CHAPTER 2

REVIEW OF LITERATURE

In this chapter, the methodologies used and the findings of various studies related to hydrologic modelling were reviewed, with the purpose of assessing their suitability for the study. The reviews have been presented as follows: rainfall and distribution (Divisions 2.2), hydrologic processes (Division 2.3), simulation and water balance modelling (Divisions 2.4 to 2.8), and scenario studies (Division 2.9).

2.1 INTRODUCTION

The hydrologic model for the study was conceptualized to simulate many hydrologic processes, water balance, and scenarios. Each part of the study depended much on methodologies and results of earlier studies. The literature directly related to process algorithms and results were presented in methodology and discussion chapters. The literature that have direct bearing on other aspects of modelling are presented in this chapter.

2.2 RAINFALL MEASUREMENT

Eagleson (1970) has observed that gauging stations when few in number, are usually located in the valleys near population centers for reasons of accessibility, and are therefore at an elevation less than the average for the entire basin, leading to under-estimation of average precipitation depths over a basin. Neff (1977) studied rain gauges of different sizes and shapes, and reported that the normal rain gauges caught between 5 and 15 % less rain for
all rainfall events. For single rainfall events, where the total catch exceeded 12 mm, the error was reported to range between 0 and 75% depending on wind characteristics during the storm.

2.2.1 Spatial Variability of Rainfall

Huff (1970) studied the spatial variability of rainfall and observed that more intense storms showed a low rate of variability, and the light showers possessed high variability. However, Niemczynowicz (1984) observed that highest deviations between point and areal rainfall were observed during the most intensive rains. The isohyets of precipitation event, representing spatial variability by closed elliptic curves were considered acceptable (Zawadzki 1973), regardless of type of rain and nature of variation (Eagleson 1970).

Aitken (1973) suggested that proper representation of the areal distribution of rainfall was essential to minimize the error between a sequence of recorded and simulated flow. Shanboltz et al. (1981), Osborn & Lane (1982) and Niemczynowicz (1984) indicated that rainfall was the most difficult variable to obtain good areal measurements, and showed that large errors could result when spatial variability of rainfall was ignored. Yair & Lavee (1985) recommended that the rain intensity measured from the rain gauge needed appropriate ratio for adjustment.

2.2.2 Areal Average of Rainfall (AAR)

Eagleson (1970) has generalized that the difference between the AAR, and the point rainfall measured at the centre, decreased with increasing duration, and increased with increasing area. Hogan (1990) reported that location of rain gauges played a larger role than number of rain gauges in reducing errors in estimates of AAR. Viessman & Lewis (2010) indicated that
when the gauge was centrally located, the observation would more closely match the AAR.

Linsley (1976) explained that the rainfall data bias could be corrected by multiplying the measured precipitation by a constant factor, as adopted in SWM IV model. Dhar & Bhattacharya (1977) estimated AAR using an exponential equation with two constants.

2.3 HYDROLOGIC PROCESSES

The hydrologic processes such as infiltration, evapotranspiration and channel flow are constituted by their respective elemental processes. When the space and time step of a model are decreased, the hydrologic models are required to consider simulating elemental processes at source level, which can then be aggregated to estimate the lumped processes.

2.3.1 Interception

Lull (1964) determined that precipitation at an intensity of 17.5 mm/hour required about 15 minutes to saturate the foliage. Lawson et al. (1981) reported that the monthly interception varied from 11.5 to 39.0 mm (13 to 24% of rainfall). Viessman & Lewis (2010) reported that for a well-developed tree, interception of the order of 1.52 mm of precipitation could be expected on the basis of an average retention of about 20 drops per leaf. They had suggested that dense closed forest intercepted as much as 25% of annual rainfall, and grasses, crops and shrubs intercepted 7 to 60% of rainfall.

Rutter et al. (1971) developed an interception model, using free through fall coefficient, canopy storage capacity, and evaporation rate. Beasley & Huggins (1981) listed the values of potential interception of rain by different crops for use in their ANSWERS model. Calder (1986) proposed
a stochastic model of rainfall interception in which drainage rate was related to maximum canopy storage capacity. Thomas & Beasley (1986) estimated interception in the FHM, using modified form of Merriam’s equation. Pitman (1989) found the values of interception (C), as $C_{\text{max}} = 0.467 \text{ LAI}$, and $C_{\text{min}} = 0.156 \text{ LAI}$. Susanto & Kaida (1991) concluded that the rate of interception decreased exponentially with increase in interception storage.

2.3.2 Detention

ASCE (1949) reported estimates of detention depth for various plants: it varied from 0.7 to 2.4 mm for grass, and from 0.1 to 0.4 mm for maize. Miller (1977) reported the approximate sizes of the depression storage for various land uses as: flat pasture = 1 to 5 mm; ploughed land = 50 mm or more; and contour furrows = 40 mm. The detention was allowed to remain trapped, until it eventually infiltrated or evaporated.

Viessman et al. (1970) considered detention depth as an optimizable parameter in their urban runoff model, and the model was found to give satisfactory results, when the average depth of detention storage was taken as 4 mm. Anderson et al. (1978) considered maximum value of surface depression storage in the water balance model as a variable that must be specified by the user. Thomas & Beasley (1986) noted that surface retention and detention were to be satisfied to sustain overland flow. Neilsen & Panda (1988) estimated detention as a calibrated parameter in water yield model, as it was reported to be difficult to measure detention quantity in the field.

2.3.3 Flow from Variable Source Area

Betson (1964) determined that a small portion of the overall watershed was largely responsible for the runoff contributions to the storm hydrographs, and that these contributing areas were variable. Kirkby (1969)
noted that the stream rise could be substantial even when less than 10 percent of the area of the basin was contributing. Dunne & Black (1970) studied runoff producing mechanism in a small drainage basin and observed that major portion of storm runoff was produced as overland flow on a small proportion of the watershed, and that runoff moved quickly to the stream at velocities 100 to 500 times that of the subsurface system. These partial areas contributing quick runoff could expand or contract seasonally or during a storm. The conversion of rainfall to quick runoff is instantaneous. Expansion of the channel area by up to 170% during storms caused a large addition of runoff to the channel from this source. Dunne et al. (1975) reported that the saturated area varied between 12% and 36% of the total watershed area. The saturated areas are common to low topographic elevations in humid areas (Yair & Lavee 1985).

