the major Indian rivers using the representative concentration pathways (RCP) 4.5 (RCP 4.5) and 8.5 (RCP 8.5) to represent different scenarios of climate change. Mishra and Lilhare (2016) suggested more runoff in the Near-(2010-2039), Mid-(2040-2069) and End-(2070-2099) term. Climate change will lead to increases in streamflow during the monsoon season and hence will enhance flooding in most of the river basins in India. Near-term projections suggest increased monsoon flow in only a few rivers, namely the Ganga, Brahmaputra, Narmada and South Coast rivers (coastal rivers in Kerala and Tamil Nadu), however most Indian rivers will be affected by increased stream flow by upto 40% by the end of the 21st century.

A set of studies using a process-based, semi-distributed model (viz., the Integrated Catchment Model INCA-N) also suggests an increase in monsoon flows and more frequent flooding in the near future due to climate change (Whitehead *et al.*, 2015; Khan *et al.*, 2018). Whitehead *et al.* (2015) made projections for the 2050s and 2090s and highlighted major increases of water flow during monsoon seasons. Further, flood hazard in river systems is not only governed by water flux but sediment load also plays an important role. Variability in sediment flux in response to climate change has also been analysed by some workers. For the Ganga River an increase (relative to the 1981-2000 baseline period) of 16-18% for the mid-century and 34-37% at the end of the century was projected by Darby *et al.* (2015) by using an empirical lumped sediment transport model (HydroTrend). Khan *et al.* (2018) further applied the Integrated Catchment Model for sediments (INCA-Sed) and standard HadCM3 coupled GCM (global climate model) to project discharge and sediment load variation in the Ganga River basin. Khan *et al.* (2018) projected an increase in peak flows and flood frequency for the mid-century (2050s) and end-century (2090s) in response to climate change. The model was also used to understand sediment transport behaviour in the Ganga river system in the near future due to climate change. Higher flow during the

monsoon season will also enhance erosion processes and will lead to enhanced sediment flux in the Ganga River system. Enhanced sediment supply will further aggravate flooding if bank full capacity is reduced by aggradation and/or reducing channel stability by creating avulsive conditions. The model projected increased sediment loads of 10-40% by mid-century and 35-79% by the end of century.

River Basin Characteristics

Runoff generation from rainfall and its distribution over the surface is governed by geologic and geomorphic characteristics along with land use-land cover patterns. The permeability property of lithology and soil controls runoff and hence has a bearing on the nature of a flood. The time lag for flooding to occur after a precipitation event is determined by the moisture condition prior to the precipitation event. A wet, saturated soil will generate runoff of higher magnitude with a low lag time after a precipitation event.

Floods also assume different characteristics based on basin morphometric parameters such as shape, relief, slope, topographic configuration, drainage density and valley geometry. Also, drainage network characteristics are important in understanding the hydrologic response of a basin (Rodriguez-Iturbe and Valdes, 1979). Identical flood generating mechanisms (for example, similar amount of precipitation) may result in very different types of floods in different catchments or within a catchment at different points of time (Rodriguez-Iturbe and Valdes, 1979; Jain and Sinha, 2003b). Downstream variability of flood severity along the Himalayan river channels has been observed in response to geomorphic and tectonic setting of the basin area (Jain and Sinha, 2003c; 2005), drainage network organisation (Jain and Sinha, 2003d) and long profile shape (Devrani *et al.*, 2015). Further, rivers draining an uplifting orogen like the Himalaya erode and transport more sediment resulting in more rapid aggradation on the plain which reduces the bankfull capacity of the channel and makes it more prone to avulsion and flooding (Sinha and Jain, 1998). Basin geometry near the sea and the nature of a coastline determine the intensity of flooding in coastal areas caused by tsunamis or cyclonic storm surges. Tidal influence in a narrow or wide estuary can greatly change the nature of flooding.

Anthropogenic Causes

Humans have contributed to aggravating floods in several ways. Increased population that has driven unplanned urbanisation and industrialisation near rivers (particularly construction on flood plains) has led to obstruction of floodways and greater damage and loss of life. Urbanisation is responsible for increased runoff generation due to decreased infiltration capacity as built and paved areas expand. Deforestation with associated reduction in infiltration of rainwater and increased runoff and soil erosion have aggravated flood hazards by increasing sedimentation in channels and discharge in river system. Construction of dams and roads also increase sediment input into rivers thereby reducing the bankfull capacity of channels. Some approaches to flood control, such as building embankments, further leads to aggradation of river channels by restricting flow and increasing the amount of sediment in the channel (Raymahashay and Sinha, 2012). This heightens flood risk as a river channel is bound to avulse towards a new path by breaching the embankment. Rail and road network on embankments in aggrading areas like in north Bihar plains are further worsening the flooding scenario by reducing the hydrological and geomorphic connectivity within the river basin (Kumar *et al.*, 2014).

