APPLICATION OF SNOWMELT RUNOFF MODEL USING GIS AND REMOTE SENSING TECHNIQUES FOR THE UPPER KABUL RIVER BASIN

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by

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CANDIDATE'S DECLARATION

I hereby certify that the work being presented in this dissertation titled "**Application of Snowmelt Runoff Model Using GIS and Remote Sensing Techniques for the Upper Kabul River Basin"** is presented on behalf of partial fulfillment of the requirement for the award of the **Master of Technology** in **"Irrigation Water Management (IWM)"** and submitted to the Department of Water Resources and Management (WRD&M) IIT Roorkee. This is authentic record work carried out during July 2018 to May 2019 under the supervision of Dr. Ashish Pandey, Associate Professor, and Er. R.D. Singh, Visiting Professor, Department of WRD&M, IIT Roorkee, Uttrakhand, India.

The matter Presented in this dissertation has not been submitted by me for the award of any other degree or diploma of this or any other institute.

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Certification

This is to certify that the above statement made by the candidate is correct to the best of my knowledge.

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ABSTRACT

The Snow which accumulates in the Upper Kabul Basin, is one of the great source of water for Kabul river, Afghanistan. The precise estimation of the volume of water stored in the snowpack and its release are essential for the efficient management of water resources. Generally, the accumulation of snow in snow-fed basins starts during winter season but its depletion starts during the spring and summer season. The extent or magnitude of the snow storage and it melting rate are based on climatic condition over the basin, once the melting era or period begin, the basin or catchment act like storage reservoir flow out continuously through the melt river or stream. The melted water routed through different stages, before reaching to the out-let of the basin. The melt water first percolates through the existing snowpack very slowly and reaches the ground surface. On the ground surface it travels as overland flow and then joins the melt stream. The present study has been carried out with the objectives of estimation of the snow cover area of the Upper-Kabul River basin, Computation of snowmelt runoff, Estimation of the flow for the Upper Kabul river basin using SRM Model.The Upper-Kabul River Basin selected for this study has an area of approx. 1654 sq. km, lies between (34^0 51’52.7” N) and (69^0 4’ 37.2” E). The elevation of the study area varies from 1822m to 4515m. The basin is located in the higher part of the Kabul river basin and has widely snow cover area during winter season. Information about snow cover area is essential for the observing or Monitoring of snowfall in the mountainous area. Consistent mapping and monitoring of snow cover by traditional means is difficult due to the less number of rain-gauges and inaccessible or remote terrain and high cost of the labor. Thus remote sensing techniques can be used to estimate the aerial extent of the snow cover in the basin. Furthermore, Digital elevation model (DEM) of the Upper- Kabul basin has been divided into 8 elevation zones with altitude variance of 322.22 m each zone. The temporal SCA for the period of 2015 to 2017 was derived from MODIS-8 days Images of snow cover data and SRTM DEM have been analyzed through GIS and ERDAS Imagine to obtain the Snow cover in different elevation zones. The daily SCA in each elevation zone has been calculated, snow cover depletion curves were derived and daily runoff was computed by SRM Model from snow cover area. In this study, 8-day snow cover products from MODIS and SRTM (DEM) of the basin with Geographic information system (GIS) and ERDAS IMAGINE and finally into the windows version of snowmelt runoff model (WinSRM or SRM) have been used.

INTRODUCTION

CHAPTER-1

4.1. GENERAL

The term precipitation means all forms of water that reach the earth from the atmosphere. The common forms of precipitation are Rainfall, Snowfall, Hail, Frost and Dew. Snow is one of the main form of precipitation, which is grainy and composed of ice crystals, and precipitate from the atmosphere (usually from clouds). It is a very important component of the hydrological cycle and; it plays a key role in the water resources in many parts of the world.

Stream-flow demonstrating the runoff phase of the hydrologic cycle is the most important basic data for hydrologic studies. The most important component for the planning of water resources development is accurate data on stream flow, or in other words, the surface runoff for a considerable period of time so as to determine the extent and pattern of the available supply of water. The practical objective of hydrologic analysis is to determine the characteristic of the hydrograph that may be expected for a stream draining any particular watershed. There is a great need of proper assessment of mountain water resources because mountain snow fields or snow cover areas are a major source of fresh water as it performs as natural reservoir that can be utilized for many purposes.

Afghanistan is a country predominated by a dry and semi dry climate with most of the area described by arid land. The effect of global climate change on hydrologic systems, especially on mountain snow and glacier melt, can modify the timing and amount of runoff in mountainous watershed. Therefore, accurate stream flow simulation and prediction are significant important for water resources management and planning. High mountain snow and glaciers are at risk due to climate change phenomenon. High snowmelt water and recession of snow cover happen in summer season. A large amount of this melt water reliably happens at this particular time of the year. To decrease the hazard from floods caused by heavy and rapid snow and glacier melt.

The upper Kabul river basin receive snow and rain in the winter season usually the basin receives the rain at the end of the winter season. The contribution from rain dominates in the lower part of the basin (about < 2500 m). the middle and upper parts of the basin (about >2500 m) receive contribution from both rain and snowmelt during summer season and this change with altitude. As the elevation of the basin increase, the rain contribution to streamflow reduces and the snowmelt contribution increases.

As a result, the main source of water in Afghanistan is snow, which falls in winter season and less rain during summer season. If these two sources are properly harnessed it can be used for water supply, agriculture, industry and energy.

The measurement of snow cover area is quite difficult and expensive task in the higher elevation. Recently remote sensing techniques have been proved to be an efficient and un inexpensive tools for such purposes. The MODIS digital image classification through ERDAS IMAGINE along with Geographic Information system (GIS) application has made it possible to determine the SCA accurately. Estimation of the daily flow from the snow cover and free snow cover area can be carried out using Snow melt runoff model (SRM). This is a suitable model for the simulation of runoff from snow cover area.

Several studies have reported the capability of SRM model for computation of snowmelt runoff and climate change effects to runoff in various basins around the world. In the present study, SRM model has been used for estimation of snowmelt runoff from the Upper Kabul river basin. The main aim of this study is to simulate daily discharge from snow cover area and examining the adaptability of the model structure based on the availability of the variable and estimation of parameters for the upper Kabul River basin in Afghanistan.

4.2. OBJECTIVES

Following are the specific objectives of this study.

- Estimation of snow-covered area of the Upper Kabul River basin.
- Estimation of flow for the Upper Kabul River basin using SRM Model.
- Calibration and validation of the SRM model for the Upper Kabul River basin

2.1 GENERAL

This chapter encompasses relevant presents a review of related literature for snowmelt runoff modeling.

2.2. MODELING OF SNOWMELT RUNOFF

Hydrological simulation model that includes snow are commonly divided into three basic components, snow cover area, precipitation runoff relationship and runoff distribution and routing procedure. Snowmelt runoff simulation models generally consist of a snowmelt model and a transformation model (WMO,1996). The snowmelt model produce liquid water from the snowpack that is available for runoff and the transformation model is an algorithm which converts the liquid output at the ground surface to runoff at the basin outlet. The snowmelt and transformation models can be distributed or lumped model in nature. Distributed models try to account for the spatial variability by dividing the basin into sub-basin and computing snowmelt runoff for each sub-basin separately with a set of parameters corresponding to each of the sub-basin. Lumped models consider whole basin as a single basin and use one set of parameter values to define the physical and hydrological characteristics. Generally distributed models use one of the following approaches to sub-divide a basin:

- Elevation zones or bands
- Basin characteristics such as slope, aspect, soil, vegetation etc. and
- A fixed or variable length, two or three-dimensional grid.

Distributed and lumped models are classified later by their use of energy balance approach or temperature index approach to simulate the snowmelt process.

2.3. SNOWMELT OPERATIONAL MODELS

A large numbers of models have been developed around the world over past four decades to describe snowmelt runoff. These can be divided into two broad categories, statistical and physical. A statistical model uses statistical relationship between inputs and outputs. A physical model based on physical processes that is related to input and output. In turn, these models can be used in a lumped or distributed model. The two more common ways of sub dividing an area of interest for snowmelt modeling is into elevation zones, or into grid squares. Some of these models are briefly discussed in Table 2.1.