2.3.4 Soil Properties

England & Holtan (1969) observed that the soil properties that influence infiltration, moisture storage, and the hydraulics of surface flow were related to topographic position. Cassel & Bauer (1975) reported coefficient of variation (CV) of bulk density as 7.3%, and that of wilting point as 31% in a study area. Cameron (1978) reported the CV as 6.9% for bulk density, 28 to 62% for textural fractions, 20 to 51% for soil water contents and 190% for saturated hydraulic conductivity. Warrick & Neilsen (1980) grouped the soil properties based on CV as: (i) low variation (7 to 10%): bulk density and water content at zero tension; (ii) medium variation (10 to 100%): sand, silt and clay fractions and water content at different tensions; and (iii) high variation (>100%): saturated hydraulic conductivity, unsaturated hydraulic conductivity, diffusion coefficient, and pore water velocity. Ahuja et al. (1988) reported that CV was less than 5% for bulk density, and greater than 20% for sand, organic matter, water content and macroporosity.
2.3.4.1 Hydraulic conductivity

It is always difficult to estimate soil hydraulic conductivity that represents a vast area or a watershed. The point value, estimated in the field or laboratory, was adjusted to get effective value for a watershed. Bouwer (1969) studied the infiltration-time relationships for flood irrigation in coarse-textured and fine-textured soils to evaluate the effect of non-uniform inundation time on irrigation efficiency, and reported that the hydraulic conductivity in the Green–Ampt equation should not be taken as the value of saturated hydraulic conductivity (Ks) estimated from laboratory experiments, but rather as 0.5 Ks due to air entrapment. Neilsen & Panda (1988) had accepted the Bouwer’s postulation and incorporated that in their water yield model. Stone et al. (1992) tested Infiltration and Runoff Simulator (IRS) program, as a part of WEPP model, and computed infiltration and rainfall excess on a plane and the overland flow hydrograph at the end of a plane; they had estimated saturated hydraulic conductivity from the estimates of Rawls et al. (1982) data, and halved it to account for the effects of crusting.

2.3.4.2 Pedotransfer functions (PTF)

Pedotransfer functions (PTFs) are statistical predictive equations used to estimate soil hydraulic and soil moisture content values from basic soil properties. A simple PTF equation is of the form: $FC = [0.2576 - 0.0020 SD + 0.0036 CL + 0.0299 OC]$ (Rawls et al 1982), where FC = moisture content of soil at field capacity (fraction), SD = sand content, CL = clay content and OC = organic carbon content of soil in weight percentages. Some PTF equations used in the WAPROS model are given in Appendix 3.

The PTFs help in translating ‘data we have’ into ‘data we need’ (Bouma 1989), and the raw soil data acquire value addition and information with the PTFs. The PTFs were developed in various parts of the world, often
with the bias towards the characteristics of local soils. Some researchers and organisations had created large international databases, such as UNSODA, HYPRES, WISEP, etc. and helped devising the PTFs for general and wider applications (Pachepsky & van Genuchten 2011).

The outputs of PTFs substituted direct measurements that were laborious, costly and time consuming (van Genuchten et al. 1991). The biggest advantage of using the PTFs is that it addresses the problem of spatial variability, by estimating soil parameters having high CVs (>100%) in the field, from the measured inputs having low CVs (< 30%) (Rawls et al. 1991).

The versatility and applicability of PTFs for humid and tropical soils across continental boundaries were proved in a few studies (Tietje & Hennings 1996). The variations of soil properties along the slope and landscape were said to be ignored by the PTFs, but it was defended as being taken care by the differences in soil particle distribution and appropriate sampling (Pachepsky & van Genuchten 2011). The presence of soil cover, crust, coarse fragments, and rock could affect the soil properties, for which necessary adjustment PTFs had also been devised (Rawls et al. 1988; Rawls et al. 1989). The omission of soil structure in soil water interactions, due to its semi-quantifiable nature, in the formulation of PTFs were reported to be offset by inclusion of organic matter content, bulk density, and porosity in the PTFs (Rawls et al. 2003).

2.3.4.3 Ensemble approach

Some PTFs were evaluated to be more suitable for one type of soil, and unsuitable for other types of soils (Minasny & McBratney 2000; van Genuchten et al. 1991). Few PTFs were evaluated to underestimate some, and overestimate other soil properties (Tietje & Hennings 1996). It was reported that no single PTF could be recommended as the most suitable for all soils.
(Guber et al. 2009; Liao et al. 2014). These constraints with single PTF necessitated shift to ensemble approach or multi-modelling (Pachepsky & van Genuchten 2011).

The ensemble method considers estimating different values from multiple PTFs, and finding the soil property value either by averaging, or by regressing (Guber et al. 2006). This ensemble approach was gaining more acceptance in applications (Guber et al. 2009; Pachepsky & van Genuchten 2011). The ensemble method was reported to decrease the errors in individual PTF (Guber et al. 2009), and improve its reliability (Pachepsky et al. 2010). The ensemble output was also reported to be significantly better than the best single PTF (Liao et al. 2014). While estimating the water retention functions, it was reported that PTF ensemble gave two times smaller errors than laboratory data (Guber et al. 2006). The multi-modelling approach has been used for streamflow forecasts at multiple locations in Colorado (Regonda et al. 2006).

2.3.5 Macropore Flow

Maxey (1964) explained that under favourable hydro-geologic conditions, a large fraction of the precipitation moved directly into the groundwater reservoir, and eventually reappeared as baseflow. Freeze & Banner (1970) found that the drying cracks caused fracture flow, resulting in deviation between the model and the field data. Ormsbee & Khan (1989) distinguished macropore flow as the turbulent movement of water through interconnected macro-channels in the soil matrix, and micro-pore flow as the non-turbulent movement of water through the soil matrix.

