Chapter 2

Review of Literature

2.1 Introduction

In the last two decades investigation on bedrock channels and fluvial erosion has seen a noteworthy increase in interest. It was accepted that these channels play a crucial role in the development of the entire landscape. They set the base-level for hillslope response, control the relief of a region and are major agents of sediment transport (Whipple, 2004). An idea of a dynamic combination between climate-driven erosion and tectonics received wide interest in the nineteen nineties (Molnar and England, 1990; Willett, 1999), and triggered exhaustive research in bedrock channels and fluvial erosion. Fluvial geomorphologists have recognized importance of bedrock channels because they behave quite differently than alluvial channels, for which river research had focused on for many decades (Tinkler and Wohl, 1998; Richardson and Carling, 2006; Wohl and Merritts, 2001; Turowski, 2011).

Previous work on bedrock channels has been scanty and frequently focused on small-scale features of rock surface such as potholes or upon the single catastrophic floods (Tinker and Wohl, 1998). Bedrock channels came into the focus of geomorphic research in the recent decades. Despite new insights, many research questions remain open. The subject of bedrock channels has a large but scattered literature dating back over a century. The world distribution of studies in bedrock channels has been shown by Tinker and Wohl (1998). Their map indicates that most of the bedrock channel investigations are from USA and Australia. Studies of bedrock channels from rest of the world are very limited. Like other countries of the world, the research on bedrock channels in India is also inadequate though the bedrock channels are present in many areas. However, some work on the bedrock channels of the Narmada and Tapi Rivers have been carried out (Kale et al., 1994; Rajaguru et al., 1995; Ely et al., 1996; Kale et al., 1996; Kale and Gadgil, 1997; Baker and Kale, 1998; Hire, 2000; Kale et al., 2003; Kale and Hire, 2004; Kale, 2005).

There are three approaches to study bedrock channels namely basin-scale, reach-scale and experimental (Tinkler and Wohl, 1998).
Basin-scale approach, which generally focus on the evolution of channel longitudinal profile at time scales of centuries or longer (Weissel and Seidl, 1998). Studies may be field-based (Merritts et al., 1994; Pazzaglia et al., 1998). Some of the field-based studies are oriented towards computer modelling of basin evolution (Howard, 1987; Howard et al., 1994; Seidl et al., 1997; Howard, 1998; Sklar and Dietrich, 1998). However, the general focus is on long-term rates of profile lowering and the development of an erosion rate law (Tinkler and Wohl, 1998; Wohl and Merritts, 2001).

The reach-scale studies are associated with the processes of erosion and deposition, as these factors have influence on channel morphology for few square meters to several widths at spatial scale. At time scales, such studies include observable processes for days to decades (e.g. Toda, 1994; Tinkler and Parish, 1998; Tinkler and Wohl, 1998; Hancock et al., 1998). An indirect approach of inferring processes from form, with the aid of palaeostage indicators and hydraulic simulation programs was adopted by O’Connor et al. (1986); Baker and Pickup (1987); Whol (1992a, b); Whol et al. (1993). Besides this, reach-scale studies include sophisticated mathematical flow modeling in bedrock channels (Miller and Cluer, 1998).

Experimental studies have used a variety of cohesive substrates to simulate either erosion of a specific feature, for instance, potholes (Alexander, 1932; Angeby, 1951) or knickpoints, (Holland and Pickup, 1976; Gardner, 1983) or general channel erosion under different conditions (Shepherd and Schumm, 1974; Wohl and Ikeda, 1997).

The literature review for the present work has been carried out on the basis of following points to match the objectives of the study and subsequent chapterization.

2.2 Channel morphological features

Straight channels, in fact, rarely exist or almost fictional among natural channels. However, exceptionally short sections or reaches of the channel are possibly straight. Nevertheless, in general, the reaches which are straight for distance more than ten times the channel width are rare in nature (Leopold and Wolman, 1957).
Meandering channels is a vast research area, covering a broad range of time and space scales, environmental dominions, and theoretical and practical approaches (Güneralp et al., 2012). A widespread review of the huge literature on alluvial river meanders is much more than that of bedrock river meanders. The research on alluvial meandering rivers had amplified to such extent by the latter part of the 20th century that in 1983 the conference namely Rivers' 83, sponsored by the American Society of Civil Engineers (ASCE), focused absolutely on meandering rivers (Elliot, 1984). Progress in research on river meandering during the 90s and at the commencement of the 21st century have focused exclusively on numerous topics such as (i) channel planform evolution; (ii) field-based or empirical research on the interactions of linking flow structure and bed morphology; (iii) research stand on experiment or laboratory on flow and sediment transport in winding channels and (iv) numerical modelling of meander morphodynamics (Güneralp et al., 2012). To analyze river-meander patterns thoroughly, two general approaches i.e. traditional approach and series approach has been given by Williams (1986). The traditional approach presumes and highlights on fundamental regularity of meander geometry (Inglis, 1947; Leopold and Wolman, 1960). However, by a thorough study of the meander trace, the series approach emphasize on the varying degrees of irregularity or quasi-randomness (Ferguson, 1976).

According to Güneralp et al. (2012) Studies on meandering river channels has mainly endeavored to explain the morphodynamic development of meandering rivers controlled by the interactions among water flow, sediment transport, channel planform and bed morphology. Güneralp et al. (2012) thoroughly introduced special issue of Geomorphology i.e. advances and challenges in meandering channels research, however, the subject matter of bedrock meanders remain ignored. Marked differences in dimensions of the alluvial and bedrock meanders have been noted by previous workers. Flows effective in meander formation may have a much larger recurrence interval than those of meandering alluvial channels (Tinkler, 1971). When meanders are observed in bedrock rivers they are classically interpreted as an antecedent feature. However, Hovius and Stark (2001) have found widespread evidence in Taiwan that this is not always the case and that instead bedrock rivers may actively meander.
In accordance with Leopold and Wolman (1960), the meander geometry has been the object of widespread statistical study and examples of that were given by Jefferson (1902), Inglis (1937; 1949), Bates (1939), and Leopold and Wolman (1957). Brice (1964) applied the Sinuosity index (Si) to differentiate straight river channels from sinuous and meandering river channels. Si ranges between 1.3 to one and four to one for the large majority of meandering rivers (Leopold and Langbean, 1966). A constant ratio between the meander wavelength and the radius of curvature has been noticed by Leopold and Langbean (1966) in a given series of meanders for the alluvial rivers. The appearance of regularity in meander depends in part on how constant this ratio is. The striking uniformity in dimensions of meanders in different physiographic settings is the result of certain geometric proportions appear common to all. For example, a nearly constant ratio of radius of curvature (Rc) to channel width (W) has been noticed by Leopold and Wolman (1960) and Williams (1986). The three empirical equations, for instance, meander wavelength (λ) and channel width (W), meander wavelength (λ) and mean radius of curvature (Rcm) and amplitude (A) and channel width (W) have been given by Leopold and Wolman (1960) to show remarkable relationship between meander wavelength, channel width and radius of curvature for alluvial rivers.