Table 2.1: List of snowmelt Models

Sl. No.	Models Name	Country which developed in	Developed by
1	National weather service River forecast system (NWSRFS).	USA	Anderson (1973)
2	Conceptual Runoff model for Swedish catchment (HBV).	Sweden	Bergstrom (1975)
3	Snowmelt Runoff Model (SRM)	Switzerland	Martinec (1975)
4	Point Energy/Mass Balance Model	USA	Anderson (1976)
5	Comprehensive Watershed Model (MOEHYDRO2)	Canada	Logan (1976)
6	Guelph Agricultural watershed storm-Event Runoff Model (GAWSER)	Canada	Ghate and whitely (1977)
5	University of British Columbia (UBC)	Canada	Quick and Pipes (1977)
6	Water Resources Branch Model (WRB)	Canada	Kite (1978)
7	systems Hydrologic European snow Model(SHE)	France	Morris and Godfrey (1978)
8	Institute of Hydrology Distributed Model (IHDM)	UK	Morris (1983)
9	Precipitation-Runoff Modeling system(PRMS)	USA	Leavesley et.al. (1983)
10	Hydrologic Simulation Program-Fortran (HSP-F)	USA	Johanson et al. (1984)
11	Revised Model of Watershed Hydrology(USDAHl-74)	USA	WMO (1986)
12	SCS snowmelt Model (SCS)	USA	WMO (1986)
13	Storm Water Management Model (SWMM)	USA	WMO (1986)
14	U.S. Geological Survey Model (USGS)	USA	WMO (1986)
15	Continuous simulation and real-time forecast Model (QFORECAST)	Canada	WMO (1986)
16	CEQUEAU	Canada	WMO (1986)
17	Empirical Regressive Model (ERM)	Czechoslovakian	WMO (1986)
18	NEDBOR-AFSSSTROMNINGSMODEL (Rainfall-Runoff Model v.11)	Denmark	WMO (1986)

19	Tank Model with snow Model (TANK)	Japan	WMO (1986)
20	YETI	Czechoslovakian	WMO (1986)
19	SCHNEE	GDR	WMO (1986)
20	Winter season Runoff Model (WSRM)	Poland	WMO (1986)
21	Hydro Resources Optimization (IIRO)	USA	WMO (1986)
22	Model of Snowmelt Formation Lowland Rivers (GMTS-1)	USSR	WMO (1986)
23	Model of snowmelt formation in a mountainous Basin(GMTs-2)	USSR	WMO (1986)
24	Model of snowmelt-Rainfall Runoff Formation (GMTs-3)	USSR	WMO (1986)
25	Hydrologic Engineering center-1 (HEC-1)	USA	USACE (1990)
26	Streamflow simulation and Reservoir Regulation (SSARR)	USA	USACE (1991)
27	Simple Lumped Reservoir Parametric Model (SLURP)	USA	Kite (1995)
28	Soil Water Assessment Tool (SWAT)	USA	Amold et al. (1998)

2.4. COMPARISON OF SNOWMELT RUNOFF MODELS

The world meteorological organization (WMO) organized an international comparison of snowmelt runoff models (WMO 1986) in which hundreds of model runs were performed in six selected test basins. Table 2 compares the numerical results of each model in the WMO project.

Table 2.2: Results of model performance in the WMO project (10 years, snowmelt season)

Model	$D_{v\,max}$ %	D_v %	$1\text{-}R^2$	1-DQ	$1\text{-}R^2_{min}$	$1\text{-}DG_{min}$
A UBC	-23.2	8.13	0.269	0.370	1.898	1.533
B CEQ	-24.7	7.37	0.334	0.558	1.371	2.337
C ERM	65.7	15.34	0.695	0.906	5.831	3.226
D NAM	51.0	10.85	0.308	0.487	1.750	2.803
E TANK	45.9	7.90	0.239	0.356	1.223	2.286
F HBV	23.2	6.82	0.286	0.501	0.955	4.489
G SRM	-27.9	5.97	0.189	0.292	0.391	0.786
H SSARR	25.4	7.30	0.294	0.510	0.756	1.447
I PRMS	24.2	10.58	0.367	0.570	0.898	2.516
J NWS	28.1	7.43	0.230	0.254	0.684	1.239

Figure 1 shows a summary of all numerical values of R^2, DG (Coefficient of Gain) and D_v published by WMO (WMO 1986). DG and R^2 have the same formula, but D_G uses the average measured discharge from a number of past years whereas R^2 uses the average measured discharge of the particular year used in the simulations. Each prism refers to a tested model. The length along the x-axis corresponds to the arithmetic mean of all $(1-R^2)$ values, length along the y-axis to the arithmetic mean of all (1-DG) values, and length along the z-axis to the arithmetic mean of all D_v values as achieved in the snowmelt season of 10 test years.

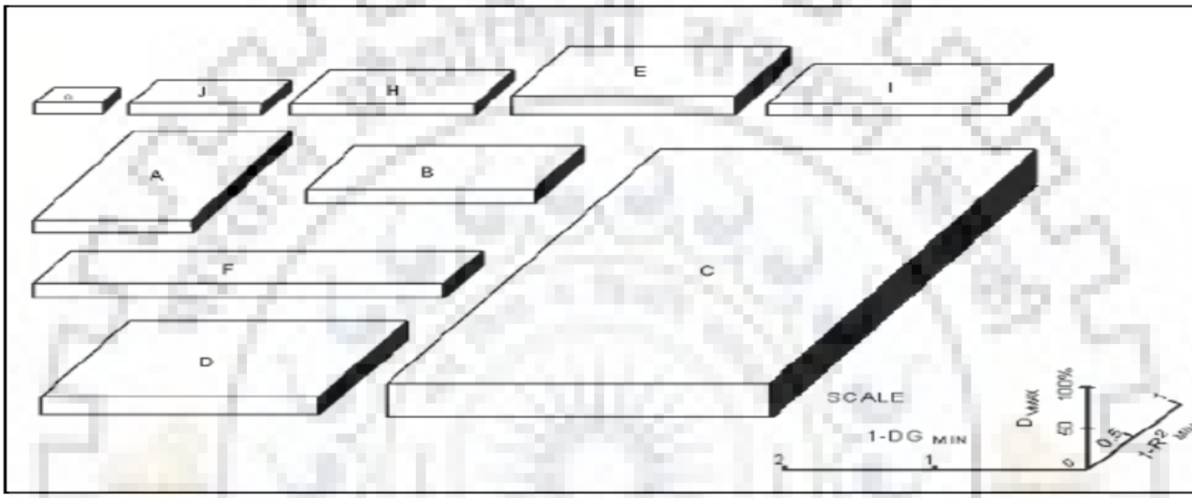

Figure 1.1: combined representation of model performance using three criteria: R2, DG and Dv. the volumes of the prisms indicate the maximum inaccuracies of the tested models form all results as listed for snowmelt seasons and individual years in the WMO

2.5. SNOWMELT COMPUTATION

The snowmelt component of snowmelt runoff simulation models usually takes the form of an energy balance or a temperature index to simulate the process of melting. the fist approach is known as energy and mass balance approach and the second is the degree days' approach. These two approaches are discussed below:

2.5.1 ENERGY AND MASS BALANCE APPROACH

The energy and mass balance approach of snowpack governs the production of meltwater. This method describes the incoming energy, outgoing energy and the change in energy storage for a snowpack of a given period of time. The net energy is expressed as equivalent of snowmelt. The energy balance equation can be written as follows (Equation 2.1):

$$\Delta Q = Q_n + Q_e + Q_h + Q_g + Q_n \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (2.1)$$

Where:

Q_n = net radiation (long and short wave)

Q_e = latent heat transfer

Q_h = Sensible heat transfer

Q_g = is the ground snow interface heat transfer.

Q_m = heat transfer by mass changes

ΔQ = change in heat storage.

In the above equation, various elements of energy are considered in the form of energy flux, defined as the amount of energy received on a horizontal snow surface of unit area over unit time. The positive value of (Qm) will result in the melting of snow. The relative importance of different heat transfer includes in melting of a snowpack depending on time and local conditions. For example, radiation melting influence the weather conditions when wind is quite. Melting due to sensible heat flux influence under warm weather conditions. When all the elements of energy balance equation are defined, the melt rate due to energy flux can be expressed as (Equation 2.2):

$$M = \frac{Q_m}{\rho w * l * \beta} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (2.2)$$

Where,

M = meltwater (m/day)

L = latent heat of fusion (333.5 kJ/kg)

ρw = density of water (1000kg/m3)

β = thermal quality of snow

the thermal quality of snow depends on the temperature of the snowpack and the amount of free water content (3-5%). For a snow that is thermally ripened for melting and contains about 3% of free water content, the value of (β) is 0.97 for such cases (2.2) reduces to, (Equation 2.3):

$$\mathbf{M = \frac{Q_m}{[1000*333.5*0.97]}} \qquad \textbf{(2.3)}$$

Which leads to simple relationship, (Equation 2.4):

$$\mathbf{A = 0.0031 * Q_m(\frac{mm}{day})} \qquad \textbf{(2.4)}$$

Data required to evaluate equation (2.1) is measurements of air temperature, albedo, wind speed, vapor pressure and incoming solar radiation (Anderson, 1973). These data are difficult to obtain on a basin scale and extrapolation to areal values form point data is another problem, especially the spatial detail is required for distributed models. This becomes further difficult when such data is required for a highly rugged terrain, such as Upper Kabul River. As such application of the energy balance equation is usually limited to small, well-instrumented or experimental watersheds.