Germann & Beven (1981) suggested a statistical approach that related radius of pore and capillary pressure, for gravity dominated flow, using the equation, \( R = \frac{(-2\sigma)}{(\psi \rho g)} \), where, \( R \) = macropore radius,
\[ \sigma = \text{surface tension at the air-water interface}, \; \psi = \text{capillary potential in units of length}, \; \rho = \text{density of water}, \; \text{and} \; g = \text{acceleration due to gravity}. \] 
As pore radius (R) corresponding to a capillary potential (\( \psi \)) of zero is undefined, they had chosen an arbitrary value of \( \psi = -1.0 \text{ cm} \) as boundary between macropores and micropores. Timlin et al. (1999) used air entry value, or bubbling pressure (\( h_b \)) (absolute value) employed in Brooks-Corey soil water retention model, for estimating radius of the pore as: \( r = (0.148/h_b) \). The value of \( r \) so obtained represents the average value of pores, including that of micropores. Buczko et al. (2003) adopted Hagen-Poiseuille equation for laminar flow through a capillary tube to determine upper bound for the number of effective macropores per unit area, and obtained estimate of the minimum value for macroporosity.

Simunek et al. (2003) explained about the unsuitability of using pore-scale physical laws, pointing out that macropores rarely flow at full saturation. As capillary potential differs from soil to soil, assumption of an arbitrary constant value of \(-1.0 \text{ cm}\) was considered not acceptable.

### 2.3.6 Overland Flow

Many localised phenomena such as the amount, angle of incidence and spatial variability of rainfall affected hydraulics of overland flow, and restricted theorisation of dynamics of overland flow for universal application (Guy et al. 1990). The overland flow eroded soil particles at one location, transported and deposited a part of it at some other location with successive loss of energy (Emmett 1970; Morgan 1980; Zhang et al. 2010; Almeida et al. 2014), affecting hydraulics of overland flow. Not all area contributed to rainfall excess (Kirkby 1988), not all rainfall excess contributed to the channel flow (De Lima 1992; Darboux et al. 2002), and not all detached sediment contributed to sediment delivery at outlet (Parsons & Abrahams 1992). The suspended sediment normally increased flow viscosity and
therefore changed flow regimes \cite{Roels1984, Zhang2010}. The roughness values were also reported to vary due to tillage operations \cite{Zobeck1987, Gilley1991}. Guy et al. \cite{Guy1990} and Myers \cite{Myers2002} opined that the flow could be more of laminar, owing to suspension of sediments. The analysis of flow hydraulics was further complicated by:

\begin{enumerate}
\item rain drop impact \cite{Emmett1970};
\item initial dry plane \cite{Freeze1974};
\item formation of coalescing micro-channels or rivulets and concentrated flow \cite{Abraham1986, Takken2000, Parsons2006, Sidle2007};
\item micro topography and depression \cite{Thompson2010}, and variable flow resistance \cite{Kim2012}; and
\item partial and full submergence of vegetation \cite{Lawrence1997, Lawrence2000, Roche2007}.
\end{enumerate}

The overland flow increases in depth and velocity across the slope downstream, and the flow regime changes from laminar to turbulent \cite{Emmett1970, Lane1977, Parsons1990}. The coexistence of both laminar and turbulent flow regimes at the same time was also reported \cite{Robinson1996}.

Zhang & Cundy \cite{Zhang1989} suggested that micro-topography, surface roughness, and soil hydraulic properties varied over short distances, and that micro-topography was the dominant factor causing spatial variation in overland flow depth, velocity and direction. Guy et al. \cite{Guy1990} and Zhang et al. \cite{Zhang2010} had pointed out that sediment load had significant effects on the resistance to overland flow. Roughness elements were fully submerged by water in case of pipe flow, while these protruded above the shallow depth in case of overland flow, requiring substantial modifications in the flow.
hydraulics (Lawrence 1997; 2000; Takken & Govers 2000; Roche et al. 2007; Antoinea et al. 2012). These uncertainties characterize the overland flow as a highly complicated natural process that is not fully understood (Rose 1985; Pilgrim & Cordery 1993; Lawrence 2000; Zhang et al. 2010).

Lack of requisite data to match the requirement of complex model was one of the reasons that often prompted hydrologists to prefer conceptual or lumped models (Brath & Montanari 2000). More modellers preferred Manning’s equation for describing overland flow for its simplicity, despite a dimension was attached to its n (Savat 1977; Morgan 1980; Thompson et al. 2010; Kim et al. 2012). In the simulation models, like SWM IV (1966), FESHM (1978), SWMM (1982), and IRS (1992), overland flow was represented by Manning’s equation. As determination of an effective average n value was difficult, it was estimated as a parameter in a few models (Esteves et al. 2000; Antoinea et al. 2012).

2.3.6.1 Runoff

Johnstone & Cross (1949) pointed out that all the propositions of unit hydrograph theory were empirical, and it was not possible to prove them mathematically, and not a single one of them was mathematically accurate. Knapp et al. (1975) have noted that the runoff from the basin moves over the surface in a complicated system of ditches and ephemeral streams, which cannot be deterministically evaluated. Singh (1977) opined that the amount and distribution of rainfall-excess in time and space was really never known, and had pointed out that errors in distribution of rainfall-excess estimates might be large, even when the runoff volume was known. Kirkby (1985) narrated that the catchment hydrographs appeared to show much less variety than catchment topography and soils, implying that massive averaging was commonly taking place to suppress local detail.
Das (1982) reported that in the Nilgiris, surface runoff was mostly around 2 to 5% of the incident rainfall. Mittal et al. (1988) had indicated that a rainfall of 482 mm during 1981 season produced 146 mm runoff (30.3%), and the same amount of rainfall in the same season during 1984 could produce only 111 mm of runoff (23.0%).