Bedrock river channels mainly flow through single path. However, several workers have described multiple-flow path channels incised into bedrock in variety of environments. Such channels are known both bedrock anastomoising channels and scabland topography or scablands (Wohl, 1998). Heritage et al., (2000) used term bedrock anastomoising for multi-thread channels in bedrock. Garner (1974) follows Bretz (1923) and define anastomoising channel as “an erosionally developed network of channels in which the insular flow obstructions represent relict topographic highs and often consist of bedrock”. According to Wohl (1998) anomalous development of multiple channels in bedrock are attributed to one or more of the three processes namely (i) inadequate channel capacity; (ii) localized uplift along the channel and (iii) preferential erosion along lines of weakness, such as joints and fractures. Kale et al. (1996) and Gupta et al. (1999) described anastomoising channels along the Narmada River of India. Kale and Shingade (1987) illustrated the formation of multiple bedrock channels along the Indrayani River by coalescence of grooves and potholes along joints in basalt bedrock.
The form of a channel is primarily a function of (i) the discharge and its variations; (ii) the texture and quantity of sediments passing through the section and (iii) the nature of the bed and bank material (Leopold et al., 1964; Schumm, 1977; Petts and Foster, 1985). In accordance with Leopold and Maddock (1953) and Maddock (1976), the alluvial channels having sedimentary particles at banks and beds are mobile in nature; these channels are self-generated through the self-governing adjustment of the morphological variables encompassing their hydraulic geometry. Nevertheless, such channels may experience infrequent high-magnitude events, their morphology have a tendency to recover to the original dimensions at varying rates depending on the series of floods and other climatic-geomorphic causes (Wolman and Gerson, 1978). Baker and Kale (1998) considered high-energy processes that are less studied by previous workers and which occur during severe floods in highly resistant bedrock channel situations in their work. According to Schumm (1977), although, alluvial channel patterns can be systematically linked to sediment types (which forms channel banks), to sediment loads, and to moderate flood characteristics, the lofty thresholds for channel alteration in bedrock rivers (Baker, 1977) bring about a diverse range of channel types and patterns (Shepherd, 1979; Wohl, 1998).

The form ratio is the ratio of channel width and depth. It is primary index of channel shape and is related to the sediment transport and boundary resistance (Schumm, 1960). Generally two groups of aspects are considered to describe channel cross sectional form – i) channel size and ii) channel shape. Perimeter lithology is an important factor to determine channel shape. Rosgen (1994) used boundary composition as one of the basic criteria to classify river channels. It is the most elaborate classification schemes yet developed. He produced 41 channel types on the basis of boundary composition. The impact of floods depends not so much on the volume of water as on the energy exerted by it. The adjustments in the width-depth ratio and hydraulic variables with discharge have been shown to very useful concepts in evaluating the potential of flows to be geomorphologically effective (Kale et al., 1994; Gupta, 1995a). Montgomery and Gran (2001) derived a fundamental set of relationships between drainage area (A) (a surrogate for discharge) and channel width (W) for alluvial and bedrock rivers. In view of their research work for alluvial and bedrock rivers, an attempt has been made to highlight how classic concepts and
As stated by Wohl (1998), the existence of exposed bedrock along a channel entails only limited and localized deposition along the channel. As a result, the morphology of many bedrock channels is dominated by erosional processes such as corrosion, solution, corrasion or abrasion, cavitation, etc. Wohl (1998) has classified bedrock erosional landforms at various spatial scales, for instance, micro-scale (mm to cm), meso-scale (cm to m) and macro-scale (m to km) (Table 2.1). The mainstreams of studies have focused on meso-scale erosional forms which are largely descriptive and empirical, as several researchers have executed experiments to compute erosive process (Wohl, 1998). As demonstrated by Blank (1958), the preliminary approach to meso-scale erosional features was 1) to illustrate a particular channel reach which have potholes or longitudinal grooves 2) to infer the erosive processes that produce these features 3) furthermore, to describe the position of the erosional features in relation to lithology, gradient, or other characteristics of reach-length exclusive to that site. Baker (1973) has developed another approach to study meso-scale erosional forms using paleostage indicators in combination with step-backwater hydraulic models to course a flood discharge along a reach of channel. A second fundamental approach towards erosive processes and channel form has been given by Wohl (1998), which focuses on modelling macro-scale channel evolution. The third approach, micro-scale studies mainly concentrate on longitudinal profile as a sign of the channel's capability to incise, or on development of channel network (Wohl, 1998).

<table>
<thead>
<tr>
<th>Scale</th>
<th>Erosional characteristics</th>
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<tbody>
<tr>
<td>Micro-scale (mm to cm)</td>
<td>Abrasion, flaking, or plucking of individual grains or small pieces of rock</td>
</tr>
<tr>
<td>Meso-scale (cm to m)</td>
<td>Selective erosion of portions of the channel boundaries across a cross section or along a reach: produces potholes, longitudinal grooves, knickpoints, undulating walls, inner channel, pool-riffle or step-pool sequences</td>
</tr>
<tr>
<td>Macro-scale (m to km)</td>
<td>Reach-to basin-scale channel morphologies in planform (meandering, downstream alternations in width and gradient), and in gradient</td>
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Table 2.1 Scales of erosional features

After Wohl (1998)
Richardson and Carling (2005) define potholes as being essentially round (in plan view), deep depressions, which are, or can be expected to be, eroded by vortices with approximately vertical axes by mechanisms other than plucking. According to them this is the most comprehensive definition because it takes into account both the process of formation and the morphological aspect. Potholes are meso-scale erosional landforms (cm to m scale) and are found in a variety of climates, lithologies and channel types (Wohl, 1998). These are formed by fluvial erosional processes like corrasion, abrasion and cavitation (Wohl, 1998; Kale and Gupta, 2001; Sengupta and Kale, 2011). Potholes are significant component of channel incision and, in turn, lead to distinctive form of bedrock channels (Kale and Shingade, 1987; Springer et al., 2006). Kale and Shingade (1987) stated that potholes are created as tiny depressions in the beginning. These small depressions trap more sediments and water, enhancing erosion through whirling movement of water and sediment. Later, the pit is abraded, deepened and widened to a typical pothole.