2.5.2 DEGREE-DAY APPROACH OR TEMPERATURE INDEX APPROACH

The specific type of data required for the energy budged method is rarely available for carrying out the snowmelt studies. The is particularly true for the Upper Kabul river basins where the network for data collection is poor. The commonly available data in Upper Kabul river are daily maximum and minimum temperatures, humidity measurements and surface wind speed. This is why the temperature indices are widely used in the snowmelt estimation. It is generally considered to be the best index of the heat transfer processes associated with the snowmelt. Air temperature expressed in degree-days is used in snowmelt computations as an index of the complex energy balance tending to snowmelt. A 'degree-day' in a broad sense is a unit expressing the amount of heat in terms of persistence of a temperature for 24-hour period of one-degree centigrade departure from a reference temperature. The simples and the most common expression relation daily snowmelt to the temperature index is, as follows (Equation 2.5):

$$\mathbf{M = D(Ti - Tb)} \qquad \textbf{(2.5)}$$

Where,

M = melt produced in mm of water in a unit time

D = degree-day factor ($mm^0\ c^{-1} day^{-1}$)

Ti = index air temperature (^{0}c)

T_b = base temperature (usually 0 ^{0}C)

Daily mean temperature is the most commonly used index temperature for snowmelt. The mean temperature is computed by (equation 2.6):

$$\mathbf{T_i = (T_{mean} = (T_{max} + T_{min})/2)} \quad \mathbf{(2.6)}$$

There are several methods of dealing with the index temperature used in calculating the degree-day value. When using the maximum –minimum approach, the most common way is to use the temperature. It was reported that sometimes the degree-days form the daily mean temperature are found to be misleading. In many parts of the western U.S. mountainous areas, the drop in minimum temperature is so much that the daily mean temperature comes to below 0 0C, thereby indicating no possible degree-days, whereas snowmelt conditions have prevailed during a part of the day when air temperature were much above the freezing point. The inclusion of minimum temperatures at an equal weight with the maximum temperature gives undue emphasis to this effect. On the other hand, the use of maximum temperature only excludes this effect entirely. In order to counteract such problems, alternatives have been suggested in which unequal weight to the maximum and minimum temperature are given. U.S. Army crops on Engineers (1956), used the following index temperatures (Equation 2.7):

$$\boldsymbol{T_i = (2\mathrm{T_{max}} + \mathrm{T_{min}})/3} \quad \mathbf{(2.7)}$$

Another approach is given by (equation 2.8):

$$\boldsymbol{T_i = T_{max} + (T_{min} - T_{max})/b} \quad \mathbf{(2.8)}$$

Where,

(B) is a coefficient less than 2

When the basin is subdivided based on elevation zones, the degree-days are extrapolated to an elevation zone by using a suitable lapse rate i.e., (Equation 2.9):

$$\mathbf{T_{ij} = \delta(h_{st} - h)} \quad \mathbf{(2.9)}$$

Where,

T_{ij} = degree-day of the elevation zone

δ = temperature lapse rate in ^{o}c per 100m

H = zonal hypsometric mean elevation in m.

h_{st} = altitude of the temperature statin in m.

in a basin with little seasonal variation, a lapse rate of 0.65 ^{0}c/100 m has been found to be suitable.

2.5.3 CLIMATE CHANGE STUDIES

Snowmelt studies, where the simulation of flow have been conducted in the context of different climatic scenarios, are (Nemec and Schaake, 1982; Gleick, 1986; McCabe and Ayers, 1989; Schaake and Liu, 1989; lettenmaier and Gan, 1990; Vehvilainen and Lohvansuu, 1991; Panagoulia, 1991; Arnell, 1992; Ng and Marsalek, 1992; Cayan et al., 1993; Chiew et at., 1995; singh and Kumar, 1997; Mehrotra, 1999; Baron et al., 2000; Xu, 2000; Fontaine et al., 2001; Singh and Bengtsson, 2004).

2.6. APPLICATION OF SNOWMELT RUNOFF MODEL

Due to the limitations of the data availability in the Upper Kabul River Basin, distributed simulation models are difficult to apply satisfactorily. However, SRM which readily uses the rainfall and temperature data, were effectively applied in this area. Several studies have been carried out to test the applicability of the SRM Model.

Leavesley (1989) studied about development of improved model capabilities and establishment of standardized techniques and measures to evaluate model performance and results.

Singh (2003) applied the SRM for Beas basin up to Manali for a limited period. The results of the model were found satisfactory with a goodness of fit as 083 and 0.61 for the years 1978 and 1979 respectively.

Thomas (1999) studied the Hydrologic modelling of a snow fed River Basin. The study involves the development of a model, calibration and it use for simulation of a snowmelt runoff. The percentage seasonal difference in volume is 1.46% and Nash-Sutcliffe goodness of fit value is 0.85 and the percentage volume difference is -5.42%.

Singh and Jain (2002) studied the glacier and snow melt in the Satluj river at Bhakra Daim in the Western Himalayan area. They have found that the average contribution of snow and glacier runoff in the annual flow of the Satluj river is about 59%, and 41% from rain.

Singh and Jain (2003) applied the SNOWMOD model in the Satluj basin to simulate the snowmelt runoff and daily stream flow for the years 1985 to 1999. Basin elevation was divided into 10 bands. Results shows that on annual basis average snowmelt runoff contribution was 67.8%.

Vajja and Partha (2005) estimated the snowmelt runoff in beas Basin at Pandoh dam by Using SRM. They founded that good agreement between the observed and computed runoff with a coefficient of determination of 0.854 and the difference in volume is +4.6%.

Arora et al., (2007) applied SNOWMOD to study the Chenab basin at salal dam for climate variability influences on hydrological responses. They found that, melt is much more sensitive to increase in temperature than to rainfall. Therefore, the project receives major share of melt contribution in the runoff. The projects have to be designed keeping in view the magnitude of variability in flows expected due to changing climate conditions.

Alam et al., (2011) Estimated the Snow Runoff Using Snowmelt Runoff Model (SRM) in A Himalayan Watershed. The simulation of SRM revealed the volume difference of only 7.8 percent between measured runoff (79 m^3/sec) and computed runoff (72.9 m^3/sec).

2.7. REMOTE SENSING OF SNOW COVER

Snow cover area has long been known as very important hydrologic variable for predicting the streamflow and also important variable for climate and hydrologic models due to its effect on energy and moisture budget. The presence of snow in a basin heavily affects moisture that is stored at the surface and is available for future runoff. For hydrological application, estimation of SCA is primarily important where spring runoff is strongly affected by winter snow accumulation. Remote sensing is a valuable tool for snow cover mapping which can be used for predicting snowmelt runoff. Remote sensing provides a new and valuable tool for obtaining snow data to predict snowmelt and runoff. Initially, snow data is obtained manually using snow courses that are very labor intensive, costly and potentially dangerous. The recent use of snowboard pillow telemetry and storage loading measurements have reduced the need for some fieldwork, however. The problem of the point measurement has not overcome. A single point measurement may or may not be representative of a large area or basin. From the perspective of remote sensing, snow cover is one of the most recognizable criteria for water resources from aerial photography or satellite imagery. Present operation satellite systems are limited to determine only the area of snow cover, snow depth and snow water equivalent. Snow physical parameters cannot by measured directly by theses system Currently, a suite of satellite for snow cover products are available:

Table 2.3: Sensor band responses relative to various snowpack properties (Rango, 1993)

Snow property	Sensor band			
	Gamma rays	Visible/near Infra-red	Thermal Infra-red	microwaves
Snow covered area	Low	High	Medium	High
Depth	Medium	If very shallow	Low	Medium
Water equivalent	High	If very shallow	Low	High
Stratigraphy	No	No	No	High
Albedo	No	High	No	No
Liquid water content	No	Low	Low	high
Temperature	No	No	Medium	low
Snowmelt	No	Low	Low	Medium
Snow-soil interface	Low	No	No	High
Additional factors All weather capability Current best spatial resolution From space platform	No Not possible	No 10 m	No 100 m	No 25 km passive 3 active

2.7.1 GAMMA RADIATION

The water content of some snow packets can be measure by low-flying aircraft carrying sensitive gamma-radiation detectors. This method uses natural emission of low level gamma radiation. Radioactive isotopes of potassium, uranium and thallium are naturally found in a typical soil. The plane or Aircraft is flying over the same flight line before and during the snow cover measuring the thickness of the snow layer, which empirically relates to an average snow water equivalent for the site (Carrol and Vanais, 1980). This method is limited to low air craft altitudes (approximately 150 meters) because it reduces the atmosphere of significant particles of radiant energy. This restriction effectively limits the use of gamma detection to relatively flat areas and cannot be used in mountainous areas due to safety considerations. Also, this approach is also limited to non-forested areas, because the impact of forest biomass to weaken the radiation signal (Glynn et al., 1988). However, past work in measured snow water equivalent by making a correction that is based on the amount of biomass and the type of radiation. In addition, sine gamma energy is relatively low, the maximum depth of snow equivalent is limited to 30-40 mm, and data interpretation can difficult if the background soil moisture changes throughout the season (Vershinina, 1985). However, Carrol and Vose (1984) have been reported in a forest environment with 480 snow water equivalent of 480 mm. in the NOAA operational airborne gamma radiation snow water-mapping program, procedure for correcting for the soil moisture were included in the system (Carroll and Carroll, 1989). This operational program flew over 1400 flight lines in the united states and Canada (Carool and Carroll,1989).

2.7.2 VISIBLE/NEAR INFRARED

Albedo is a snow-covered property that is easily measured by remote sensing. Albedo, A, is defined as:

A= Reflected solar radiation/ incoming solar radiation.