Methods of Estimating Flood Magnitude

Thousands of floods—small and great—occur annually. Nevertheless from time to time, meteorological and hydrologic conditions combine to generate flood events that engender loss of life and property in the floodplain and coastal areas (IFCRCRS, 1997; Hoyt and Langbein, 1955). For designing effective structures and strategies that help reduce flood-related hazards, good estimates of possible peak flow magnitude/s are needed (Linsley, 1986). Given the general paucity of comprehensive hydrological datasets, no methodology is yet available to determine precisely the flow magnitudes that may occur in a given year. However, the following are the most commonly employed methods for estimating the peak discharge for a location:

- a. Physical indications of past floods
- b. Rational Method
- c. Empirical formulas and envelope curves

- d. Unit hydrographs
- e. Flood Frequency Studies

It should be noted that the application of any of the above mentioned methods for appraising peak flow discharges depends on the (i) desired objective (ii) available data (iii) desired level of protection (Subramanya, 1994). Reliability and performance of these peak flow estimation techniques is given in Linsley (1986).

Physical Indications of Past Floods

Historical flood level marks at sections of rivers, trees, bridges or buildings are used to reveal the magnitudes of peak flows in past. To this end, the cross section of the river channel is measured during the dry season to calculate the cross-sectional area. By adopting an appropriate value of Manning's n to represent the roughness coefficient of the bed and sides of the channel, the velocity may be computed using the Manning's or Chezy's formula. The peak discharge is then calculated as a product of the velocity and river cross-sectional area.

Empirical Formulas and Envelope Curves

These are simplified flood estimation methods that are based on the statistical relationship between observed flood peaks and important catchment properties such as catchment area, basin relief, rainfall and drainage network characteristics. Only a few of the parameters governing the peak discharge are used in these equations, and consequently these formulae perform well in regions where they are developed (Chow *et al.*, 1988). While there are a number of such empirical formulas for various parts of the world, they can generally be categorized into the following three groups:

1. Formulas in which peak discharge is a function of watershed area:

$$Q = f(A) \quad (2)$$

where, A is drainage area

2. Formulas in which the peak discharge is related to catchment area, basin topographical and meteorological characteristics

$$Q = f(A, h, s, P) \quad (3)$$

where h is local relief, s is slope parameter and P is precipitation intensity

3. Formulas in which parameters such as rainfall intensity (I), time of concentration (t_c) and return period (T) are involved

$$Q = f(A, I, t_c, T) \quad (4)$$

Chow *et al.* (1988) and Subramanya (1994) lists several of these empirical formulas.

Rational Method

This is the most commonly applied method for estimating peak discharge in smaller watersheds (Pilgrim and Cordery, 1993; Young and McEnroe, 2014). The origin of this method is not clear however, in the United States, Linsley *et al.* (1992) was the first to mention such approach in the scientific literature. This method has been found to be effective for peak flow predictions in small sized ($A < 1$ acre) catchments (Linsley *et al.*, 1992) and finds considerable application in urban drainage design such as for storm drains and in the design of culverts and bridges.

The rational method is based on a simple approach that relates the peak discharge to rainfall intensity for a particular length of time (time of concentration) and watershed drainage area. Often, the functional form for the rational method is given as

$$Q_p = C.i.A \quad (5)$$

where Q_p is the peak discharge; C is the runoff coefficient; i is average rainfall intensity and A is drainage area. The runoff coefficient (C) is a dimensionless empirical coefficient related to the catchment losses, and thus depends on the nature of the catchment surface, surface slope and rainfall intensity. Furthermore, C ranges between 0 and 1 where a value of 0 (1) indicates that none (all) of the rain falling on the basin is converted to runoff. Composite, volumetric and rate-based methods can be applied for estimating the runoff coefficient in a specific catchment (Young *et al.*, 2009; Dhakal *et al.*, 2011; Dhakal *et al.*, 2012). On the other hand, rainfall intensity (i) corresponds to the average precipitation intensity selected from intensity-duration-frequency (IDF) curves for a specified frequency (probability) using the duration equal to the time of

concentration, t_c , of the watershed (Viessman *et al.*, 1977). The time of concentration (t_c) is often defined as the minimum time required for a parcel of water to travel from the most hydraulically distant point in the watershed to the outlet (McKay, 1970; Wigham, 1970). It depends on lengths of channel and overland flow paths, average slope of a watershed, and roughness characteristics of the watershed. The IDF relations for a region are generated by fitting annual maximum rainfall intensities for a specific duration to a Gumbel-probability distribution (McKay, 1970), while several empirical methods are available for estimating the time of concentration such as the methods of Morgali and Linsley (Morgali and Linsley, 1965), Kirpich method (Kirpich, 1940), Kerby-Hatheway method (Kerby, 1959).