Longitudinal grooves parallel to flow result from longitudinal vortices and turbulent vortices during high-magnitude flood flows (Wohl, 1993). Longitudinal groove are associated with zones of enhanced erosion.

Shepherd and Schumm (1974) hypothesized that inner channels are formed by the high flow stresses generated during large floods in the steep reaches of the bedrock rivers. Further, investigations by Baker (1988), Wohl (1992a) and Wohl and Ikeda (1997) supported their explanation. Inner channels play significant role for incision of channels into resistant substrate, which maximize shear stress and unit stream power for a given stage.

Channel gradient is another important channel morphologic variable dictating the flood power and impact. Gradient is controlled by the lithology of a basin (Hack, 1973). Generally, areas of resistant bedrock exhibit steeper channel gradient. Such high-gradient channel reaches are efficient in terms of erosion and transportation of material during large floods. Channel boundary shear stress and unit stream power are vital parameters in determining geomorphic response. These parameters are greatly influenced by channel slope (Baker and Costa, 1987).
In open bedrock channels, Richardson and Carling (2005) projected a comprehensive explanation and an organized taxonomy for the typology of sculpted shapes. The dimensions of such sculpted marks are varied, and an extensive range of approaches have been applied to study them thoroughly (Velázquez et al., 2016). Wohl and Merritt (2001) and Wohl and Achyuthan (2002) have carried comprehensive reviews in this perspective, with regard to factors such as 1) hydraulic driving force, 2) physical resistances of the substrate and 3) morphological features.

As opined by Leopold and Maddock (1953); Schumm (1977) and Schumm and Winkley (1994), alluvial channels shape their channel in bed and bank sediment that can readily entrain and transported by rivers for a broad range of flows. Consequently, according to Leopold and Maddock (1953) and Hey (1982), these channels regulate their geometry, pattern, and gradient to frequent flows of low to moderate magnitude that transport the most sediment and that are close to bankfull conditions. Cenderelli and Cluer (1998) stated that, for large part, alluvial channels supply abundant sediment due to availability of sediment at the channel bottom and banks and because of the capability of the stream to readily entrain and transport this sediment. On the contrary, in resistant-boundary channels and valleys, coarse-grained deposition mainly cobbles and boulders and fine-grained deposition essentially sand and fine pebbles, usually remain in association with infrequent and extreme floods (Cenderelli and Cluer, 1998). Infrequent and extreme floods produce flows that are "out-of-bank" and extend across the whole valley bottom in the resistant-boundary valleys. Such flows are responsible for widespread geomorphic activities along the course of the flow (Cenderelli and Cluer, 1998). The process of erosion primarily takes place in constricted reaches where valley side slopes are embraced of coarse and unconsolidated sediment. Quite the reverse, according to Martini (1977); Baker (1978, 1984); Church and Jones (1982); Carling (1987, 1989, 1995); Wohl (1992) and O'Connor (1993), in general, deposition of coarse-grained material occurs at particular locations for instance (i) where the channel and/or valley widen; (ii) upstream and downstream of obstructions and; (iii) along the margins of channel bends. In resistant-boundary channels, the supply of coarse-sediment is spatially irregular (Baker,1988) and forcefully controlled by the factors like (i) the availability of sediment in constricted reaches; (ii) the capability of the flow to entrain and transport the sediment and; (iii) the number and closeness of depositional areas to the sediment source.
The situation of fine-grained sediment transport remains different in resistant-boundary channels. During frequent low to moderate flows, amount of fine-grained sediment are entrained transported and deposited (Cenderelli and Cluer, 1998). According to Schmidt (1990) and Cluer (1995) such phenomenon normally takes place immediately upstream and downstream of constricted reaches and beside the channel margins where flow recirculates. Two case studies, specifically (i) coarse-grained deposition in the Mt. Everest Region of Nepal due to infrequent and extreme flood and (ii) fine-grained deposition along the Colorado River in and near the Grand Canyon, U.S.A. owing to low to moderate floods have been examined by Cenderelli and Cluer (1998). The above-mentioned unique case studies assess the significance of sediment supply in influencing coarse as well as fine-grained deposition in resistant-boundary channels.

Substantial depositional features located at sudden expansions immediately downstream of constricted reaches are called as expansion bar or boulder delta (Baker, 1978, 1984; Elfstorm, 1987; O’Conner, 1993). Surfaces of this bar consist of multiple linear and lenticular bars separated by shallow channels. The bars are outcome of rapid reduction of flow energy and flow competence. Longitudinal bars are narrow, linear to curvilinear, elongated along the axis of channel that formed at local flow expansions along the valley margins. They are extended in the direction of flow. Sometimes, longitudinal bars form in the centre of the channel, typically where the channel is relatively wide. Longitudinal bars tend to taper off in a downstream direction (Robert, 2003). Point bars are plainly an accumulation of deposited material along the inner margins of channel bends where flow energy is reduced and secondary currents transport sediment from the main channel to this reduced flow region (Knighton, 1984; Dietrich and Smith, 1984; de Jong and Ergenzinger, 1995).