Typically, new snow will have an albedo of 90% or more whereas older snow that have been weathered and has accumulated dust and litter can have an albedo as low as 40% (Foster eto al., 1987). The reflector depends on the snow properties, such as grain size and shape, water content, surface roughness, depth and presence of impurities. The reflection of the new snow is reducing d due to its age in both the visible and infrared regions, however, the reduction in the infrared region is more prominent. This increase in sensitivity the infrared is due to an increase in the grain size of the snow that result in milting and refreezing. In most cases, the reflection can decrease in the visible area it is attribute to pollutants such as dust, pollen and aerosols. Most current snowmelt models use either point snow course estimates of the snowpack or the total area of the snowpack. In the latter case there is usually an implicit assumption that the snow cover area and its changes are somehow consistently related to the snow water equivalent in the basin. Through careful analysis of satellite images or aerial photograph, snow can be identified and the boundaries of the snow/ no snow areas accurately located. However, a certain amount of subjective interpretation may be necessary to identify and separate the effects of shadows and forests.

In non-forested terrain, all areas with continuous brightness distinctly greater that the normal dark background, and that have been assure dot be cloud free, should be mapped as snow areas. the snow line that encloses these areas can be assume to represent accumulated snow depths of 2.5 (Bowley *et al.*, 1981) or more. Areas on imagery that appear mottled (alternating dark and light reflectance) can be mapped as areas of between zero and 2.5 cm of snow depth. In mountainous terrain, the snow line is mapped with visible bands of satellite imagery because of its high reflectance in comparison with no- snow areas. Generally, this means selectin the NOAA VHRR visible channel, Landsat Mss channels 4 or 5, SPOT, or Landsat TM Channels 2 and 4. If there is a choice of bands, it is better to choose a spectral band closer to the infrared region because at higher sun angles the Landsat MSS band 4 and TM band 1 and other bands near the blue region may be near saturation, causing a loss of detail in identifying snow and now-snow areas. Although snow can be detected at longer wavelengths, that is in the near infrared, the contrast between a snow and a no-snow area is considerably lower than with the visible region of the spectrum. However, the contrast between clouds and snow is greater in Landsat Tm Band 5 (1.57-1.78µm) and this serves as a useful discriminator between clouds and snow (Dozier, 1984).

2.7.3 THERMAL INFRARED

Thermal data may be the least of the common use of the remote sensing products to measure snow and its properties. In order to determine the snowpack temperature, the amount of spectral radiation of the snow must be known. This in turn requires the knowledge of liquid water content and grain size as well as other factors. Despite these constraints thermal data can be useful for identification snow/ snowless boundaries. Thermal infrared data is also useful for discriminating between clouds and snow with AVHRR data because the 1.57-1.78 µm band is not available on this sensor. The emissivity of snow approaches that of a black body at the 10.5

– 12.5 µm atmospheric window. Griggs (1968) has shown that melting snow can have an emissivity as high as 99%, whereas the emissivity for no-snow areas is typically 95% or less.

2.7.4 MICROWAVE

Microwave remote sensing offers several benefits that are not provided by other satellite sensors. This is because microwave data can clear information about snow properties that are of great interest to hydrologists; that is; snow cover area, snow water equivalent (or depth) and the presence of liquid water in the snowpack which signals the onset of melt (Kunzi et al., 1982). Snow on the surface of the earth, in other words, is the accumulation of crystals or ice grains, which makes it possible to have a closed snow, which in one area may completely or partially cover the earth. The physical properties of the snowpack determine its microwave properties; the microwave radiation released from the underlying ground is dispersed by snowflakes in many ways within the snow layer, as a result, a microwave release above the snow level is less than the emission of the earth. Effective properties included in the microwave response of the snowpack: depth and water equivalent, liquid water content, density, grain size and shape, temperature and stratification as well as snow state and land cover. The sensitivity of microwave radiation to the snow layer on the ground provides snow monitoring using passive microwave, remote sensing techniques to extracting information on snow extent, snow depth, snow water equivalent (SWE) and snow state (wet/dry). Because the number of scatter within a snowpack is proportional to the thickness and density, equivalent of snow water can be seen at the lighting temperature of the scene (Hallikainen and Jolama, 1986). Deeper snowpacks generally result in lower lighting temperatures. Active microwave observation of the snowflakes is, very few and almost not-being, especially about 10 cm and less wavelengths are suitable for snow. Most of information about microwave remote sensing of snow is made from a truck or tower system. Preliminary results by stiles et al. (1981), matzler and Schanda (1983) and Rott (1986) determine that active microwave region has a potential similar to the passive microwave region. The analysis of active microwave data is more complex than passive data due to the confusion caused by the effect of surface properties (including soils) and geometry consideration on the reflected radar wave. A significant advantage of the active microwave approach is an improved resolution capability (of about 10 m form space) when compared with the passive.

2.7.5 MODERATE RESOLUTION IMAGING SPECTRORADIOMETER (MODIS)

With the launch of MODIS in December 1999, a new era in hyper spectral satellite remote sensing is started. MODIS makes it possible to monitor the environment by measuring atmospheric spectral ranges from the blue to thermal infra-red.

The first MODIS sensor went into orbit with the launch of the TERRA satellite on December 18, 1999. With the successful launch of AQUA from Vandenberg Air Force Base, CA, on 4th May 2002, a second MODIS sensor was put into orbit for studying the Earth's water cycle and our environment. TERRA and AQUA (both with a 705km orbit) have a sun-synchronous, near polar, circular orbit. AQUA will cross the equator daily at 1:30 P.M. as it heads north (ascending mode). With this formation it is expected that AQUA's afternoon observations combined with TERRA's morning observations will provide important insights into the daily cycling of global precipitation and ocean circulation.

MODIS is a 36 band spectrometer providing a global data set every 1-2 days within 16-days repeat cycle. The spatial resolution of MODIS (Pixel size at nadir) is 250 m for channel 1 and 2 (0.6 μm-14.4 μm), respectively Figure 2.4

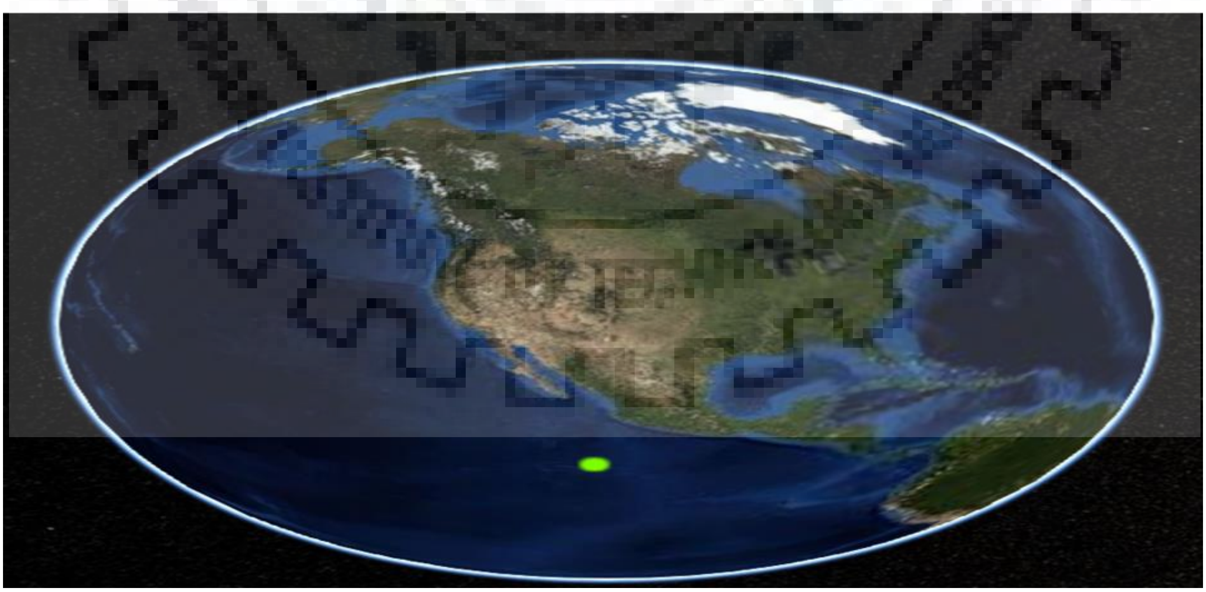

Figure 2.2: MODIS maximum snow cover extent during an eight –day period in 1200 km x 1200 km tiles

Satellite images are made by combining the reflected light detected by the sensor at various wavelengths (spectral bands) and making them into a single image. The MODIS Rapid Response system makes use of MODIS broad range of spectral observation by creating both true-color and false-color images each tailored to highlight different land surface, atmospheric, and oceanic features. One such band combination is 7-2-1. In this composite, MODIS bands 7, 2 and 1 are assigned to the red, green, and blue portions of the digital image. Vegetation appears bright green and bare soil red, and water appears dark black in this combination leaving snow light blue very prominent and hence making it easy to distinguish snow form all three classes i.e. vegetation, bare soil, and water. MODIS rapid response system provides on its website a number of image subsets that are automatically generated in near-real-time for various applications users.