2.3.7 Baseflow

Takeda & Ishii (1968) observed that while the subsurface flow encountered more friction than the surface runoff, it occupied a thicker cross section and the slower speed of subsurface flow was compensated by greater cross sectional area of flow. In northern Japan, the lag of flow was found to decrease in large storms. In the experimental basin of Iwate university, this lag was found to be 25 hours for storms of 10 mm, but only 5 hours for storms of 40 mm. Appleby (1970) related baseflow process to memory concept and defined baseflow as propagating antecedent or historical effects, carry-over signals from the past, and storage in transit. Rushton & Ward (1979) cautioned that unless recharge calculations were made on a daily basis, a significant underestimate of the recharge to groundwater might occur; weekly and monthly input data resulted in underestimates of 10% and 25% respectively.

2.3.8 Evapotranspiration (ET)

The Penman Monteith ET equation (PME) is the most widely used method to estimate ET. The performances of other ET equations were evaluated against PME as a standard (Amatya et al. 1995; Xu & Singh 2002; George & Raghuwanshi 2012; Efthimiou et al. 2013; Nikam et al. 2014). However, the data requirement for PME is highly demanding, which could not be fulfilled in data scarce regions and in developing countries (Droogers & Allen 2002; Tabari et al. 2011; Zhao et al. 2013; Nikam et al. 2014). The
non-availability of quality data was reported to cause serious limitations in using PME (Efthimiou et al. 2013).

2.3.8.1 Sensitivity of ET to wind speed

While wind is needed to move out saturated layer to generate steady ET, its sensitivity to ET is reported to be low. Saxton (1975) reported low sensitivity of PET to the wind profile parameters. Goyal (2004) estimated that PET had a low sensitivity of 7% to wind speed. Sharifi & Dinpashoh (2014) found that ET was less sensitive to wind speed. Ambas & Baltas (2012) have suggested that wind speed was not so important for the calculation of ET. Alexandris & Kerkides (2003) proposed a new empirical formula for estimating hourly reference ET, without using wind speed data, and reported it as functioning satisfactorily.

2.3.8.2 Hourly distribution of ET

Hillel (1977) used a sine function to partition daily PET estimates into hourly values. Anderson et al. (1978) calculated evapotranspiration in the water balance model six times a day at 4 hour interval. Any potential evaporation energy remaining after the evaporation from interception storage was divided between potential soil evaporation and potential plant transpiration.

The percentage distribution of PET at 4 hour interval commencing from midnight was 2.4, 4.8, 29.0, 39.7, 19.5 and 4.6%. Campbell (1985) estimated the hourly rate of PET, using the hourly temperature and radiation data obtained from their respective daily data. Sachan & Srivastava (1988) explained a procedure to distribute daily evapotranspiration values into hourly values. The temporal distributions of temperature and PET nearly follow the same pattern.
2.3.8.3 Modelling ET

In the Stanford Watershed Model (SWM IV), evapotranspiration was considered to occur from interception storage, and upper zone storage at the potential rate. If the potential demand could not be satisfied from these sources, the remaining potential depleted the soil moisture (Crawford & Linsley 1966). Campbell (1985) partitioned PET by two methods: (i) 10% evaporation and 90% transpiration; and (ii) based on the ratio of the radiation intercepted by the crop to total incident radiation. Kropff (1993) noted that actual evaporation was partly taken from the topsoil layer (26%) and the rest from the rooted layer (74%).

2.4 SIMULATION MODEL

Robinson (2004) describes simulation as an experimental approach to modelling, and the model user can explore alternative scenarios that improves understanding of the real system. A model plays the role of a surrogate for the system it represents (Birta & Arbez 2007), and the simulation represents the behaviour of the system under given conditions.

Simulation model is more suitable for experimentation, as it reduces the cost and duration of experiments, and facilitates experimentation under controlled conditions (Chung 2004). Models are found to be more appropriate to handle systems that are rare, costly, and dangerous, and the interventions that are disruptive, irreversible, and ethically unacceptable (Birta & Arbez 2007). Hence, Robinson (2004) designates the simulation model as a form of decision support system.

Hydrologic models and environmental models simulate natural processes that are not fully understood. However, these models improve the understanding of the interactions between those processes.
2.5 HYDROLOGIC MODELS

Many rainfall-runoff models have been used for research purposes to enhance the knowledge and understanding about the hydrological processes that govern a real world system (Moradkhani & Sorooshian 2008). These models are also used for estimating sensitivities of watershed parameters, and for driving reservoir operation, irrigation, erosion, and pollution models.

2.5.1 Model Development

An overly comprehensive and complex model is too difficult to construct, analyze and communicate. Its voluminous results may cloud the important issues in irrelevant detail. On the other hand, an overly simplified model may not be capable of exhibiting some effects or relationships which are essential in the operation of the system (Hillel 1977). Dent & Blackie (1979) opine that model building is not an exact science, because the format of the model, and degree of detail in the model, to mimic the behavior of a real object, remains a matter of judgment on the part of the model builder.

2.5.2 Use of Hydrologic Models in India

Venugopal et al. (1983) applied WBNM proposed by Boyd and others, for modeling rainfall excess-surface runoff relationship. The model, which relied on routing, was run after dividing the catchment into ordered basins and inter-basins. Gosain & Kapoor (1988) had developed a tank model, based on Sugawara’s tank model, and applied that on the Banchhor tank catchment of Madhya Pradesh. Shrivastava & Bhatia (1992) explained that they had to use the USDA–SCS model built with curve numbers, to compare the rainfall runoff data collected, as other models developed in India or in other countries required detailed input data, which were generally not available in India.
2.5.3  Seasonal Models

The continuous models are generally used to simulate runoff continuously for one, or more years, but sometimes for less than a year too. Anderson et al. (1978) used Water Balance Model for simulating 78 day record. Sachan & Srivastava (1988) had tested the models by comparing seasonal runoff data from watersheds and reported fairly good agreement between simulated and observed seasonal runoff using RUNMOD. Mittal et al. (1988) had conducted hydrologic studies in Chandigarh, for a season spanning from July to September.