The limitation of the rational method is that it is developed for relatively frequent events. Consequently, the peak flow rate as calculated by the rational method should be increased for more extreme events (Viessman and Lewis, 2003).

Unit Hydrograph Method

For small to medium sized catchments ($A < 5000 \text{ km}^2$), a flood hydrograph can be developed using its unit hydrograph, which is defined as the direct runoff hydrograph (stream flow vs. time) resulting from a unit volume of precipitation-excess (typically 1 cm) of constant intensity and uniformly distributed over a catchment area (Chow *et al.*, 1988; Maidment, 1993). Precipitation excess is the fraction of the total precipitation, that is available for direct surface runoff. Sherman (1932) introduced the concept of the unit hydrograph and the principal concept underlying the application of a unit hydrograph is that for a given precipitation duration, the unit hydrograph does not change unless the watershed physiography changes. Consequently, flood hydrographs for real and complex precipitation events generally can be simulated by superimposing and adding unit hydrographs (Dooge, 1959). If dealing with precipitation of different durations, the linearity property implicit in the Unit Hydrograph analysis can be used to generate Unit Hydrographs associated with larger or smaller excess precipitation duration. This procedure is sometimes referred to as the S-curve Hydrograph method (Haan *et al.*, 1994).

For a site with a gauged stream flow and precipitation records, a unit hydrograph can be generated from a trial-and-error analysis of recorded flood hydrographs (stream flow vs. time) and hyetographs (precipitation vs. time) (Linsley *et al.*, 1992). First base flow is determined (e.g. using the Area method, the master recession curve and precipitation loss (e.g. using the f-index, Horton, Green and Ampt to determine infiltration losses) need to be estimated and subtracted from recorded flood hydrographs and hyetographs to compute direct runoff and the hyetograph of precipitation excess. Base flow and precipitation loss separation is somewhat arbitrary, however it is to theoretical concepts of catchment response. Next, a unit hydrograph procedure that relates unit-hydrograph shape and timing to the recorded hydrograph and hyetograph is used to derive the unit hydrograph. If the calculated hydrograph of direct runoff plus base flow does not match the recorded hydrograph, the unit hydrograph variables are adjusted and the trial and error procedure is repeated. When the computed hydrograph of direct runoff plus base flow closely matches the recorded hydrograph, the resultant unit hydrograph is considered appropriate for the catchment.

For sites where direct precipitation and stream discharge observations are not available, flood hydrographs to be utilized for design purposes are developed by using synthetic hydrograph procedures. Synthetic unit hydrograph procedures can be categorized as those based on (1) models of watershed storage (e.g. Nash, 1957; Nash 1959; Dooge 1959); (2) the relationship between hydrograph characteristics (e.g. time to peak, peak flow) and watershed characteristics (e.g. Snyder, 1938, Geomorphologic instantaneous unit hydrograph) (Rodriguez-Iturbe and Valdes, 1979); and (3) a dimensionless unit hydrograph (e.g. SCS 1972).

A limitation of the unit hydrograph approach is the underlying assumption of spatial uniformity of precipitation (Chow *et al.*, 1988; Maidment, 1993).

Flood Frequency Analyses

Floods are complex natural events that are outcomes of the multiple interactions between several precipitation (e.g. intensity, duration, spatial coverage) and catchment characteristics (e.g. antecedent moisture conditions, land cover, slope). They are

therefore very difficult to model analytically. An alternative method for predicting flood peaks is the statistical approach of flood frequency analysis. This technique involves the analysis of flood peak data over an extended period of time using probability distributions that relate flood peak magnitudes to their frequency of occurrence (Cunnane 1987; Chow *et al.*, 1988). In flood frequency analysis, the underlying assumption is that floods are outcomes of stationary, independent and identically distributed stochastic processes. Often this assumption is found invalid, and questions about this assumption have been discussed extensively in published studies (Yevjevich, 1968; Klemes, 1987; Jain and Lall, 2000, 2001).

Three types of flood peak series are employed in flood frequency analyses: the annual maximum series (AM), peaks over threshold (POT) series, and the time series (TS) model. The annual maximum flood series is based on the maximum peak flow for each year (Cunnane, 1973). However, the use of AM may involve some loss of information as the second or third highest peak floods within a year, which may exceed many of the annual maxima, are ignored (Kite, 1977; Chow *et al.*, 1988). In POT, all flood peaks above certain base value are considered for analysis regardless of the time interval. The base value (threshold) is usually low enough to include at least one flood event for each year (Ashkar and Rousselle, 1983; Ashkar and Rousselle, 1987). The disadvantage of POT is that observations may not be independent. The time series model is based on the properties such as trend, and seasonal or stochastic processes deduced from a series of flood peaks at equally spaced time intervals (Matalas, 1967; Adamowski and Oosterveld, 1975). The commonly used time intervals for flood frequency analysis are in days, although months or years can also be used. The relative merits of TS, POT and AM are discussed in various papers and reports (Chow, 1964; Cunnane, 1973; NERC, 1975; Kite, 1977; Rosbjerg, 1977; Tavares and Da Silva, 1983; Chow *et al.*, 1988; Cunnane, 1989; Madsen *et al.*, 1997; Coles, 2001).