2.3 Erosional processes and sediment transport

The bedrock channels are supply limited (since the transport capacity of flow is greater than the supply of sediment) and the morphology of bedrock channels is dominated by the processes of erosion. As per Wohl (1998), the bedrock substrate is dominantly eroded by processes of (i) corrosion, or chemical weathering and solution; (ii) corrasion, or abrasion by sediment in transport along the channel and (iii) cavitation and other hydrodynamic forces associated with flow turbulence. Knighton
(1998) described that under conditions of very high flow velocity, sudden changes in pressure can cause the formation and implosion of vapour bubbles. The shock waves generated by implosion that weaken the bed by the process of cavitation. This effect is mainly caused by the abrupt collapse of vapour pockets within the flow. The cause behind the process of cavitation may be flow separation induced by joints, bedding planes, or other surface irregularities in bedrock (Barnes, 1956). The erosive potential of this process can be phenomenal, under sustained high flow (Eckley and Hinchliff, 1986). Erosional features such as flute marks, polished rock surfaces and pot holes are indicators of intense bedrock scouring, resulting from cavitating flow conditions (Baker, 1988; Kale et al., 1993b; Kale et al., 1994). Embleton and King (1968) opined that the process of cavitation may causes quarrying of the scablands. One of the crucial causes behind entrainment of particles is fluid stressing or shear detachment in which flowing water exerts a shear force upon the bed it overflows. It is distinguished that the sediment transport rates and sediment entrainment are driven by excess shear stress over a threshold value, and a similar mechanism can be predicted for bedrock erosion (Turowski, 2012). However, according to Howard (1998) this process is important only in weakly consolidated rocks and clays. The process of quarrying or plucking in bedrock erosion is the removal of loose blocks of rock from the bed of channels by drag and lift forces. It is dominant process of bedrock erosion (Hancock et al., 1998) rather than abrasion (Brez, 1924). Chatanantavet and Parker (2009) introduced the concept of macro-abrasion, this process is major cause for formation of blocks. In this process the existing cracks, joins and plane of weakness in the material are enlarged by the impact of particles until individual blocks are loosened. These loose blocks of rocks can be separated by shear detachment and entrained. Only minute study and introductory laboratory work has been available regarding this process (Dubinski and Wohl, 2005), though quarrying is thought to be significant in joined rocks (Brez, 1926; Hartshorn et al., 2002).

Impact erosion or abrasion is the process of scraping or wearing. Moving sediment particles in the flow may strike the bed and remove small fragments of the impacted rock material, it also drives crack proliferation and weakens the substrate (and thus prepares for plucking) (e.g., Bitter, 1963; Wilson, 2009). The most rapid rates of abrasion perhaps take place during turbulent floods, along channels of weakly resistant bedrock, accompany with large and moderately coarse suspended sediment
loads. This process can initiate the development of potholes and deep circular scour features, these formations affect the flow and accelerate the rate of erosion. The accumulated coarse material in pothole swirled around by the flow and it deepens as well as enlarges the potholes through drilling process into the channel bed. Over the time the bed elevation lowers due to coalesce of potholes. The other forms such as longitudinal grooves, knickpoints, and similar erosional features along the channel bed and walls are indicators of abrasion dominated erosion.

Hancock et al. (1998) for the first time documented and termed the process of hydraulic wedging. According to them hydraulic wedging is the process which loosens and prepares blocks for quarrying through wedging fragments of rocks/clasts into fractures and joints. Channel bed with wider or various preliminary cracks and bedload sediment is requisite for this process to function. The clasts ranges in size from fine sand to boulders and are wedged very tightly into joints of bedrock in such a way that removal of clasts necessitates noteworthy force (Hancock et al., 1998). Two possibilities have been given by Hancock et al. (1998) for encroachment of clasts into joints (i) the clasts are either emplaced forcefully by very high flow velocities (ii) clasts passively accepted into a crack that was momentarily widen while sediment was nearby, however, there is scarcity of data and no experiments to validate this process.

Knickpoint are sudden break or irregularity in the gradient along the long profile of a river. The migration of knickpoint is not erosional process as such, but interplay of several processes mounted by a channel-spanning bedform (Turowski, 2012). There are several viewpoints regarding formation of knickpoints. According to Whipple and Tucker (1999) knickpoints can be formed by changes in the climate or local tectonics. Korup (2006) stated that blocking of the channel by material of landslide may be responsible. Miller’s (1999) view proposes that lithologic contrasts may possibly form knickpoints. Chatanantavet and Parker (2009) opined that knicks can arise autogenically. The knickpoints play crucial role in the channel dynamics since these contribute information on base level through the channel network (Whipple and Tucker, 1999; 2002). Studies of Bishop and Goldrick (1992) described knickpoints for which pothole erosion at the lip is an important component of headward retreat. As flow approaches the lip of knickpoint, width decreases, but depth, velocity, and bottom shear stress increases (Gardner, 1983). As a result of this, the slope of the
incising channel reaches increases above the lip of knickpoints. Very few actual measurements exist for rates of bedrock knickpoint retreat.

Infrequent and large magnitude floods produce massive discharges into channels. The geomorphic works associated with such floods are variable, in some cases these floods generate minute geomorphic response (Costa, 1974), and in other cases magnificent effects are observed (Baker, 1977; Gupta, 1983). The geomorphic effectiveness of a flood, which relates to its ability to affect the form of the landscape (Wolman and Gerson, 1978), is commonly linked to flood power and the degree of turbulence (Baker and Costa, 1987; Wohl, 1993; Baker and Kale, 1998; Kale and Hire, 2004; Hire and Kale, 2006; Kale and Hire, 2007).

According to Baker and Costa (1987) the unit stream power and shear stress are measures of existing energy and have verified valuable notions in assessing the function of large floods in generating major channel changes along with movement of cobbles and boulders. The data collected by Baker and Costa (1987) for some large flash-floods as well as for some great historic and prehistoric floods exhibit that the power values coupled with such floods are a number of orders of magnitude upper than those produced in alluvial rivers. Furthermore, according to Baker and Costa (1987) and Baker (1988), these investigations specify that, very high values of actual energy consequence in cavitational erosion and erosionaly efficient macroturbulence. Therefore, magnificent changes, yet in the resistant channel boundaries, have been credited to such high-energy flood conditions.

Tinkler and Wohl (1998) opined that, flows in bedrock systems usually have highly aerated and turbulent flow structure and in general show greater velocities and shear stresses than those in alluvial reaches, in addition, substantial sections of the flow are critical (Fr = 1 or close to 1) or supercritical (Fr > 1). However, flow remains unsteady and gradually varied i.e. subcritical for several locations. Supercritical flow is more common in bedrock channels and can be sustained for longer period of time (Baker and Costa, 1987). This flow move rapidly and efficiently through the channel due to less intensive turbulent mixing and less deviation from the main downstream direction of flow. Supercritical flow may overshoot tight bends and can also be highly erosive (Kay, 1998). Supercritical flow in general occurs when increase in channel slope increases the flow velocity, resulting in a reduction in depth (the hydraulic
Standing waves form in critical flow when the Froude number is 1 or close to 1. Normally standing waves form over deforming or non-deforming boundaries, however, form more easily over rough boundaries (Alexander, 2008). The frequent occurrence of unbroken standing waves is caused by the presence of boulders and cobbles in the channel bottom, as well as by a considerable increase in gradient of the channel in some of its parts, in addition, broken standing waves are formed in channel i.e. turbulent flow with foamy water and breaking wave crests, generally appears like ‘white water’ (Wiejaczka et al., 2014). Instability occurs in channel if Froude number exceeds a critical value (i.e. 1), and it gives rise to supercritical flow. Whenever the Froude number is in excess of 1.6, roll waves or slug flow appears (Hjalmarson and Phillips, 1997). In general, these waves more probably to be initiated on wide, shallow, steep systems, and over gravel surfaces (Tinkler and Wohl, 1998). Roll waves then travel downstream, and they sustain for periods of hours during peak flow. These waves appear like “walls of water” in channel (reported by eyewitness) and are almost certainly roll waves (Tinkler and Wohl, 1998).