2.5.4  Performance of Available Models

Viessman et al. (1970) studied error distribution in the runoff model and reported that the absolute average error was 9%, and the average error in peak discharge prediction was 13.8%. About 80% of the computed peak outflows differed from the gauged peak flows by less than 20%, while about 75% of the computed peak flows were in error by less than 15%. Errors of this magnitude were considered to be reasonable. Wagner & Linsley (1975) applied SWM IV to Wagli catchment in India and observed that the error between simulated and observed flow varied from −76% to +35%. The deviation of simulated discharge from the observed discharge, for the little Arkansas river, using British Hydrological Society (BHS) model varied from −52% to +136% (Knapp et al. 1975).

Crow et al. (1980) applied USDAHL−74 model to two grassland watersheds, and found that the error between simulated and observed runoff varied from −80% to +450%, and the larger runoff events were closely simulated than the smaller ones. Montas & Madramootoo (1991) applied ANSWERS (Areal Nonpoint Source Watershed Environmental Simulation) model to watersheds in Canada and found that error in peak flow ranged from
–89.3% to +198.6%, the error in time to peak ranged from –66.7% to +492%, and the error in runoff volume ranged from –87% to +207%.

Zepp & Belz (1992), when using SWATRE model, noted that the simulated soil water balance and groundwater recharge showed a variation of 180% and 50% respectively. Refsgaard et al. (1992) applied the SHE model to Indian catchments and observed that it was costlier to use SHE, than traditional, and simpler, hydrological models, owing to its complexity and data requirements. Xu (1997) applied 18 models in seven catchments in China and reported Nash Sutcliffe Efficiency (NSE) values from 0.80 to 0.92 for monthly flows. Waichler et al. (2005) applied DHSVM, a process-based distributed hydrologic model and reported overall efficiency in simulating hourly streamflow > 0.7, and mean annual error < 10%. Tripathi et al. (2006) applied SWAT model to Nagwan watershed and reported Volume Deviation Error (VDE) from –10 to +11%. Bari & Smettem (2006) found the VDE was +15% in Lemon catchment, and +2% in a catchment in Western Australia. The compilation of performance data by Moriasi et al. (2007) showed that the highest daily and monthly NSE values for stream flow validation of SWAT model were 0.83 and 0.93; and that of HSPF model were 0.87 and 0.92 respectively. Chiew (2010) used five rainfall-runoff models on 240 catchments in south eastern Australia for daily streamflow and reported NSE values > 0.6 in more than 80% of the catchment studies. Sintondji et al. (2013) reported a coefficient of determination of 0.89, a model efficiency of 0.81, and an index of agreement of 0.96 for weekly stream flow.

2.5.5 Rating Values for Evaluation

Love & Donigian (2002) suggested rating values of VDE for HSPF simulation, which were converted to Volume Handling Efficiency (VHE) as >90% for very good, 0.90 - 0.85 for good, and 0.85 - 0.75 for satisfactory ratings. Boughton (2005) considers calibration results with a coefficient of
determination < 0.6, or VDE > ± 10% are generally too poor to be considered acceptable. Bari & Smettem (2006) have recommended that both the NSE and $r$ are to be > 0.8 for an acceptable calibration for monthly streamflow.

Moriasi et al. (2007) reviewed the results of many hydrologic models and recommended the following values of evaluation criteria for monthly simulation data: (i) NSE: > 0.75 for very good; 0.75 - 0.65 for good and 0.65 - 0.50 for satisfactory ratings; (ii) Mean Error (ME) (Bias) ($\pm$): < 0.10 for very good; 0.10 - 0.15 for good and 0.15 - 0.25 for satisfactory ratings; and (iii) Ratio of RMSE to standard deviation of observed flow (RSR): < 0.50 for very good; 0.50 - 0.60 for good and 0.60 - 0.70 for satisfactory ratings.

### 2.5.6 Effect of Time Step on Evaluation

The values of evaluation criteria are more influenced by the time step, or interval of simulation. As the time interval becomes coarser (hourly to daily to monthly to yearly), the values of error criteria will decrease and that of efficiency criteria will increase, due to large scale averaging. Silberstein et al. (1999) showed Coefficients of determination ($r^2$) were 0.84, 0.92, 0.96 and 0.99 for daily, weekly, monthly and yearly accumulation intervals. Varado et al. (2006) reported that NSE varied from 0.58, 0.66 to 0.74 for daily, 10 day and monthly time steps. Engel et al. (2007) caution that the time period considered can impact model performance, and model performance is poorer for shorter periods than for longer periods. Moriasi et al. (2007) have suggested that as the evaluation time step increases, a stricter performance rating is warranted.

### 2.5.7 Evaluation of Evaluation Criteria

Legates & McCabe (1999) and Willmott & Matsuura (2005) have suggested that when large outliers are present, mean absolute error (MAE) is
preferable to RMSE. Moriasi et al. (2007) recommended use of median values, especially for skewed data, as medians were less sensitive to extreme values when compared to mean values. As MSE and RMSE are quadratic or squared functions, their estimates are reported to be affected by extreme values (Legates & McCabe 1999; Tedeschi 2004; Krause et al. 2005; Berthet et al. 2010). While comparing and evaluating different evaluation criteria, the researchers found a few criteria as unscientific (Willmott et al. 2009), inappropriate (Willmott & Matsuura 2005), misleading and even unfit and not to be used (Willmott 1982).

2.6 WATER BALANCE EQUATION

Beven (2006) states that the hydrologists cannot currently close the water balance (WB) strictly by measurement, and calls the closure problem metaphorically as the Holy Grail of Scientific Hydrology, suggesting that the solution might be out there somewhere in principle, but may be impossible to find. It implies that it is difficult to close the WBE by measurement.

Many hydrologists had reported WBE, without an error term, and assigned the unbalanced value to ET, assuming that change in storage was zero. Nemec & Schaake (1982) have observed that the longer the time-step involved in the WBE, the less significant the terms of storage become and the easier it is to accomplish a balance, and vice versa. Tomasella et al. (2008) have pointed out that groundwater storage and baseflow vary considerably from year to year, and the assumption that the overall system returns to the same state of storage each year becomes invalid.