From ordered peak flood data, the flood frequency analysis approach aims at estimating the probability of exceedance or return interval of flood peaks. Probability of exceedance is the likelihood that a given peak flood magnitude will be equalled or exceeded in each year. Recurrence interval (also

referred to as return period) is the average time period usually expressed in years when a flood peak of given magnitude will be equalled or exceeded (Cunnane, 1989). Recurrence interval is the reciprocal of probability of exceedance. For example, a flood peak having a probability of exceedance of 0.01 each year has a return period of 100 years. This means that the flood peak can occur on average once every 100 years. But it should be noted that floods do not occur in exact cyclic events and therefore it is conceivable that a 100-year flood peak event can occur in several consecutive years. Exceedance probability is given by:

$$P = \frac{1}{T} \quad (6)$$

$$P' = 1 - P$$

where P is the probability of exceedance of a flood having a return period T -year/s and P' is the probability of non-exceedance (probability that the flood of equal or greater magnitude will not occur for each year).

Flood frequency analysis includes the fitting of extreme value distributions to measured peak flows to estimate flood frequency curves. However, there is no physical basis to select any particular type of probability distributions for flood frequency analysis. Often the range of the variable in the distribution function, the general shape of the distribution, and parameters such as skewness and kurtosis are used to indicate whether a particular distribution is appropriate for a given flood peak distribution. Log-Normal (Durrans, 1992; Haktanir, 1992), Pearson type III (Chowdhury *et al.*, 1989; Wu *et al.*, 1991), Gumbel (Koch, 1991; Guo and Cunnane, 1991), log Pearson type III (Durrans, 1992; Ashkar *et al.*, 1993), Generalized extreme value (Hosking, 1986), Generalized Pareto (Wang, 1991; Rosbjerg *et al.*, 1992), log-logistic (Ahmad *et al.*, 1988a; Ahmad *et al.*, 1988b; Singh *et al.*, 1993) distributions to name a few have been used and may be appropriate for a given sample of data. To check the validity of a distribution, goodness of fit tests (e.g. χ^2 , Kolmogorov-Smirnov tests, Akaike's information criterion, L-moment, Cramer von Mises) are used. Even if a statistically acceptable distribution is selected, proper estimation of parameters is essential as some of the parameter estimation methods may not yield good

estimates, or even converge. The method of moments (Guo and Cunnane, 1991), maximum likelihood functions (Koch, 1991), probability weighted moments (Wang, 1991; Greenwood *et al.*, 1979), L-moments (Hosking, 1990; Stedinger *et al.*, 1993) are the most commonly applied methods for estimating parameters.

Regional Analysis

Flood frequency analysis is a challenge, as sufficient flood discharge data is not always available at a site to determine the frequency of rare flood events. In such instances, hydrologists use the regional flood frequency relationships, based on pooled flood data from many 'hydrologically similar' basins, for flood risk estimation at-site to compensate for short data records relative to the return period of interest. Hydrologically similar basins are group of watersheds with sufficient homogeneity in flood generating mechanisms, and the procedure to identify such basins is referred to as regionalization.

Two successful examples of regionalization are the index method, and the regional regression method. The basic concept underlying the index procedure is that the distribution of floods at every site in a region be the same except for the scale parameter (or index flood) that reflects the size and runoff characteristics of each basin (e.g. Wallis, 1980; Lettenmaier *et al.*, 1987). On the other hand, regional regression techniques employ simple models that predict peak flows exceeding various probabilities as a function of physiographic basin characteristics and other parameters. Flood regionalization methods have the advantage over other methods in deriving design floods from design rainfall because (i) they are easier to apply (ii) there are no questions about the relationships between rainfall and flood exceedance probabilities (iii) no assumption needed about antecedent watershed conditions. A review of the regional analysis approach, and its merits can be found in Cunnane (1988).

Non-stationarity of Flood Frequency Analysis

Hydrologists have long employed the assumption of stationarity—variables whose underlying statistical properties do not change over time—as a foundational concept in flood frequency analysis. This in turn implies that the flood peaks are outcomes of physical processes that remain unchanged over time, and that the probability of occurrence of a T -year flood peak

is the same for each year. On the other hand, there is a body of evidence that indicates the presence of temporal trends or other non-stationarities (e.g. shifts, clusters) in peak flood flows for various regions (Changnon and Kunkel, 1995; Porporato and Ridolfi, 1998; Jain and Lall, 2001; Groisman *et al.*, 2001; Cox *et al.*, 2002). Such dependence in peak flood characteristics can arise from natural variability (e.g. low-frequency climate variability such as ENSO, natural climate change) as well as from anthropogenic influences (e.g. land-use/cover changes, installment/removal of hydraulic structures, climate change).