Reynolds Number is the ratio of inertial and viscous forces acting on a body of fluid, it is dimensionless coefficient, Re number measures the degree of turbulence, or random changes in flow direction and/or velocity superimposed on the main downstream movement of water (Richards, 2004).

The process critical velocity for inception of cavitation (Vc) can occur only for certain critical conditions. According to Hjulstrom (1935) the minimum velocity necessary for cavitation to take place in river is about 12 m/s. However, this figure is applicable for relatively shallow and swift streams (Baker, 1973). The critical velocity for inception of cavitation in m/s is given by Baker (1973) and Baker and Costa (1987).

The upstream migration of knickpoints has been recognized as significant means of bedrock channel lowering, however, little is known about the mechanisms that control the shapes and migrations of knickpoints (Miller, 1991; Seidl and Dietrich, 1992; Seidl, 1993; Whipple et al., 2000 (a, b); Zaprowski et al., 2001). According to Baker (1988); Wohl (1992, 1998 and 2000) and Wohl and Ikeda (1997) headward migration of a knickpoint through resistant substrate can leave behind a deep and narrow gorge, it reflects the erosional resistance of the channel boundaries, and maximizes the shear stress and stream power per unit area of a given discharge and channel gradient.
Several equations and models have been developed by researchers to predict channel incision of a river into its bed. However, the comprehensive and most commonly used stream power erosion model (SPEM) is of great use since there are few variables and can be measured against topographical data (Howard and Kerby, 1983; Skylar and Dietrich, 2001). It is argued by Howard and Kerby (1983) that the Stream Power Law/model (SPEM) is most applicable because it is related to physics of erosion. The family of stream power models is based on the principle that bedrock channel incision rate can be estimated by a power law function of mean bed shear stress or stream power (Howard and Kerby, 1983; Howard et al., 1994; Whipple et al., 2000a; Kobor and Roering, 2004; Whipple 2004). Stock and Montgomery (1999) have applied stream power erosion model for Kaulaula and Waipao Rivers, which flow through basalt lithology.

The investigations of Leopold and Maddock (1953); Schumm (1977) and Schumm and Winkely (1994) reveals that alluvial channels shape their form in bed and bank sediment that the stream can readily entrain and transport for a wide range of flows. In contrary, according to Baker (1988), the resistant-boundary channels are supply-limited, coarse sediment entrainment and deposition is usually associated with infrequent and extreme floods, since, energy required to transport a particle of sediment increases with particle size.

2.4 Role of lithology and tectonics

The morphology of channel is predominantly function of fluvial forces applied and bedrock resistance offered. The rock resistance to flow dynamics noticeably varies with respect to lithological considerations. In accordance with Goudie (2004) the erodibility of rocks relies on the lithology which strongly controls the erosional processes. In this standpoint, rocks are frequently referred to as ‘hard’ or ‘resistant’ or ‘weak’ and ‘non-resistant’ to erosional processes. In order to find out effects of rock strength/role of lithology in shaping the landforms, weathering phenomena and relative dating the Schmidt hammer (SH) has now been adopted by Geomorphologists (e.g. Ericson, 2004). The instrument was devised by E. Schmidt in 1948. Primarily Schmidt hammer has been used by civil engineers to test the strength of concrete. However, from last few decades, Geomorphologists and Geologists have started using SH to estimate the strength of rocks for numerous reasons (Goudie, 2006). SH
measures the distance of rebound of controlled impact on a surface and represents a relative measure of surface hardness or strength (Goudie, 2006). Yasar and Erdogan (2004) stated that, several studies on the investigation of efficacy of the Schmidt hammer test on diverse rock types have been made by numerous investigators. Goudie (2004) used the Schmidt hammer rebound values (N) to estimate the Rock Mass Strength (RMS) i.e. the specific properties of the rock mass that control its strength and subsequent slope stability.

The commonly-used geomorphic indices of active tectonics (GAT) have been developed as basic investigation tools to assess the relationship between tectonics and basin morphology on the regional or basin scale and to identify areas experiencing tectonic deformation (Table 2.2) (Bull and McFadden, 1977; Keller, 1986; Keller and Pinter, 1996; Burbank and Anderson, 2001; Della Seta et al., 2004; Kale and Shejwalkar, 2008). According to Keller and Pinter (1996) the results of several geomorphic indices can be combined to provide an assessment of a relative degree of tectonic activity in an area. Geomorphic indices can be obtained easily from topographic maps or aerial photos (Strahler, 1952). The analysis of Kale and Shejwalkar (2008) and Troiani and Della Seta (2008) states that, in recent decades, the increasing usefulness of GIS software has made it possible to undertake quick and detailed processing of data. Recently, in morphotectonic studies, traditional geomorphic analysis has been integrated with morphometric analysis of landforms and with geostatistical topographic analysis (Keller et al., 1982; Mayer, 1990; Cox, 1994; Merritts et al., 1994; Lupia et al., 1995; Lupia et al., 2001; Currado and Fredi, 2000; Pike, 2002; Della Seta, 2004; Della Seta et al., 2004; Kale and Shejwalkar, 2008; Troiani and Della Seta, 2008; Figueroa and Knott, 2010; Dehbozorgi et al., 2010; Font et al., 2010; Jayappa et al., 2012). Geomorphic indices appropriate to fluvial systems in different regions and of varying size (Strahler, 1958), associate with independently derived uplift rates (Rockwell et al., 1985; Merritts and Vincent, 1989; Kirby and Whipple, 2001) and are applicable to a variety of tectonic settings where topography is being changed (Bull and McFadden, 1977; Wells et al., 1988; Azor et al., 2002; Figueroa and Knott, 2010).
Table 2.2 Geomorphic indices of active tectonics (GAT) and their calculations