2.6.1 Water Balance Error

The errors in measured water balance (in mm or % of rainfall) from various studies were reported as: (i) 12 mm, or 2.1% of the precipitation
(Sokolov & Chapman 1974); (ii) – 45.8 mm (Waichler et al. 2005), and (iii) 6%, 10% and 7% of precipitation (Chauvin et al. 2011). Tripathi et al. (2006) applied SWAT model to Nagwan watershed and reported WB errors from 0.010 to 1.62 %. Harder et al. (2007) used Thornthwaite monthly water budget models, and reported that the closure errors varied from 0.1 to 8.8 % of precipitation. Xu & Singh (1998) reported that MAE varied between 1% and 20% with monthly water balance model. Sciuto & Diekkruger (2010) reported that WB error varied from 4.82 to 7.27 mm.

2.6.2 Rating Values for WB Error

A limit for maximum error in WB equation was also suggested: Harder et al. (2007) suggested 5% error target for estimation of water balance components, and Chauvin et al. (2011) compared their data to 10% error.

2.7 MODELLING LIMITATIONS

Pilgrim et al. (1978) studied surface and subsurface flow processes under natural and artificial rains, and reported that the spatial variabilities of these two flow processes were so great that a truly deterministic model would be difficult to develop even for small study plots. Sachan & Srivastava (1988) stated that it was difficult to simulate all hydrologic processes in a watershed with adequate precision. Oreskes et al. (1994) asserted that a perfect model, accurately representing actual processes in a real system, was not even theoretically possible. Tedeschi (2004) has argued that a model can never fully mimic the reality under all conditions. The model developers and users shall take cognizance of this fact, and a model could only be expected to give a better fit, rather than a perfect fit.

2.8 METHODS OF SENSITIVITY ANALYSIS

Cacuci (2003) pointed out that the first-order sensitivity of Local Sensitivity Analysis (LSA) was more stable linearly, and it decreased
dramatically with increasing order. Bastidas et al. (2006) pointed out that one at a time (OAT) methods were more common among the land surface modellers. Saltelli et al. (2006) observed that the Global Sensitivity Analysis (GSA) practices were familiar among applied statisticians, but not among the scientific community. Saltelli & Annoni (2010) had verified with the publications and stated that most often LSA (OAT) was performed regularly, in spite of advances in GSA. They had further indicated that ‘OAT did not make Type I errors’, or it always rejected the wrong, and never reported insensitive factors as relevant.


### 2.9 SCENARIO DEVELOPMENT

Hydrologic models have been used to generate hydrologic scenarios for anticipated changes in climate and land use. Gleick (1986) has reviewed different types of hydrologic models and recommended water balance models owing to its distinct advantages over other models for developing scenarios. The study reports the advantages of water balance models as: historical variations in temperature and precipitation can be modelled to determine the climatic sensitivity of runoff and soil moisture in any given watershed; hypothetical data can be used to test the sensitivity of watersheds to different climatological variations and a series of possible transient climatic responses can be evaluated; and these models can be linked with general circulation model output. Sankarasubramanian et al. (2001) have carried out studies on adopting climate elasticity as an unbiased estimator of the sensitivity of streamflow to climate and constructed regional maps of the sensitivity of streamflow to climate for the continental United States; they have suggested that conceptual watershed models are more suitable for
modeling climate change impact. Bronstert (2004) has pointed out that climate change impact assessment concentrates on the changed meteorological forcing, and the land-use-change impact assessment focuses more on the internal dynamics of the hydrological system and affirmed that rainfall-runoff model is the only economically possible way to study the impact of environmental changes on the hydrological cycle.

2.9.1 Climate Change in India

Govinda Rao et al. (1996) analysed the climatic data over 115 years (1878-1993) and estimated the season-based temperature changes in °C/100 years, as: spring: 0.12; summer: 0.20; post monsoon: 0.52; and winter: 0.50. Arora et al. (2005) studied the data over the last century and indicated that annual mean temperature, mean maximum temperature and mean minimum temperature increased at the rate of 0.42, 0.92 and 0.09 °C/100 years, respectively.

Naidu et al. (2009) reported that the summer monsoon temperature was rising at the rate of 0.212 °C and 0.188 °C /10 years in the southern and northern regions respectively. In another study, the rates of warming, in °C /year, were reported as 0.0126 in the southern region and at 0.0098 in the northern region (Naidu et al. 2010). Jain & Vijay Kumar (2012) have reported that mean minimum temperature showed a rising trend in the south, central and western parts of India; annual mean temperature showed a falling trend in the north and north eastern India; the rising trends in maximum and minimum temperature were different at different locations. Khan et al. (2015) analysed monthly and seasonal temperatures in Kolkata and reported the warming trend for the summer and winter seasons at 0.034°C/year, and 0.032°C/year, respectively.

Govinda Rao et al. (1996) estimated the changes in rainfall to vary between +10 and +12% /100 years of normal rainfall. Dash et al. (2007)
reported that Indian summer monsoon rainfall decreased by about 1.6 cm during 131 years (1871–2002). The summer monsoon rainfall exhibited a very small negative trend of 4 mm/decade (Naidu et al. 2009). It was also reported that Karnataka, Andhra Pradesh and parts of Rajasthan and some parts of eastern India were experiencing more wet days, while Kerala and Tamil Nadu were experiencing more dry days (Guhathakurta et al. 2011). Pal & Al-Tabbaa (2011) have projected a decreasing trend in spring and summer (monsoon) seasons, and an increasing trend in the winter and autumn seasons for Kerala. Menon et al. (2013a) have generalised with the aid of Coupled Model Inter-comparison Project Phase 5 (CMIP5) models that monsoon circulation due to global warming would be strengthening in the northern parts and weakening in the southern parts of India, showing a northward shift in the monsoon.