Conventional approaches for flood frequency analysis will not work for non-stationary time series data as the distribution parameters, and the distribution itself change over time. Furthermore even mild trends (linear or non-linear) in the statistical characteristics of peak flood flow have been revealed to have strong impact on the occurrence and magnitude of flood hazards (Porporato and Ridolfi, 1998; Cox *et al.*, 2002; Cunderlik and Burn, 2003). Such evidence have led to the development of flood models or methods in which (a) the concept of non-stationarity is loosely or not at all imposed (b) parameters (e.g. mean and variance) of a given distribution change with time or other geophysical dependencies. They include (a) models considering covariates (Coles, 2001; Katz *et al.*, 2002) (b) probability distributions embedded with trend components (Strupczewski *et al.*, 2001) (c) probability distributions with mixed components (Waylen and Caviedes, 1986) (d) stochastic models considering shifting patterns (Sveinsson *et al.*, 2005). A review of the basic concepts of non-stationary hydrologic extreme events and current approaches to at-site and regional flood frequency of non-stationary peak flood flows can be found in Khaliq *et al.* (2006) and Obeysekara and Salas (2014).

Geomorphic Characteristics of Floods

Flood Power

The capacity of a river to erode and transport sediment increases manifold during a flood event. The concept of flood power quantifies such an increase in a river's ability to perform geomorphic work. Flood power is defined as the energy available to carry out geomorphic work during a flood event. The geomorphic work by a flood event includes erosion of beds and banks of the channel and entrainment

and deposition of eroded material. Bed shear stress and/or unit stream power are good measures of the geomorphic effectiveness of a flood. Several studies have shown that the flood power rather than flood magnitude is more significant with regards to geomorphic work (Baker and Costa, 1987; Kochel, 1988). It has been parameterized in terms of channel boundary shear stress (τ) and flood power per unit area of bed (ω), which are given as -

$$\tau = \gamma RS \quad (7)$$

$$\omega = \frac{\gamma QS}{w} \quad (8)$$

In the case of wide channels, where the width-to-depth ratio is greater than 20, the above equations are reconstituted by replacing hydraulic radius, R with mean flow depth (D) -

$$\tau = \gamma DS \quad (9)$$

$$\omega = \frac{\gamma QS}{w} = \gamma DVS = \tau V \quad (10)$$

where, τ is the boundary shear stress (in N/m²), ω is unit stream power (in W/m²), γ is the specific weight of water (9800 N/m²), Q is discharge (in m³/s), S is the slope (in m/m), w is the water surface width (in m), R is the hydraulic radius or mean depth of water (in m), and V is the average flow velocity (in m/s).

As the above equations suggest, higher slope, discharge, average flow velocity, mean flow depth and low channel width or valley width increases bed shear stress and flood power thereby increasing the ability of a flood event to do more geomorphic work. In general, these parameters are expected to follow a fairly uniform trend from upstream to downstream. For instance, slope decreases and channel depth increases downstream. But local geologic and geomorphic factors can directly or indirectly affect these parameters to give rise to an irregular pattern of flood power distribution along a river. Downstream distribution of flood power is strongly controlled by valley width (Magilligan, 1992). A narrow channel or valley, which is mainly controlled by lithology, results in high flood power whereas an abnormally wide floodplain results in significantly lower flood power for the same flood event. Further, as the unit stream power increases for narrower and deeper channel,

flood power in a river channel further increases by significantly increasing their depth and velocity during large-magnitude floods (Kale, 2007). The overall spatial distribution pattern of flood power thus reflects both the control by local factors and the general control of basin characteristics. Also, the nature of the channel also controls the spatial distribution of flood power. In bedrock reaches, the unit stream power values are higher by one or two orders of magnitude than in the alluvial channels (Baker and Kale, 1998). Hydraulic modelling has also shown that flood produces high flood power in a confined reach compared to an unconfined reach (Thompson and Croke, 2013). For a given site, a large magnitude flood results in a nonlinear increase in flood power. In general, variations in rainfall intensity and distribution, basin geomorphic and geologic characteristics and channel geometry cause spatio-temporal variation in flood power.

Sediment Transport

Flood events strongly affect the sediment budget through sediment erosion, remobilisation, transportation and depositional processes. The estimation and prediction of sediment budget (a quantitative account of sediment inputs, outputs and storage in a known volume of river channel, floodplain or catchment) has great significance in hazard assessment and engineering applications. Further, the morphologic response of channels to large floods is dominantly governed by high volumes of sediment transport. The total bed material transportation during a large event is also used as an important indicator to characterise a large flood event (Eaton and Lapointe, 2001).