<table>
<thead>
<tr>
<th>Sr. No.</th>
<th>Index</th>
<th>Formula</th>
<th>Variables</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Hypsometric Integral (HI)</td>
<td>HI = (Em – Emin)/(Emax – Emin)</td>
<td>Em = mean elevation</td>
<td>Bull and McFadden (1977)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Emax = maximum elevation</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>Emin = minimum elevation</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Valley width-height Ratio (Vf)</td>
<td>Vf = Vfw/(((Eld-Esc)+(Erd-Esc))/2)</td>
<td>Vfw = width of valley floor</td>
<td>Bull and McFadden (1977)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Eld = elevation of the left</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>valley divide</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Erd = elevation of the right</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>valley divide</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Esc = elevation of the valley</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Floor</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Asymmetry Factor (AF)</td>
<td>AF = 100(Ar/At)</td>
<td>Ar = area of the basin to the</td>
<td>Keller and Pinter (1996)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>right of the trunk stream</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>AT = total area of the drainage</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>basin</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Stream Length-Gradient Index (SL)</td>
<td>SL = (H1 – H2)/(ln L2 – lnL1)</td>
<td>H1 and H2 are the elevations</td>
<td>Hack (1973)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>of each end of a given reach</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>L1 and L2 are the distances</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>from each end of the reach to</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>the source</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Basin elongation ratio (Re)</td>
<td>Re = (2√A :√x)/LB</td>
<td>A = basin area</td>
<td>Bull and McFadden (1977)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>LB = length of the basin</td>
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</tr>
</tbody>
</table>

Hypsometric analysis (or area-altitude analysis) is the study of the distribution of horizontal cross-sectional area of a landmass with respect to elevation (Strahler, 1952). Classically, hypsometric analysis has been used to differentiate between erosional landforms at different stages during their evolution (Strahler, 1952; Schumm, 1956; Strahler, 1964). Hypsometric Integral (HI) is a relief variable which is widely used to measure the degree of fluvial landscape erosion and describes the distribution of elevations across the drainage basin area (Strahler, 1952). Hypsometric Curve (HC) of a catchment represents the relative area below (or above) a given altitude (Strahler, 1952). Hypsometric curve can also be used to infer the stage of development of the drainage network and can be considered as a powerful tool to differentiate between tectonically active and inactive areas (Keller and Pinter, 1996).

Vf is basically an index of the form or shape of the valley cross-section. The Vf ratio is a good measure that indicates whether the river is actively downcutting and incising (Bull and McFadden, 1977). The Vf index reflects the difference between V-shaped valleys that are down cut in response to active uplift (low values of Vf) and broad-
floored valleys that are eroding laterally into adjacent hill slopes in response to base level stability (high values of $V_f$) (Bull, 1978).

The effect of tectonics on the drainage pattern is also reflected by the asymmetry of drainage basins (Molin et al., 2004). AF can be used to evaluate tectonic tilting at the regional or basin scale (Hare and Gardner, 1985; Keller and Pinter, 2002).

The Stream Length-Gradient Index (SL Index) is considered as one of the quantitative geomorphic parameters incorporated in morphotectonic analysis (Hack, 1973). In tectonically active regions, and/or at the basin scale of investigation, the SL Index can be a useful tool to detect tectonic displacements (Keller and Pinter, 1996; Chen et al., 2003; Zovoili et al., 2004; Kale and Shejwalkar, 2008, Troiani and Della Seta, 2008; Monteiro et al., 2010). Nevertheless, the effectiveness of the parameter in detecting local active structures has not been confirmed for small catchments and/or in regions where tectonic activity is subtle (Chen et al., 2003 and references therein; Verrios et al., 2004; Troiani and Della Seta, 2008). In small river basins the contribution of the lithological effect to anomalous values of the SL Index seems indistinguishable from the tectonic one (Troiani and Della Seta, 2008). However, in spite of all the difficulties, SL index has been widely used as a proxy to identify areas of anomalous uplift within a landscape (Kale and Shejwalkar, 2008).

Basin elongation ratio (Re) is an areal morphometric variable that quantitatively describes the planimetric shape of a basin and, thus, indirectly provides information about the degree of maturity of the basin landscape (Kale and Shejwalkar, 2008). Basins draining tectonically active areas are more elongated and become more circular with the ending of uplift (Bull and McFadden, 1977). Elongated basin shapes are also associated with high local relief and steep valley slopes (Molin et al., 2004).

An idea of a dynamic combination between climate-driven erosion and tectonics received wide interest in the nineteen nineties (Molnar and England, 1990; Willett, 1999), and triggered exhaustive research in bedrock channels and fluvial erosion. River incision into bedrock is a significant erosion process that has an impact on the rate of landscape response to changes in rock uplift rate and climate (Howard et al., 1994). Considerable attention has been given to morphology of bedrock channels and dynamics and to fluvial erosional processes in recent years (e.g. Turowski et al., 2008;
Howard, 1994; Wohl et al., 1994; Tinkler and Wohl, 1998a; Stock and Montgomery, 1999; Whipple et al., 2000a, b; Wohl and Merritt, 2001; Finnegan et al., 2005; Stark, 2006; Wobus et al., 2006a, b; Whittaker et al., 2007).

2.5 Flood hydrometeorology, flood hydrology and flood geomorphology

(i) Flood hydrometeorology

As stated by Wohl (1992b) and Gupta (1995a), floods play a dominant role in shaping the river channel and the landscape in certain hydro-geomorphic environments, such as the seasonal tropics. In accordance with Baker’s (1988) view, flood geomorphology is concerned with the processes, forms, effects, and causes of floods. The frequency and hydraulic properties of the high flows play foremost important role to shape the channel and to carry the sediment. Infrequent large floods that occur at an interval of several decades are associated with much higher levels of power expenditure and thus are capable of producing major channel changes and movement of coarse sediments (Baker and Kale, 1998). The major reason of occurrence of floods was given by Hirschboeck (1991), according to her, floods are produced due to extraordinary synoptic situations that deliver more precipitation to a drainage basin than that can be readily stored or absorbed in the basin. In the humid and seasonal tropics, large floods are mostly associated with high-magnitude rainfall caused by synoptic events ranging in force from lows to cyclones (Gupta, 1988; 1995a). Almost 80-90% of the annual rain over most parts of the country falls during the period of summer monsoon season i.e. from June to September due to the monsoon circulation. During this period cyclonic disturbances from the Bay of Bengal and the Arabian Sea produce widespread and heavy rainfall which often causes severe floods in Indian rivers (Rakhecha, 2002). As per earlier and recent studies of the synoptic situations associated with the rainstorms, flood-generating rainstorms are connected with (Abbi and Jain, 1971; Ramaswamy, 1985) -