Jain & Vijay Kumar (2012) have analysed basin-wise rainfall pattern and reported that annual rainfall is increasing at 0.27–10.16 mm/year in six river basins, and decreasing at 0.45–4.93 mm/year in 15 river basins in India. Deka et al. (2013) analysed 110 year (1901–2010) rainfall records of Brahmaputra and Barak basins of Assam, NE India, and reported that annual and monsoon rainfall showed decreasing trends in both the basins. Chakraborty et al. (2013) observed decreasing trends in annual rainfall from +0.1 to -13.6 mm/year in Seonath River Basin, Chhattisgarh; the decrease was noticed to be more in winter and monsoon rainfall than during summer rainfall. Deka et al. (2015) have observed from recent 30-year data a decreasing trend in monsoon rainfall, and increasing trends in pre-monsoon and post-monsoon seasons in the Brahmaputra valley. Kumar et al. (2015) studied rainfall pattern in the Sutlej river basin and reported great temporal and spatial variations, unequal seasonal and geographical distribution with frequent departures from normal. Gajbhiye et al. (2015) analysed the rainfall pattern in Sindh basin, India and observed that monthly rainfall showed: (i)
positive trend during April, May, August and October; (ii) negative trend during July and September; and (iii) no trend during November and December.

The futuristic climate change predictions for India are found to be more ambiguous (Rosenberg 1991). Jain & Vijay Kumar (2012) have observed that the results of various climate change studies on India widely differ and a clear and consistent picture of rainfall trend has not emerged.

Govinda Rao et al. (1996) estimated the anticipated rise in temperature for the year 2050 to be 0.75 °C in the south and 1.5 °C in the north, averaging 1.0 °C for India. Meenu et al. (2012) projected that the maximum daily temperatures would increase by 1.0, 2.1, and 3.4 °C respectively, in the 2020s, 2050s, and 2080s. Islam et al. (2012) have estimated from GCM studies that maximum temperature would increase by 4°C and annual rainfall change would vary between −3.30 and +29.6% in Brahmani River basin.

Govinda Rao et al. (1996) have predicted that the largest increase in rainfall will be 15 % over the northwest by the middle of the 21st century, and the increase will be considerably lower in the extreme north and south. Gosain et al. (2011) have projected marginal decrease in rainfall in Brahmaputra, Cauvery and Pennar river basins up to 2050, and increase in rainfall for all the river basins from 2071 onwards.

Menon et al. (2013b) have explained that enhanced rainfall with global warming in a few locations is due to (i) increase in water holding capacity of the atmosphere and higher precipitable water content, and (ii) the upper-tropospheric cooling that destabilises the atmosphere. They have pointed out that surface fluxes from ocean increases due to a warmer sea surface temperature and enhances the rainfall intensity over regions that
already have strong moisture convergence, calling the phenomenon as 'rich-get-richer' mechanism.

Rosenberg (1991) has cautioned that the harsh predictions for more increase in temperature and sharp decrease in rainfall may be possible, but not plausible, and hence moderate changes in temperature are considered credible. Arora et al. (2005) has advocated to take into account the possible cooling effect of sulphate aerosols, which requires downscaling of increase in air temperatures by about 0.7–1.0°C in the 2040s, in comparison to the 1980s.

Refsgaard et al. (1989) have categorised the changes in climate into climate variability and climate change. They have defined ‘climate variability’ as up or down variation of climate variables in a year, and ‘climate change’ as a long term change in climate variables over the years. They have described that climate variability is routine, reversible and temporary, generally caused by natural conditions, while climate change is irreversible and permanent, caused by both natural conditions and human interference.

2.9.2 Scenario Studies: Climate Variability

For scenario study under climate variability, Nemec & Schaake (1982) adopted rainfall changes at intervals of –25%, –10%, +10% and +25%; Chiew (2006) used: –15%, –10%, 0% and +10%; and Wang et al. (2008) used: +3%, +5%, +10%, +15%. Nemec & Schaake (1982) used temperature changes at increments of –1 °C, +1 °C, and +3 °C; and Wang et al. (2008) used +1 °C, +2 °C, +3 °C, and +4 °C.

When precipitation was increased by 10%, Nash & Gleick (1991) reported 11% increase in streamflow. Chiew (2006) reported that a 1% change in rainfall resulted in a 2.0–3.5% change in annual streamflow. Fu et al. (2007) estimated that a 1% precipitation increase generally resulted
in a 1.1 – 1.4% increase in streamflow and a 1% precipitation decrease resulted in 1.6% decrease in streamflow. Wang et al. (2008) projected that a 10% increase in precipitation resulted in a 13% increase in annual streamflow. Xu et al. (2012) found a +1% change in annual precipitation resulted in 2.0–4.0% change in runoff.

Nemec & Schaake (1982) found that a change of 1°C resulted in a 4% change in evapotranspiration. Fu et al. (2007) noticed that 1.5°C temperature increase resulted in a streamflow decrease of 15%. Wang et al. (2008) estimated that the channel flow was reduced by about 1.1% when temperature was increased by 1°C. Xu et al. (2012) observed that a 1°C increase of the temperature resulted in a −0.05% change in runoff.

2.9.3 Scenario Studies: Climate Change

For climate change study, Nash & Gleick (1991) used 10 combinations of temperature and rainfall changes as: T+2°C; T+4°C; and P−20%; P−10%, P+0, P+10%, P+20%. In a scenario study in Brahmani basin, rainfall changes varying from ±10 to 30% with an increment of 10% and temperature changes varying from 0 to 4°C with an increment of 2°C were considered (Islam et al. 2012).

Gleick (1986) found that in a humid basin, a 1°C increase in temperature coupled with a 10% decrease in precipitation reduced runoff by 25%. Fu et al. (2007) estimated that a 20% precipitation decrease resulted in a 25 – 30% decrease in streamflow at same temperature, but a 50% decrease in streamflow if the temperature increased by 1.5°C. Yang & Yang (2011) estimated that a 1% change in rainfall led to a 1.6% – 3.9% change in runoff, and a 1°C temperature increase produced a 2% - 11% decrease in runoff.
Islam et al. (2012) have carried out scenario studies in Brahmani basin and reported that a 4°C rise in temperature resulted in 11.40% decrease in annual streamflow, whereas 10% decrease in rainfall resulted in 22.90% decrease in annual streamflow. The combined effect of 4°C temperature rise and 30% rainfall increase resulted in 62% increase in annual streamflow. A 10% decrease in rainfall resulted in 25.00, 12.40, and 21.10% decrease in streamflow during monsoon, pre-monsoon, and post-monsoon season respectively, whereas 4°C increase in temperature resulted in 12.00, 2.70, and 11.20% decrease in streamflow during the same seasons.