Stream capacity and competence of a flood event are used to analyse the impact of floods on a sediment budget. Competence of a flood refers to the largest particle a stream can mobilize during the event whereas capacity refers to the total sum of load a stream can mobilize, entrain and transport. Capacity and competence are greatly increased during a flood, which results in bulk sediment movement. Floods not only transport bedload, suspended load and dissolved load, but can also generate and transport debris flows in the headwater catchments (Rickenmann and Koschni, 2010).

Measured discharge and suspended sediment

load show variable relations besides the often reported lag between the peaks in suspended sediment concentration and discharge. The proportion of suspended sediment in the total sediment load in mountainous rivers varies from 20% to 90% (Bathurst *et al.*, 1987; Whittaker, 1987; Billi *et al.*, 1998) compared to 70% to 95% in low-land rivers (Walling and Webb, 1981). Among the Indian River basins, the Himalayan rivers are characterised by 90-99% of transport of suspended sediment load (Sinha and Jain, 1998; Bawa *et al.*, 2014), though it may be lower (65-95%) in the Peninsular rivers (Fig. 5).

Flux measurement during floods has many practical difficulties, which necessitates the use of new approaches for flux estimation. Sediment transport of an event can be estimated inversely by calculating the net changes in sediment storage within the channel by computing the topographic difference between pre- and post-event (Carson and Griffiths, 1989; Lane *et al.*, 1995; Martin and Church, 1995; Ham and Church, 2000). This method is called the “morphologic” (or “inverse”) method for estimating bed material transport although it only provides an estimate of the net result of all the processes at work. Availability of high resolution digital elevation data, such as from LiDAR, has made this approach feasible.

Theoretical equations based on shear stress and/or unit stream power are used to reconstruct/estimate/predict the sediment transport rate (Gomez and Church, 1989). However, results of the theoretical equations can be in substantial discord with the actual transport in natural stream. Further, sediment transport can differ widely in the mountains and lowlands. Sediment transport in mountainous streams displays a greater variability from event to event and within events during the snowmelt period or during single high-magnitude floods, in comparison to larger rivers in low relief areas. Application of theoretical sediment transport equations can also be rendered problematic by the local constraints such as the presence of woody debris, channel roughness and bedrock constriction (Rickenmann and Koschni, 2010).

Geomorphic Response of Rivers

Floods are important in creating geomorphic diversity in channels. The geomorphic response of rivers to floods can be seen in the framework of driving and resisting force. The flood power and shear stress