3.1. DESCRIPTION OF THE UPPER KABUL RIVER BASIN

This study has been carried out for the Upper Kabul River basin, Afghanistan. The basin is bounded between longitude 68^0 38' 41.31" E and latitude 34^0 30' 42.3" N. Upper Kabul River basin is sub divided in 3 sub basin i.e, Maidan , Nerkh and Durrani basins. The study area starts from 75 km west Kabul City, from Paghman Mountain range and belongs to the HinduKush Mountain range in Afghanistan. It takes off from the west slope of Maidan and from the south slope of Nerkh and joined together at Maidan and flows nearly north-east direction up to Kabul City. The area of the Upper Kabul basin is 1654.2 km^2. Kabul river is one of the greatest river, it is 700-kilometer-long river of which 560 km are in Afghansita. Rising in Upper Kabul River basin in the Sanglakh range of the HiduKush Mountains in Afghanistan. It flows east past Kabul and Jalalabad, and is separated from the watershed of the Helmand River by the Uani pass, the Major tributaries of the Kabul River are the Logar, Panjshir, Alingar, Surkhab, Kunar. The Kabul river empties into the Indus river near Attock, Pakistan. It is the main river in eastern Afghanistan. The Kabul river crosses two major climatic belts. It upper reaches have a continental warm-summer climate with a mean July temperature of about (25 ^oC) and mean January temperature reaches to (-8 ^oC). The gradient of the river is very steep in upper reaches and in lower reach it become less steep. There are high peaks in the west as well as in the south of the river valley and its altitudes varies from 1800m near Kabul to more than 4500 m. During summer these tributaries are mainly fed by snowmelt and have got perennial flow, which varies considerably during different months of the year. A major portion of this basin is covered by mountains which receive a heavy snowfall in winter and less amount of rainfall in summer and minor portion of the basin is lies under degraded forests and cultivated land and there for the proportion of the silt and sand are of fine, medium and coarse configuration. Less Steep slopes are common but are terraced at several places in the lower ranges up to an elevation of 1800 m for an agricultural purpose. There are one hydrological and one meteorological station in the study area (Figure 3.1). The discharge data from Tangi saydan hydrological station 34^0

51’52.7” N and 69^0 4’ 37.2” E, precipitation and temperature data from Qalaye Malak station has been obtained.

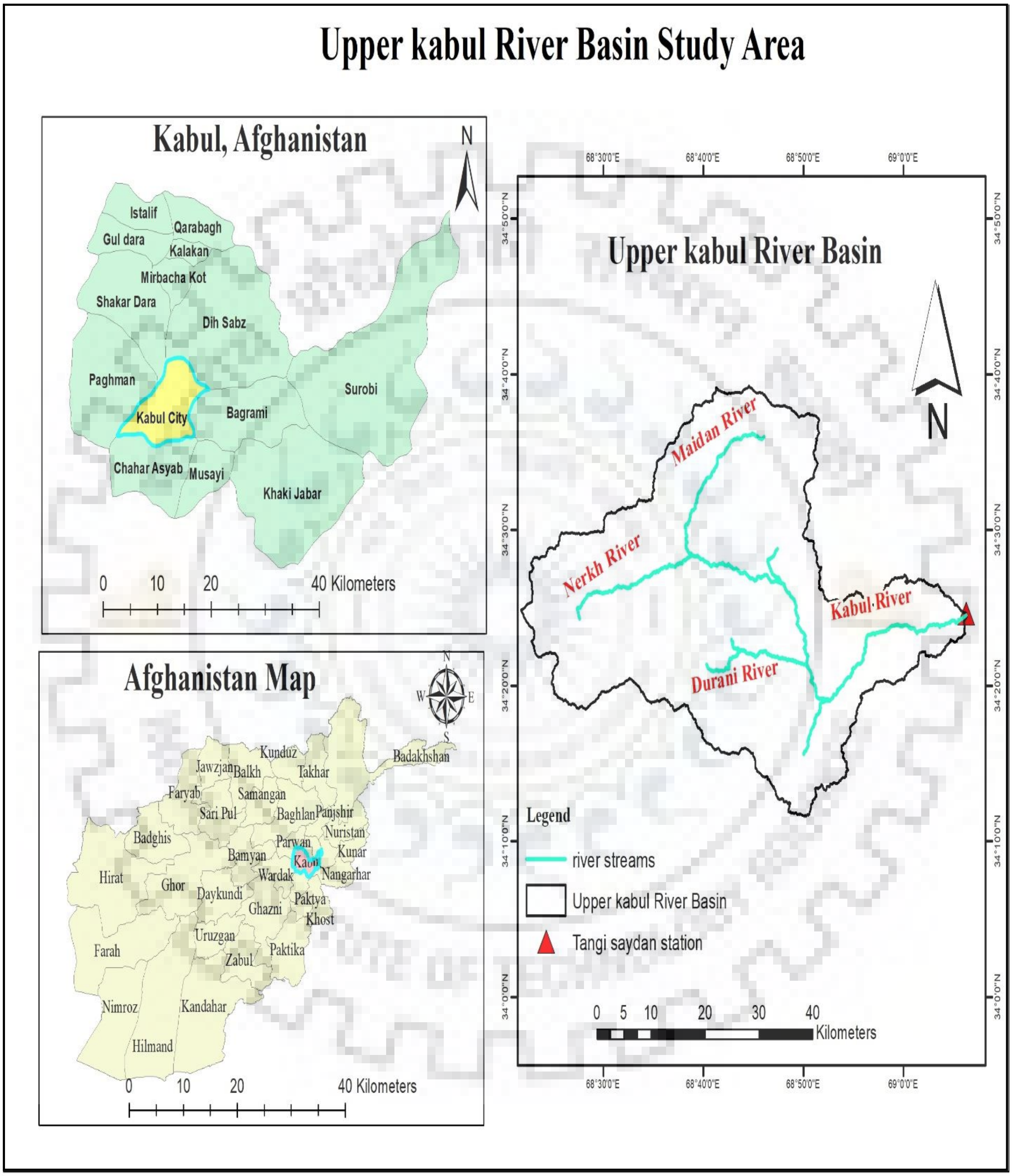

Figure 3.1: Location map of the study area

3.2. PRECIPITATION

The Upper Kabul River basin receive snowfall during winter season, generally extending from December to March, but less rainfall in summer season. The summer season is generally marked by high river flows and sometimes flood. There are no regular records of the snowfall experienced in the upper Kabul areas and the exact amount of precipitation on this account cannot be assessed with any degree of certainty. The climatic conditions vary greatly in the different parts of the basin area. Variation are related to changes in elevations and aspect. The climate of the study area cold in the winter and hot in summer season.

3.3. RUNOFF

The runoff in the Kabul river comprises of two parts, the one that is derived from the melting of the snow and the other resulting from the rainfall in the basin. The runoff which received from the snowmelt makes the river perennial and the runoff due to the snowmelt varies from year to year due to the changes caused by the annual variations in temperature and extent of snowfall in the catchment area during any year. The discharge in the Kabul river is measured at the gauge and discharge station established at Kabul, Afghanistan.

3.4. GEOLOGY AND SOIL

The geology of the Upper Kabul River described as valley-fill basin and range setting Figure 3.2. Where the valleys are filled with Quaternary and Tertiary sediments and rocks. And the ranges are composed of uplifted crystalline and sedimentary rocks. Quaternary sediments are typically less than 80m thick in the valleys. The underlying Tertiary sediments have been estimated as much as 800m thick in the area and may be more 1000 thick in some areas of the valley. Based on some afghan and soviet efforts, the divided Quaternary and Tertiary sediments and rocks into younger and older basin deposits. The younger deposits described as gravel and sand and talus. The gravel and sands were deposited mainly in the river channel. And the older deposits are the Lataband, Kabul and the Buthkahk series. The latband series includes gravels and conglomerate ranging in thickness from several meters to several hundred meters. The Kabul series consist of marls, clays, siltstones and fine grained sandstones. The surrounding mountains are primarily composed of Paleoproterozoic gneiss and Late Permian through Late Triassic sedimentary Rocks.(Mack et al. 2009)

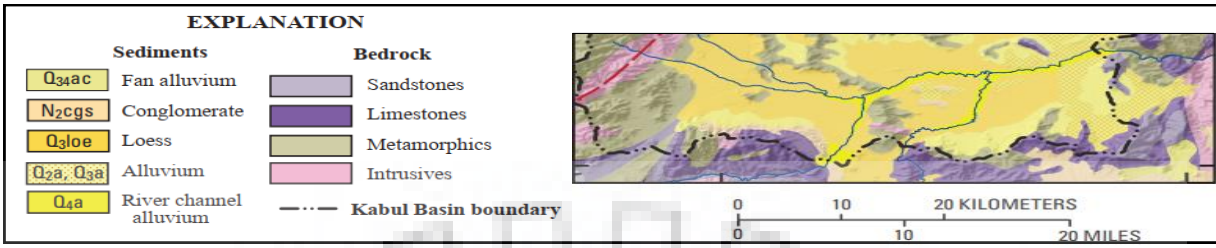

Figure 3.2: Geology of the Upper Kabul River Basin

3.5. REMOTE SENSING DATA

3.5.1 SNOW COVER DATA FOR THE UPPER KABUL RIVER BASIN

In this study MODIS data have been selected and used for preparation of snow cover maps and for Topography, SRTM DEM has been used (Figure 3.3). The maximum and minimum snow covered are determined using remote sensing data (MODIS) for all the months.