Narsimlu et al. (2013) have carried out scenario studies on climate change impacts on water resources of the Upper Sind River Basin using SWAT model. They have reported from model studies that annual streamflow could increase by 16.4 % for the mid-century and a significant increase of 93.5 % by the end-century; the streamflow is projected to rise drastically during monsoon season than during non-monsoon seasons.

2.9.4 Scenario Studies: Land Use Change

Bultot et al. (1990) increased the impervious area at 0, 5, 10, 15 and 20% of the total area of the watershed for scenario studies, and reported that 5% positive increment in the area of impervious surfaces caused: (a) + 16mm increase in streamflow and + 33 mm increase in surface runoff, but 17 mm decrease in baseflow; the enlargement of impervious areas also resulted in more irregular streamflow and larger amplitude variations. The effects of urbanization can be considered analogous to enlargement of impervious area. Rose & Peters (2001) summarised the effects of increased imperviousness as: (1) a higher proportion of precipitation appeared as surface runoff; (2) the lag time between precipitation and runoff was decreased; (3) peak flow magnitudes were increased; and (4) base flow was decreased. Jacobson
(2011) confirmed that urbanisation was positively correlated with stream flashiness. Shaw et al. (2011) reported that enlargement of impervious area caused a steeper rising limb of the flow hydrograph, and a high peak flow.

Hibbert (1967) has concluded his review of paired-catchment experiments with three statements: (i) that deforestation increases water yield, (ii) that afforestation decreases it, but (iii) that the response is variable and largely unpredictable. Samraj et al. (1988) had investigated the effect of afforestation on annual water yield, and observed that the black wattle and blue gum plantation reduced the annual runoff by an average of 87 mm (16%), which was statistically significant at 1%. Chhabra & Geist (2006) have suggested that afforestation causes reversal of the hydrological responses to deforestation. Schleppi (2011) has stated that forests generally deliver less water than other types of vegetation during periods of low flow, and that the “sponge theory” appears to be falsified in most cases. He has further pointed out that when afforestation is accompanied by ploughing, peak flows are modified. Raftoyannis et al. (2011) have concluded that the high transpiration demand of afforestation would immediately bring about water problems in the more arid environments.

Shaw et al. (2011) compared the two contrasting views on the hydrological role of forests: (i) sponge theory considers forests act like sponges, hold up more water during rains and release it during dry periods; this concept treats forests are good for promoting groundwater recharge, and maintaining dry period river flows; and (ii) pump theory considers forest roots act like pumps and remove more water for transpiration, and allow less water to groundwater, and consequently less baseflow. Though sponge theory has been promoted for afforestation activities on conceptual grounds, experimental and model studies have often falsified sponge theory, and supported pump theory.
2.9.5 Effect of Crop Cultivation Practices

Raghunath et al. (1967) conducted experiments at Ootacamund and reported that by adopting contour-farming for potato on 25% slope, the runoff was reduced from 52 to 29 mm. Tejwani et al. (1975) reported that the annual runoff was reduced to 41.2% of rainfall under contour-cultivation, when compared to 54% under up-and-down cultivation. Miller (1977) has generalized that plant population increases infiltration and surface detention; plant stems and leaves slowdown the flow of surface water; greater plant activity dries out the root zone, increasing its capacity to take in more water from the next rain, causing reduction in runoff.

Peck & Williamson (1987) estimated that when agriculture area is enlarged, the potentiometric surface moved upward at 2.6 m yr\(^{-1}\), which was equivalent to increased recharge of 6-12% of rainfall. Kuchment (1989) observed that different types of ploughing changed the volume of runoff by 30-40%, leading to opposite results when estimating the impact of deforestation on the runoff. Kale et al. (1992) reported that intercropping system reduced the runoff. Tomer & Schilling (2009) noted that increasing baseflow in Iowa’s rivers was significantly related to increasing agricultural intensity; increasing streamflow in the Mississippi river was mainly due to an increase in baseflow resulting from land use change from perennial to annual cropping systems.

2.9.6 Effect of Conservation Measures

Onstad & Jamieson (1970) described that the terraced watershed had a large depression storage compared to un-terraced watershed, which tended to increase the subsurface flow. Rambabu et al. (1974) observed that field bunding of agricultural watershed in Dehra Dun reduced total runoff by 62%, and the peak runoff rate by 40%. Sud et al. (1977) determined that
bunding of agricultural land conserved on an average 62 mm of rainfall per year and reduced the runoff and the peak rate of runoff by 36.35% and 41.7% respectively. Sato et al. (2009) found that the soil and water conservation measures decreased not only soil erosion, but also decreased surface runoff by about 14 to 74% respectively.

2.10 REVIEW OF HYDROLOGIC MODELS

Franchini & Pacciani (1991) evaluated SWM IV, Sacramento, Tank, APIC, SSARR, Xinanjiang and ARNO models, and pointed out that: inclusion of the unit hydrograph lead to overestimation of the concentration time; increase of parameters and consequent increase in difficulty of the calibration procedure; and difficulty in supplying an adequate quantitative description of the flows within the soil.

Yan et al. (2014) have compared three models for application in Little River Catchment in Georgia, USA and concluded that SimHyd and Tank models were not suitable for the study area, as these models showed more than 10% errors in calibration and verification.

Gayathri et al. (2015) have reviewed the performance of SWAT, MIKE SHE (physically based), HBV, TOPMODEL (semi distributed conceptual) and VIC (semi distributed grid based) models and pointed out that the major constraint is in getting large quantity of input data. They have concluded that research is still going on to make better predictions and there is still scope for improving the existing theories and for developing new theories to assess the impact of climate change and land use changes.