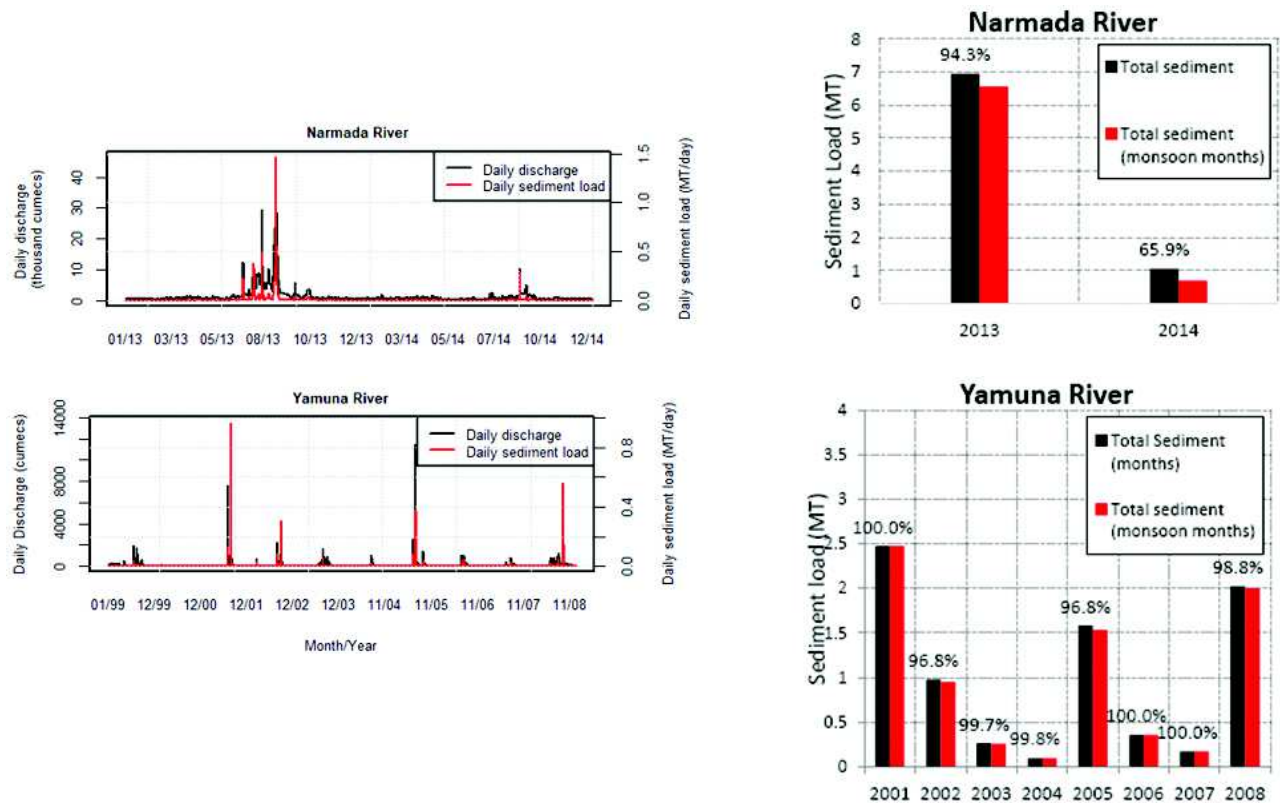


Fig. 5: Sediment transport and discharge characteristics of Himalayan and Peninsular river during monsoon period. Flux variability in a year is governed by cyclic pattern, where most of the flow and sediment transport occur during monsoon period. Flow during monsoon period is responsible for 99-95% of suspended sediment transport in a Himalayan river and 90-65% of suspended sediment transport in a peninsular river

represent the driving force. The driving force is resisted by inherent geomorphic characteristics of a river system which is a function of the erodibility of beds and banks (a function of lithology and material chemistry), channel geometry, channel form, sediment supply and sediment grain size, soil type and vegetation. The dominance of degradation or aggradation in a river response is determined by the balance of the two forces at play. It will also determine the effectiveness of floods in sculpting the morphology of landscape in the long term. When driving force is dominant, the river will erode and transport. But when the resisting force exceeds driving force, deposition is dominant. Degradation occurs with high discharge, high slope, low sediment load and fine sediment size. Aggradation on the other hand is favoured by high sediment load, coarse sediment size with relatively low discharge and a gentle slope.

The geomorphic sensitivity of river channel in response to floods is also governed by the nature of these driving and resisting forces. Drastic modification

of valleys can occur during catastrophic flood events, while, in some other cases, the channel response to floods can be muted owing to high resistance of the channel or low channel gradient. Floods likely to result in significant geomorphic changes are those that produce flood power higher than the threshold of geomorphic change; i.e. flood power should be high enough to entrain the bed sediments and exceed the resistance provided by bed and banks (Baker and Costa, 1987). For alluvial channels in wet climates, 300 W/m² of unit flood power or 100 N/m² of bed shear stress has been identified as the minimum threshold for geomorphic change (Magilligan, 1992). The spatial pattern of deposition and erosion is generally governed by the downstream variation in flood power distribution. As gradient and flood power decreases downstream, erosion is replaced by deposition. In general, narrow reaches are characterized by net erosion, whereas wider unconfined reaches are characterized by net deposition (Thompson and Croke, 2013).

Effects of flood on rivers can be observed at different scales with significant spatial variability. A large flood may affect the channel planform and channel geometry through channel widening, erosion or deposition on the channel bed, overbank deposition, floodplain stripping or scouring in alluvial streams. Channel widening is more common in sparsely vegetated arid and/or alpine environments (Huckleberry, 1994; Warburton, 1994). Geomorphic changes in bedrock channels are characterized by bed-scouring and channel erosion through hydraulic action and cavitation, deposition of gravel bars or erosion of gravel deposits and deposition of sand and fines at tributary mouths, on bends and in other protected areas as slackwater deposits and/or deposition of boulder berms.

Flood waves in a river channel give rise to a complex assemblage of morphological units that may vary from reach to reach depending on the relative dominance of erosion and deposition processes (Sridhar, 2007; O'Connor *et al.* 2013; Yousefi *et al.*, 2018). Depositional landforms include debris flows, lahar deposits, large boulders (upto 10s of meter diameter), and large bars (mostly above present day flow level). Reaches dominated by erosion are characterized by large channel scours, severe bank erosion and eroded channel tracts and/or significantly wider and deeper channels. Some of these geomorphic features, namely flood trim lines, slack water deposits (SWD), flood benches, and gravel and boulder bars are used to quantitatively estimate palaeoflood flows (Kale *et al.*, 2000; Benito and O'Connor, 2013).