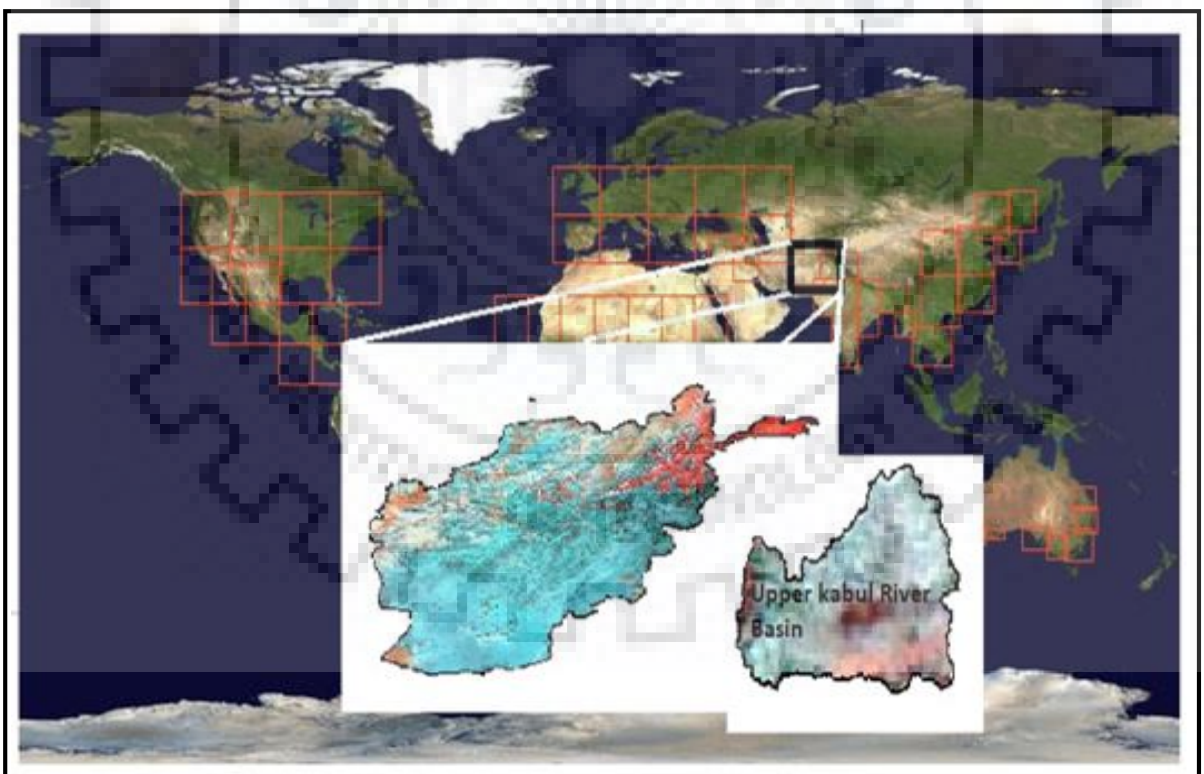

Figure 3.3: MODIS subset of Afghanistan covering part including the study area of Upper Kabul

The archive imagery is available online, images for the years 2015, 2016, 2017 have been downloaded from MODIS rapid response system website (http://rapidfire.sci.gsfc.nasa.gov/subsets).

3.5.2. MODIS DATA

The moderate Resolution imaging Spectroradiometer (MODIS) lunched in December 1999, produces a snow- covered area (SCA) Products. a new are in hyperspectral satellite remote sensing began. MODIS makes it possible to monitor the environment by measuring atmospheric trace gases and aerosol density, and mapping the surface of clouds, land and sea in a variety of spectral ranges from the blue to the thermal infra-red.

In this study, we have been used MOD10A2 images have been used , the summary of MOD10A2 data is listed in table 3.1. These Satellite images are made by combining the reflected light detected by the sensor at various wavelengths (spectral bands) and making them into a single image. The MODIS rapid response system make use of MODIS broad range of spectral observation by creating both true-color and false-color images, each tailored to highlight different land surface, atmospheric, and oceanic features, one such band combination is 721. In this composite, MODIS bands 7, 2, and 1, are assigned to the red, green, and blue portions of the digital image. Vegetation appears bright green and bare soil red, and water appears dark black in this combination leaving snow light blue very prominent and hence making it easy to distinguish snow from the others classes.

Table 3.1: Summary of the MOD10A2 Products

Earth science data type (ESDT)	Product Level	Nominal data dimensions	Spatial resolution	Temporal resolution	Map projection
MOD10A2	1.3	1200kmx1200km	500m	8-day	GCTP sinusoidal

3.5.3 SRTM DIGITAL ELEVATION DATA

For the first time, space shuttle Radar Topographic Mission (SRTM) carried onboard February 19, 2000. It was successfully Mapped the topographic features of Earth's land masses using radar interferometry (leblance et al., 2006). A radar interferogram is produced by measuring the radar phase difference between two spatially separated antennas, A1 and A2 (Zebker et al.m 1994). The near global coverage DEM was produced from the C-band data and processed by NASA's JPL and the X-band data provided slightly higher resolution and were processed by the German space agency's aerospace center (DLR). This method requires no ground control, and hence is very useful for inaccessible regions. The overall absolute horizontal and vertical accuracy of these 1 arc second data is estimated to be significantly better than the original mission requirement of 20 and 16m respectively. The spatial resolution of the SRTM DEMs is the WGS84 EGM96 geoid. This data is currently distributed free of charge by USGS and is available for download from the National Map Seamless data distribution system, or the USGS ftp site (http://edscns17.cr.usgs.gov/srtm/index.html). The DEM files have been mosaicked into a seamless global coverage, and are available for download as $5^0 *5^0$ tiles, in geographic coordinate system- WGS84 datum. These files are available for download in both Arc-info ASCII format and as GeoTiff, for easy sue in most GIS and remote sensing software applications. The DEM of the study area is shown in Figure 3.4.

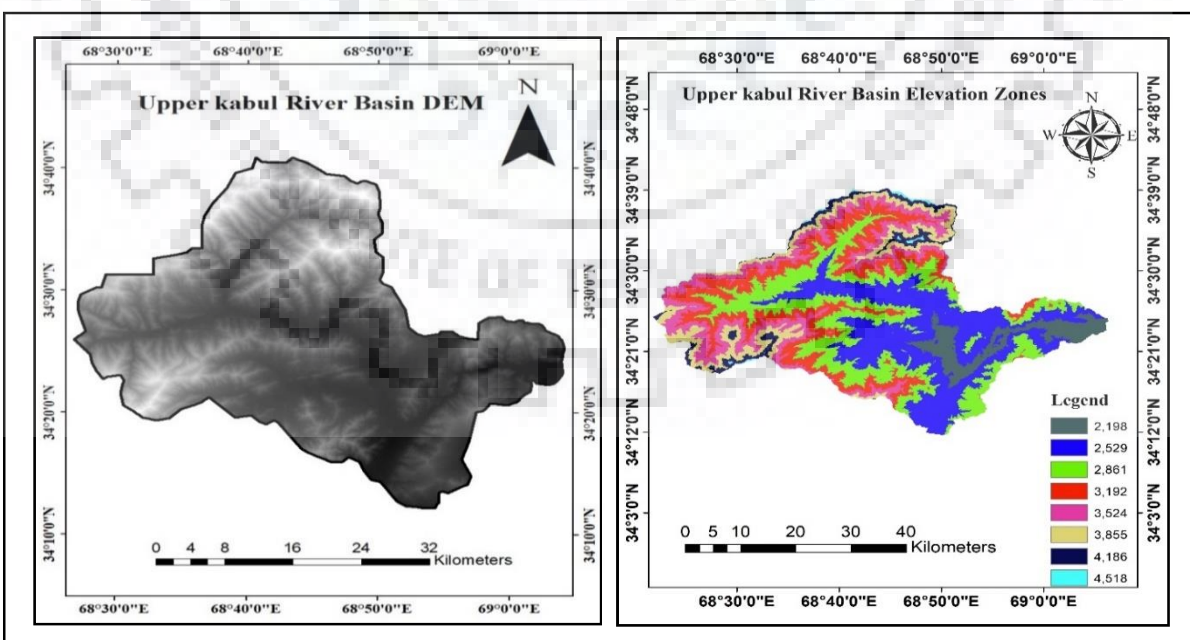

Figure 3.4: DEM and the elevation zones of the Upper Kabul river basin

4.3. SNOWMELT RUNOFF MODELING

In most part of the world the seasonal short-term variation in streamflow reflects the vacation in rainfall. But in higher latitude and altitudes where snowfall is predominant, runoff depends on heat supplied for snowmelt rather than the timing of precipitation. Hence, to understand the hydrological behavior and simulate the stream flow it is very important to model the snowmelt runoff. one of the objectives of this study is to simulate the snowmelt runoff in Upper Kabul River basin

Modeling of streamflow from basin is based on the transformation of incoming precipitation to outgoing streamflow by considering losses to the atmosphere, temporary storage, lag and attenuation, in most part of the world the seasonal short-term variation in streamflow reflects the variation in rainfall. But in higher latitude and altitudes where snowfall is predominant, runoff depends on heat supplied for snowmelt rather that the timing of precipitation. Hence, to understand the hydrological behavior and simulate the stream flow it is very important to model the snowmelt runoff. One of the objectives of this study is to simulate the snowmelt runoff in Upper Kabul River basin located in the upper part of Kabul city. This chapter provides details of snowmelt model (SRM) a conceptual hydrological model that is used in the study. The structure of the model, data input, procedure used for the simulation of the snowmelt runoff are described and simulation results are presented.