(1) Bay of Bengal depressions moving westwards
(2) General active monsoon conditions over Madhya Pradesh and Gujarat
(3) Land depressions moving westwards
It is well known that the monsoon rainfall of the same region goes through variations from one year to another. However, departures of rainfall from its long-term mean in any two years are not same (Gadgil, 2002). The year to year fluctuations in rainfall of the region cause complexity in recognition of the direction of change in the rainfall. Thus, some effective statistical methods are to be applied to identify the nature of long-term variability in monsoon rainfall. The frequently used method to study the variability of rainfall is Normalized Accumulated Departure from mean (NADM). Successive properties within long-term data can be merely resolved by NADM method (Riehl et al., 1979; Mooley and Parthasarathy, 1984; Probst and Tardy, 1987; Kale, 1999b). Consequently, the NADM plotting method has been used to emphasize the long-term variability by minimizing short-term fluctuations in the monsoon rainfall.

According to earlier studies, based on data for rivers from around the world, floods are not randomly distributed. Nevertheless, there is a tendency for periods of high and low floods to match with periods of high and low rainfall (Burn and Arnell, 1993; Chiew and McMohan, 1993; Kale, 1999b).

El Niño is a phenomenon in which episodic warming of the ocean occurs in the central and eastern Pacific, and Southern Oscillation is the seesaw pattern of atmospheric pressure change that takes place between the eastern and western Pacific (Lutgens and Tarbuck, 1995). The ENSO event was discovered by Gilbert Walker with ascertaining the fact that Indian monsoon was not an isolated system but had strong teleconnections with the global climate (Kelkar, 2009). Numerous studies evaluated the probable linkages between the ENSO and Indian summer monsoon rainfall (ISMR) and revealed diverse aspects of the relationship between ISMR and ENSO (Khandekar, 1979; Sikka, 1980; Rasmusson and Carpenter, 1983; Shukla and Paolino, 1983; Ropelewski and Halpert, 1987; Kane, 1989; Simpson et al., 1993; Khole, 2004; Lutgens and Tarbuck, 2007; Ihara et al., 2007) after Walker (1924). These studies have deduced that usually, ISMR is inversely correlated with Sea Surface Temperature (SST) of the Pacific Ocean. Lutgens and Tarbuck (2007) observed that El Niño is indeed a part of the global circulation and influences the weather at great distances from the Pacific Ocean. Additionally, it is marked by an anomalous weather patterns. Indian monsoon is more prone to drought situations
during El Niño events, on the contrary, wet monsoons are more likely to prevail, during the La Niña events. (Krishnan and Sugi, 2003). Similar relationship between them has also been recognized by Saha et al. (2007).

Long-term changes in seasonal and annual rainfall have been evaluated using Mann-Kendall test by Hollander and Wolfe, 1973. In accordance with Subramanian et al. (1992), Mann-Kendall test is a powerful statistical technique for randomness against trend. Numerous workers have reported the use of Mann-Kendall test in trend analysis of meteorological parameters, particularly of rainfall. Krishnakumar et al. (2009) established the long-term changes in seasonal and annual rainfall over Kerala by Mann-Kendall trend test. Several workers have also applied this non-parametric method for quantifying the direction and magnitude of trends in the streamflow and rainfall records (Chiew and McMahon, 1993; Kale, 1998; Hire, 2000; Marengo, 1995; Probst and Tardy, 1987; Gunjal and Hire, 2007). Moreover, some other workers (Sahu (2004), Seetharam (2003), Lal et al. (1993) and Suresh et al. (1998)) used this test for detection of the nature of changes in the rainfall of the small regions or stations.

The important question to rainfall studies in India is whether the future is likely to see the condition of rainfall decreased, unchanged or exacerbated. Even though, it is complex to envisage the direction and magnitude of change, it is possible to approximate the percentage change required in the future data series before it can be considered to be statistically significant (Kale, 1998). Student’s t-test has been used by Chiew and McMahon (1993) and Marengo (1995) to find out the percentage change essential in the mean of the future rainfall data series prior it can be considered to be appreciably different from the historical gauge record.

(ii) Flood hydrology

It is obvious from the above discussion that monsoon regime plays an important role to determine the river regime conditions of the river under study. Nonetheless, the efficacy of discharge regime characteristics is inadequate for geomorphological purposes since it is based on monthly or ten-daily means. The extensive work on some large Indian rivers designate that the channel forms and processes are associated to very large, but relatively infrequent flood events (Goswami, 1985; Kale et al., 1994; Gupta, 1995a; Gupta et al., 1999).
According to Rostvedt et al. (1968) and Ward (1978), in a broad intellect, rise in the water level/stage or discharge that result in overtopping of natural or artificial banks of a stream is known as flood. In hydrology, a flood perhaps any relatively high water level or discharge above a pre-determined flood level or discharge magnitude (Ward, 1978). From the geomorphic standpoint, a flood has been defined as a high flow for which the stream channel is clearly inadequate transportation system and whose passage occupies at least the lower part of the valley flat (Gupta, 1988). Moreover, in India for meteorological functions, flood is recognized with reference to a danger level (DL). For instance, Ramaswamy (1985) considers a flood as ‘severe’ if the highest flow level is at least 2 m above the danger level. Large floods are often expressed in terms of return period or recurrence interval (100-yr, 500-yr or 1000-yr flood).

The definitions given above are usually accepted, however, they are neither applicable across the world nor to the study area since, the river under review is deeply incised in bedrock. Due to which even high flows are incapable to fill the entire channel, and overtopping is infrequent. At the same time, there is little uncertainty that significant positive departures from mean flows occur, and such flows cannot be treated as just high flows (Hire, 2000). Thus, there is a need to have an alternative definition of flood for the study area.

In general, for hydrologists and geomorphologists, the single maximum instantaneous discharge for every year of gauge record is of great interest. Consequently, the simplest and most suitable definition of flood for an incised river should be based on the statistical parameters such as mean and standard deviation of the annual peak discharges (Hire, 2000). According to Petts and Foster (1985), in fluvial geomorphology flows having a recurrence interval of 2.33 years to about 5 years are considered to be significant from the standpoint of geomorphic work (Petts and Foster, 1985). For the river under review none of the definitions given by earlier workers are applicable. However, Hire (2000) has defined floods on the basis of recurrence interval. According to him, a discharge having a recurrence interval of 2.33 years is the same to the mean annual peak discharge (Qm), whereas, a flow having a return period of 5 years is close to mean plus one standard deviation (Qm+1σ), which has a recurrence interval of 6.93 years. The above depictions given by Hire (2000)
seem to be appropriate for the Par River. Moreover, annual maximum series data are available for about 50 years for a gauging site on the river. Therefore, floods are defined as under;

- **Floods** ($Q_f$): all annual peak discharges above mean annual peak ($Q_m$), but below mean plus one standard deviation (i.e. $Q_m < Q_m + 1\sigma$).