Palaeofloods

Floods of the distant past that were not recorded using the standard modern techniques of gauging and modeling or were not recorded in history are referred to as palaeofloods. Imprints of these floods are inferred from sediment deposits and effects of floods on landscapes and vegetation. These palaeoflood indicators include:

- a) Slack water deposits (SWD) - these are coarse grain (and sometimes fine grain) sediments that are entrained during the high energy flow stage and are deposited in sheltered locations as the flow recedes. Because of good preservation potential, they form useful palaeoflood indicators (Baker *et al.*, 1983).

- b) Other palaeostage indicators (PSI) include silt and stains on the bark of trees, flotsam and erosional trim lines, and injury marks on trees caused by moving boulders entrained during a flood. Landscape position controls the accuracy of these palaeostage indicators. High-water marks are often more accurately indicated in caves and on sheltered terrace surfaces (Kite *et al.*, 2002).

Palaeoflood indicators help in deducing palaeoflood stages, followed by estimation of palaeodischarge and inverse modelling of the palaeoregime (Baker, 2008). Palaeoflood interpretation is an interdisciplinary affair and requires inputs from geochronology, sedimentology and stratigraphy (Kite *et al.*, 2002). New geochronological techniques and improved hydraulic modelling have improved the results of palaeoflood studies. Palaeoflood estimations have been carried out for various rivers of India using sedimentological and geomorphic features. Studies exist for the Luni River in western India (Kale *et al.*, 2000), Tapi, Narmada, and Mahi rivers in central India (Ely *et al.*, 1996; Kale *et al.*, 2003; Sridhar, 2007), Kaveri River (Kale *et al.*, 2010), and the Alaknanda River catchment (Wasson *et al.*, 2013). These datasets provided evidence of major flood events in the last few thousand years. The palaeoflood records show large scale temporal variability of flood frequency in the past with some periods of clustering of flood events over short periods of time. The clustering of large flood events is closely related with Holocene climate change in India. In general, flood magnitudes in the past millennium were significantly larger in comparison to the modern day flooding based on historical data. Only on the Kaveri River were the 20th century floods larger than the mid-Holocene palaeofloods (Kale *et al.*, 2010).

Palaeoflood Studies Have Applications in the Following Areas

- a) Role of Climate change on flood hazards - Palaeoflood studies provide flood records from the distant past thereby providing much longer records of floods than can be derived from gauged records, enabling the role of climate change to be clearly analysed. Many incidences of palaeofloods have been found to be correlated with wet climatic regimes (Ely *et al.*, 1993; Ely,

1997; Knox, 1993; Knox, 2000; Benito *et al.*, 2003). In general, flood events tend to concentrate during humid periods highlighting a strong dependence on climate and change regimes (Kale, 2012; Wasson *et al.*, 2013; Foulds *et al.*, 2014). The temporal distribution of extreme floods in large rivers of South Asia is characterised by clustering of extreme floods around 840 and 2000 AD, and show a significant link with above-normal rainfall during the same time. Similarly, flood magnitude and frequency in Peninsular India was reduced significantly during the Little Ice Age (1300 to 1850 AD) (Kale and Baker 2006).

- b) Flood planning and flood risk assessment – Using only the gauged data in flood risk assessment has the effect of under assessing flood risk because these records are not long enough to capture the real extremes. Hence integration of palaeoflood data with gauged and historical data can improve flood probability estimation through flood frequency analysis (Sheffer *et al.*, 2003; Black and Fadipe, 2009). Such holistic understanding of flood occurrences is especially useful in safety assessments of engineering structure and for better floodplain planning.

Conclusions

Flood studies have grown as a multidisciplinary science. Sustainable flood management practices are moving towards an ecosystem based approach, which incorporates the process based on understanding from

different disciplines. Estimation of flood magnitude and sediment flux dynamics is still a challenging task in hydrological studies. Depending on the data and model availability, a broad spectrum of approaches—ranging from highly empirical to physically based models—have been developed over the past century. In all the approaches discussed in this article, an underlying concern is that of attendant uncertainty and the reliability of resulting flood estimates. New technological advances in the fields of remote sensing and sensor based digital measurement are enabling every greater use of multiple data streams in flood estimation studies. Concurrently, improvements in the underlying physics of hydrologic phenomena are also contributing to more accurate model representations and a reduction in uncertainties. It is generally considered that river management strategies need to be basin specific, as: (a) rivers show great diversity in the flood-producing mechanisms and meteorological characteristics, and (b) flood management and infrastructure design occurs at particular scales. From a hydrologic science standpoint, flood hazard in any area, which may occur through combinations of various meteorological, hydrological, glacial and geomorphic processes, contribute to their uniqueness. Knowledge from multiple disciplines has an important role in shaping and informing flood studies. All of this is likely to pose multiple challenges and opportunities for a fruitful hydrologic research in this century, one where changing climatic and land-use baselines will require a deeper understanding of time-varying causative processes.

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