Over the last few decades a wide range of hydrological models were used for various application from purely statistical methods which neglects the physics of snowmelt process to the complicated energy budget equation but the most important and popular among them are the conceptual models, which represent a compromise between scientifically realistic complexity and practically realistic simplicity because of the difficulties in obtaining input data varying in time and space. (Kumar, 2009).

The conversion of snow into water is called snowmelt, which needs input of energy (heat). Hence the process of snowmelt is linked to the flow and storage of energy into and through the snowpack. The data required to run an energy balance model for snowmelt runoff estimation needs information on air temperature, albedo, solar radiation, wind speed and vapor pressure. The main difficulties are to obtain such data in basin scale and extrapolate the point data in area values. Another difficulty is to obtain such data for highly rugged terrain like Upper Kabul river basin. Hence application of energy balance equation is limited to small and well networked watersheds.

The conceptual model fully employs the concept of an index where a known variable is used to explain a phenomenon is a statistical rather than in a physical sense. The most commonly available data for any basin is air temperature and it is considered the best index of heat transfer processes associated with snowmelt estimation. There are several temperature index based snowmelt models like SRM, SNOWMOD, SSARR Model, PRMS Model. Every one of these model is used for different purpose. The SRM has been widely used for snowmelt modeling for the different basin in different countries. The snowmelt runoff model (SRM) uses precipitation, temperature, snow cover area as input variables, to simulate the seasonally or yearly runoff which generates form snow and rain. Kabul river has a major contribution of snow and rain during summer season. in this study, we carried out SRM model for the upper Kabul River basin, to find out the runoff for the Kabul River.

4.4. DESCRIPTION OF SNOWMELT RUNOFF MODEL (SRM)

The Snowmelt Runoff model (SRM) has been developed by Martinec (1975) in small European basins and gradually this is used by many researches. SRM can be applied in mountain basins of almost any size (so far from 0.76 to 917,444 km^2) and any elevation range. SRM is a temperature index model, which is designed to simulate and forecast daily streamflow for mountainous basins, where snowmelt is a major runoff factor and having contribution from both snowmelt and rainfall. The generation of stream flow from such basins involves. The determination of the input derived from snowmelt and rain, and its transformation into runoff. It is a distributed model and for simulating the streamflow, the basin is divided into a number of elevation zones and different hydrological processes related to snowmelt and rainfall runoff are evaluated for each zones. A model run starts with a known or estimated discharge value and can proceed for an unlimited number of days, as long as the input variables - temperature, precipitation and snow covered area - are provided. As a test, a 10-year period was computed without reference to measured discharges (Martinec and Rango, 1986). In addition to the input variables, the area-elevation curve of the basin is required. If other basin characteristics are available (forested area, soil conditions, antecedent precipitation, and runoff data), they are useful for facilitating the determination of the model parameters.

SRM can be used for the following purposes:

(1). Simulation of daily flows in a snowmelt season, in a year, or in a sequence of years. The results can be compared with the measured runoff in order to assess the performance of the model and to verify the values of the model parameters. Simulations can also serve to evaluate runoff patterns in ungauged basins using satellite monitoring of snow covered areas and extrapolation of temperatures and precipitation from nearby stations.

(2). Short term and seasonal runoff forecasts. The computer program (WinSRM) includes a derivation of modified depletion curves which relate the snow covered areas to the cumulative snowmelt depths as computed by SRM. These curves enable the snow coverage to be extrapolated manually by the user several days ahead by temperature forecasts so that this input variable is available for discharge forecasts. The modified depletion curves can also be used to evaluate the snow reserves for seasonal runoff forecasts. The model performance may deteriorate if the forecasted air temperature and precipitation deviate from the observed values, but the inaccuracies can be reduced by periodic updating.

(3). In recent years, SRM was applied to the new task of evaluating the potential effect of climate change on the seasonal snow cover and runoff. The microcomputer program has been modified and supplemented accordingly.

4.5. SRM METHODOLOGY

Snow melting and runoff Model is used for estimating the runoff due to snowmelt. The stepwise methodology is illustrated below:

- The river basin of interest area (Upper Kabul River basin) has been selected and extracted from the SRTM data through GIS.
- Snow covered area of the (Upper Kabul River basin) has been extracted from the series of MODIS image and analyised through ERDAS IMAGINE.
- Snow Depletion curves were derived from the snow covered area of the Upper Kabul River basin
- In a selected elevation zones of the basin, snow cover period was estimated.
- Hydrological and meteorological data of the upper Kabul river basin was analyzed.
- SRM model Variables like temperature and precipitation are obtained from the above meteorological data.
- Temperature lapse rate was recorded from the temperature data.
- Model calibration was carried out for the period of 2016-2017.
- Sensitivity analysis of SRM model was carried out.
- Model validation was carried out for year 2016-2017.
- Runoff has been predicted using the SRM model.
- Model accuracy was also checked.

flow chart of the SRM model with all its input parameters as well as output are illustrated in Fig. 4.1.

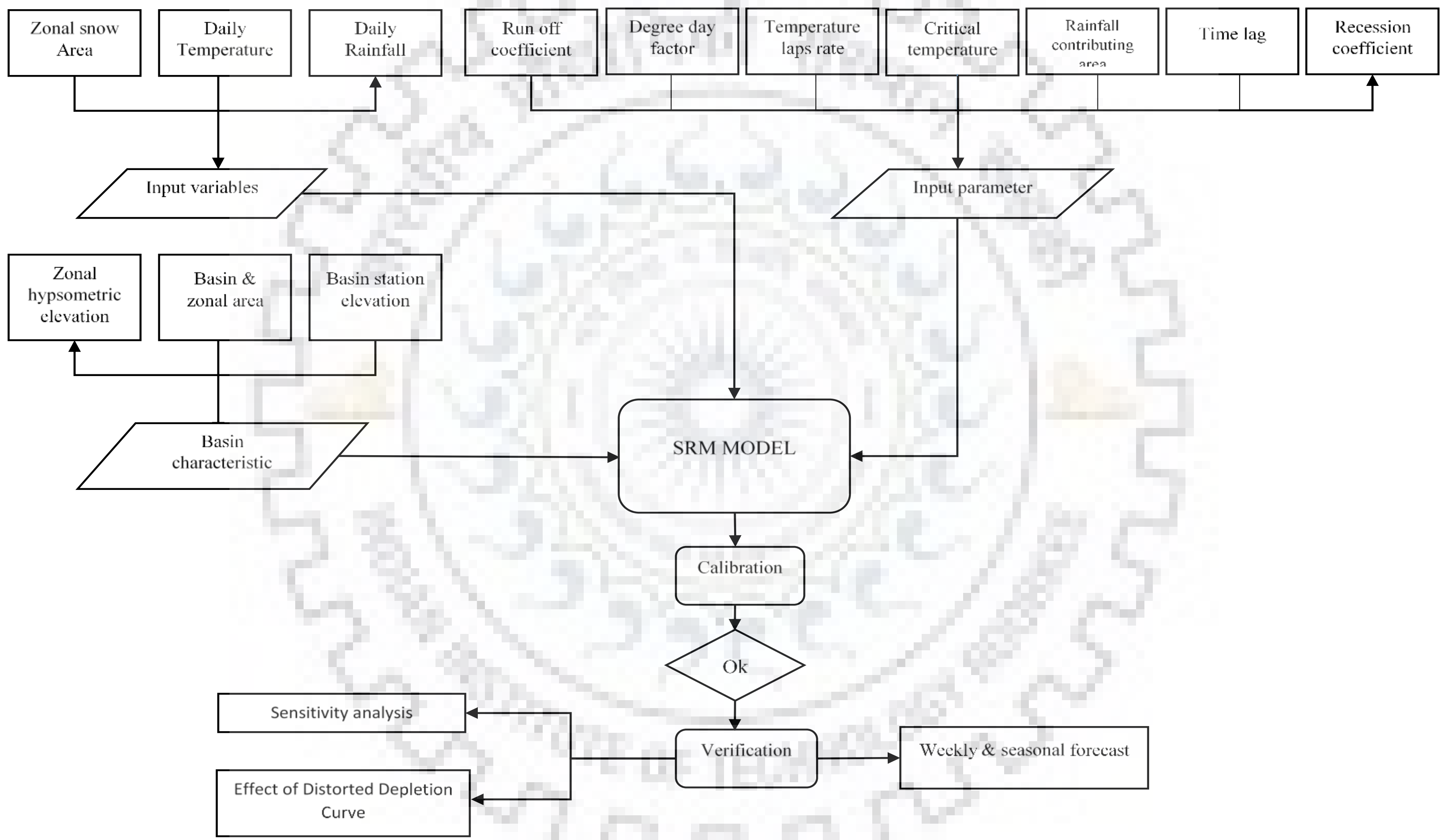

Figure 4.1: Flow chart of Research Methodology.

4.6. GENERAL EQUATION OF SRM MODEL

$$Q_{n+1} = [C_{Sn}a_n(T_n + \Delta T_n)S_n + C_{Rn}P_n] + \frac{A.1000}{86400}(1 - k_{n+1}) + Q_nK_{n+1} \ldots\ldots\ldots\ldots\ldots\ldots (4.1)$$

Where,

Q = Average daily discharge [$m^3\ s^{-1}$]

C = Runoff coefficient expressing the losses as a ratio (runoff/precipitation), with C_S referring to snowmelt and C_R to rain.

a = Degree-day factor [cm oC-1d-1] indicating the snowmelt depth resulting from 1 degree-day.