- **Large floods** ($Q_{lf}$): all floods that exceeds mean plus one standard deviation ($> Q_m + 1\sigma$).

- **Peak on record** ($Q_{max}$): highest annual peak flood discharge on record during the gauge period. This is the highest $Q_{lf}$.

The measured instantaneous peak flood discharges encompass one of the most important datasets for hydrologists, engineers and geomorphologists. According to hydrologist annual peak discharge series or annual maximum series (AMS) is highest peak discharge recorded in each year for a series of years at a gauging site (Ward, 1978).

In general sense, in hydrology, stage discharge curve or rating curve is a graph of discharge versus stage/gauge for a given point on a stream, normally at gauging stations, where the stream discharge is measured across the stream channel with a flow meter. According to Giovanni (2008), the empirical as well as theoretical relationship exist between the water-surface stage (i.e. the water level) and the concurrent flow discharge in an open channel, this relationship is known as stage-discharge relation or rating curve, or just rating. Numerous measurements of stream discharge are made over a range of stream stages.

An evaluation of the effectiveness of flows depends much on the magnitude and frequency of the events than mean discharges. Magnitude-frequency analysis is one method that identifies the hydrological and geomorphological importance of these events quantitatively, particularly the frequency of flood events of various magnitudes (Chow, 1964; Leopold et al., 1964; Morisawa, 1968). In flood hydrology, flood frequency analysis (FFA) is a statistical measure and is considered to be an effective tool to interpret past records of gauge data in terms of future probabilities of occurrences (Mutreja, 1995). One of the most familiar means to indicate the probability of a flood event is to assign return period or recurrence interval to the
event. In general sense, the recurrence interval or return period is defined as “an annual maximum event having a return period or recurrence interval of T years, if its magnitude is equalled or exceeded once, on the average, every T years. The reciprocal of T is the exceedance probability of the events i.e. the probability that the event is equalled or exceeded in any one year” (Bedient and Huber, 1989). Floods are analyzed and explained in a probabilistic sense because of their inherent randomness.

There are numerous probability distributions that are used in flood hydrology. The most commonly used probability distributions are Lognormal, Gamma (Pearson type III), Log Pearson type III (LP III), Gumbel extreme value type I (GEVI) etc. Since the objective of the present study is not to find out the most appropriate probability distribution(s) for the river under study, but to estimate the recurrence interval of high flows, the FFA is mainly based on the GEVI distribution. GEVI probability distribution have been selected mainly on the basis of its applicability to the monsoon-dominated Indian rivers. On the basis of analysis of long records available for 92 gauging stations, Garde and Kothyari (1990), have proposed GEVI distribution for the AMS data from Indian catchments. Therefore, in order to understand the hydrological characteristics of floods in terms of size and frequency, the GEVI probability distribution has been applied to the AMS data.

E. J. Gumbel put forward the concept of extreme-value, in which he conceived that the largest daily discharge in a year was the upper extreme of the 365 daily flows, and this value of a year formed part of an extreme-value series (Petts and Foster, 1985).

According to Costa and O’Connor (1995), the maximum discharge is typically considered as a measure of the potential of flow to be an effective geomorphic agent. Large discharges indexed by area or recurrence interval are supposed to generate large forces to cause enduring changes in the channel and valley morphology. Nonetheless, given the hydro-climatic conditions there is an upper physical limit to the magnitude of floods that can be produced (Enzel et al., 1993), and consequently the maximum possible force that can be generated. Thus, to assess the potential of a region to produce a maximum possible peak discharge, regional envelope curves encompassing the maximum flood peaks experienced in a region have often been used to define the natural upper bounds to flood magnitudes (Enzel et al., 1993). This graphical and empirical approach is based on two assumptions: (1) that there are physical limits to
supply of precipitation to a basin (Enzel et al., 1993), and (2) the maximum flood per unit drainage area in one basin is likely to be experienced in nearby basins having similar hydro-geomorphic conditions (Mutreja, 1995).

(iii) Flood geomorphology

In flood geomorphology, the measurement and evaluation of the geomorphic effectiveness of flows of different magnitude has been one of the significant themes. Efficacy of events in shaping landforms is measured by the magnitude of flows, by the frequency with which they occur, and by the amount of suspended sediment they transport (Wolman and Miller, 1960). Recently, the potential of flood flows has also been assessed in terms of the channel boundary shear stress and stream power per unit boundary area (Baker and Costa, 1987), as well as the flood flow duration (Costa and O’Connor, 1995). Channel morphology is the dimensions of a river channel in cross section and in plan (Petts and Foster, 1985). The morphologic properties of a channel vary in different reaches throughout the course of a river and are governed by the factors given below (Morisawa, 1985):

- The interaction of the hydraulic of flow (velocity, discharge, roughness, and shear stress),
- The channel configuration at the reach and immediately upstream (width, depth, shape, slope and pattern),
- Sediment load entering the reach (caliber and amount), and
- The composition of bed and bank material.

Therefore, according to Morisawa (1985) the river channel morphology is an expression of equilibrium between stream power and the resistance of material comprising the channel perimeter. The appearance of a river can be divided into following categories.

- Channel size and shape: The size includes channel width (W), mean depth (d), cross sectional area (A_C), wetted perimeter (Wp), and hydraulic radius (R), and the shape of the channel is characterized by the width-depth (w/d) ratio.
- Channel slope or gradient
- Channel pattern: i.e. the form of the channel in two dimensions.
The form of the channel is a function of the energy available to erode or deposit materials of different caliber along the bed and banks at different flood stages. The channel form is function of bed and bank material texture, fluctuations in discharge, sediment load, the balance between aggradation and degradation and the resulting pattern, rates of bank erosion and deposition at any cross-section or along any reach (Fryirs and Brierley, 2009).

The literature review indicates that a few studies have highlighted on meteorological, hydrological and geomorphological aspects of floods on bedrock rivers.