T = Number of degree-days [^oC d]

ΔT = The adjustment by temperature lapse rate when extrapolating the temperature from the station to the average hypsometric elevation of the basin or zone [^oC d]

S = Ratio of the snow covered area to the total area

P = Precipitation contributing to runoff [cm]

K = Recession coefficient indicating the decline of discharge in a period without snowmelt or rainfall.

k = Q_{m+1}/Q_m (m, m + 1) are the sequence of days during a true recession flow period).

N = sequence of days during the discharge computation period. Equation (1) is written for a time lag between the daily temperature cycle and the resulting discharge cycle of 18 hours. In this case, the number of degree-days measured on the nth day corresponds to the discharge on the n + 1 day. Various lag times can be introduced by a subroutine.

$\frac{\mathbf{10000}}{\mathbf{86400}}$ = conversion from cm·$km^2\ d^{-1}$ to $m^3\ s^{-1}$

(T, S and P) are variables to be measured or determined each day, C_R, C_S, lapse rate to determine ΔT, T_{CRIT}, k and the lag time are parameters which are characteristic for a given basin or, more generally, for a given climate.

If the elevation ranges of the basin or watershed exceed 500 m, the basin should by subdivided into elevation zones of about 500m each. The elevation range of the upper Kabul River baisn is from 1800 m to 4500 m and eight elevation zones A, B, C, D, E, F, G, H, the model equation becomes:

$$
\begin{aligned}
\mathbf{Q_{n+1}} = \Big\{ & \left[C_{SAn}\, a_{An}(T_n + \Delta T_{An}) S_{An} + C_{RAn} P_{An} \right] \frac{A_A * 100}{86400} \quad \text{for zone 1} \\
+ & \left[C_{SBn} a_{Bn} (T_n + \Delta T_{Bn})\, S_{Bn} + C_{RBn} P_{Bn} \right] \frac{A_B * 100}{86400} \quad \text{for zone 2} \\
+ & \left[C_{SCn}\, a_{Cn} (T_n + \Delta T_{Cn}) S_{Cn} + C_{RCn} P_{Cn} \right] \frac{A_C * 100}{86400} \quad \text{for zone 3} \\
+ & \left[C_{SDn} a_{Dn} (T_n + \Delta T_{Dn}) S_{Dn} + C_{RDn} P_{Dn} \right] \frac{A_D * 100}{86400} \quad \text{for zone 4} \\
+ & \left[C_{SEn} a_{En} (T_n + \Delta T_{En}) S_{En} + C_{REn} P_{En} \right] \frac{A_E * 100}{86400} \quad \text{for zone 5} \\
+ & \left[C_{SFn} a_{Fn} (T_n + \Delta T_{Fn}) S_{Fn} + C_{RFn} P_{Fn} \right] \frac{A_F * 100}{86400} \quad \text{for zone 6} \\
+ & \left[C_{SGn} a_{Gn} (T_n + \Delta T_{Gn}) S_{Gn} + C_{RGn} P_{Cn} \right] \frac{A_G * 100}{86400} \quad \text{for zone 7} \\
+ & \left[C_{SHn} a_{Hn} (T_n + \Delta T_{Hn}) S_{Hn} + C_{RHn} P_{Cn} \right] \frac{A_H * 100}{86400} \Big\} (1 - K_{n+1}) \\
+ & \; Q_n K_{n+1} \} \qquad \text{for zone 8} \cdots\cdots\cdots\cdots \quad (4.2)
\end{aligned}
$$

Equation (9) and (10) are written for the eight elevation zones of the upper Kabul river basin and only for metric system of the SRM model but for the English units. The user of the SRM model can change the equation from metric to English units.

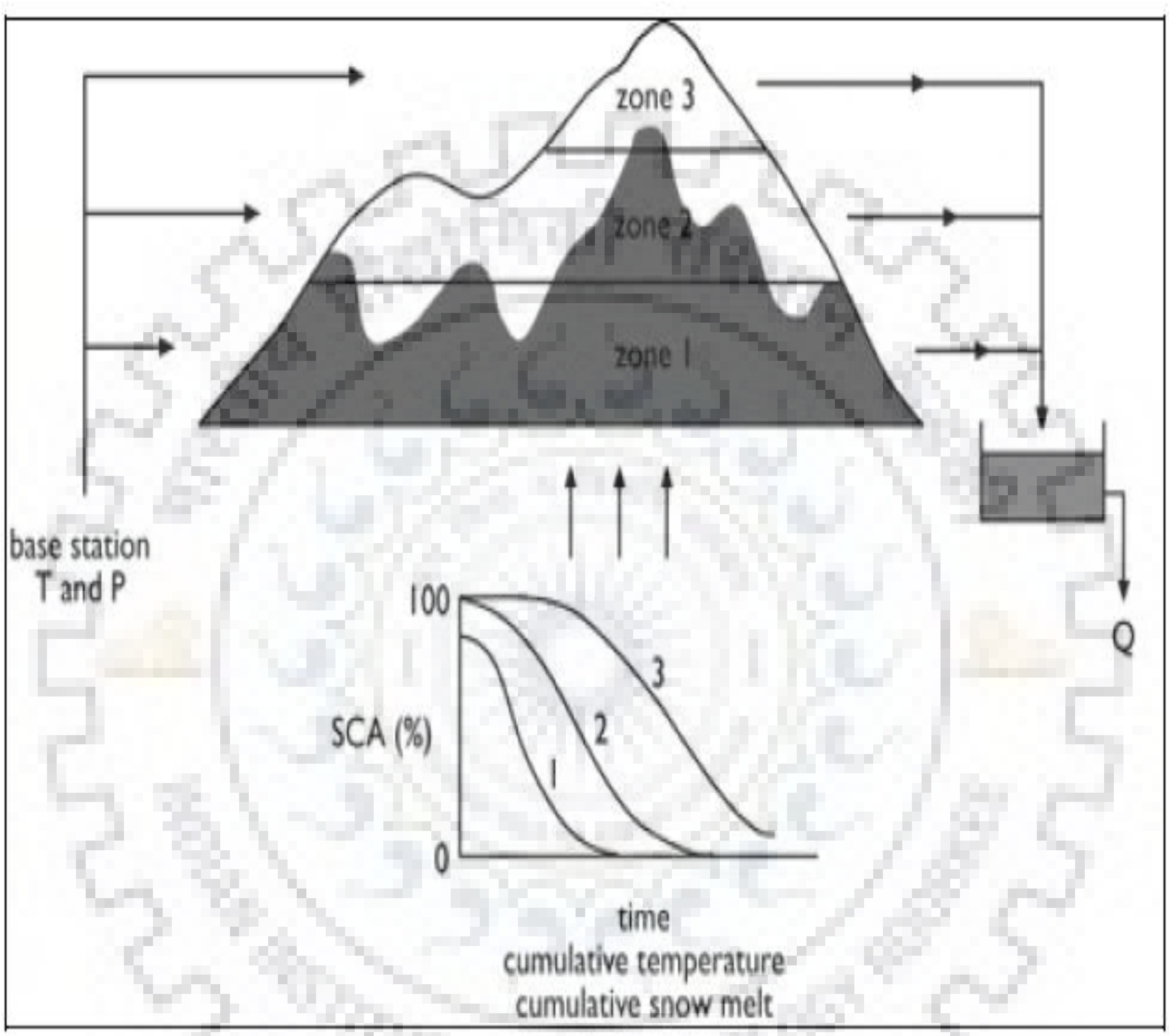

Figure 4.2: Structure of snowmelt runoff model

4.7. DATA REQUIREMENT

There are three input tables needed to run the model.

1. The Features of the basin with inputs are of each elevation zone and hypsometric mean elevation of that zone, that can be obtained from area elevation curve of the study area.

2. The snowfall cover, precipitation, and temperature are input variables.

3. Basin parameters which are: snow runoff coefficient, critical temperature, runoff coefficient for rain, rainfall contributing area, degree day factor, recession coefficient, temperature lapse rate and time lag. These parameters are used to calibrate the model.

In SRM model, one of the important input is snow covered area. In this study, this inputs have been prepared using remote sensing data and the methodology is described in the following section.

4.8. DETERMINATION OF MODEL VARIABLES

4.8.1. DIVISION OF THE WATERSHED INTO ELEVATION BANDS

The basin boundary is defined by the watershed divide and the location of the stream gauge as identified in the topographic map. In the mountainous watersheds where the the snow depth and temperature vary with elevation, the basin area is divided into number of the elevation zones and each elevation band is treated as a separate watershed with its own characteristics and initial snow water equivalent. The number of zones will depend on the topographic relief of the basin. There is no specific range of altitude for slicing or dividing the basin in the zones. But an altitude difference of about 500 m or so is considered appropriate for dividing the basin into elevation zones. Once the elevation zones are defined, the various model variables and parameters are applied to each zone for the calculation of snowmelt runoff. To facilitate this application, the mean hypsometric elevation of the zones must be determined through the use of area-elevation curve.

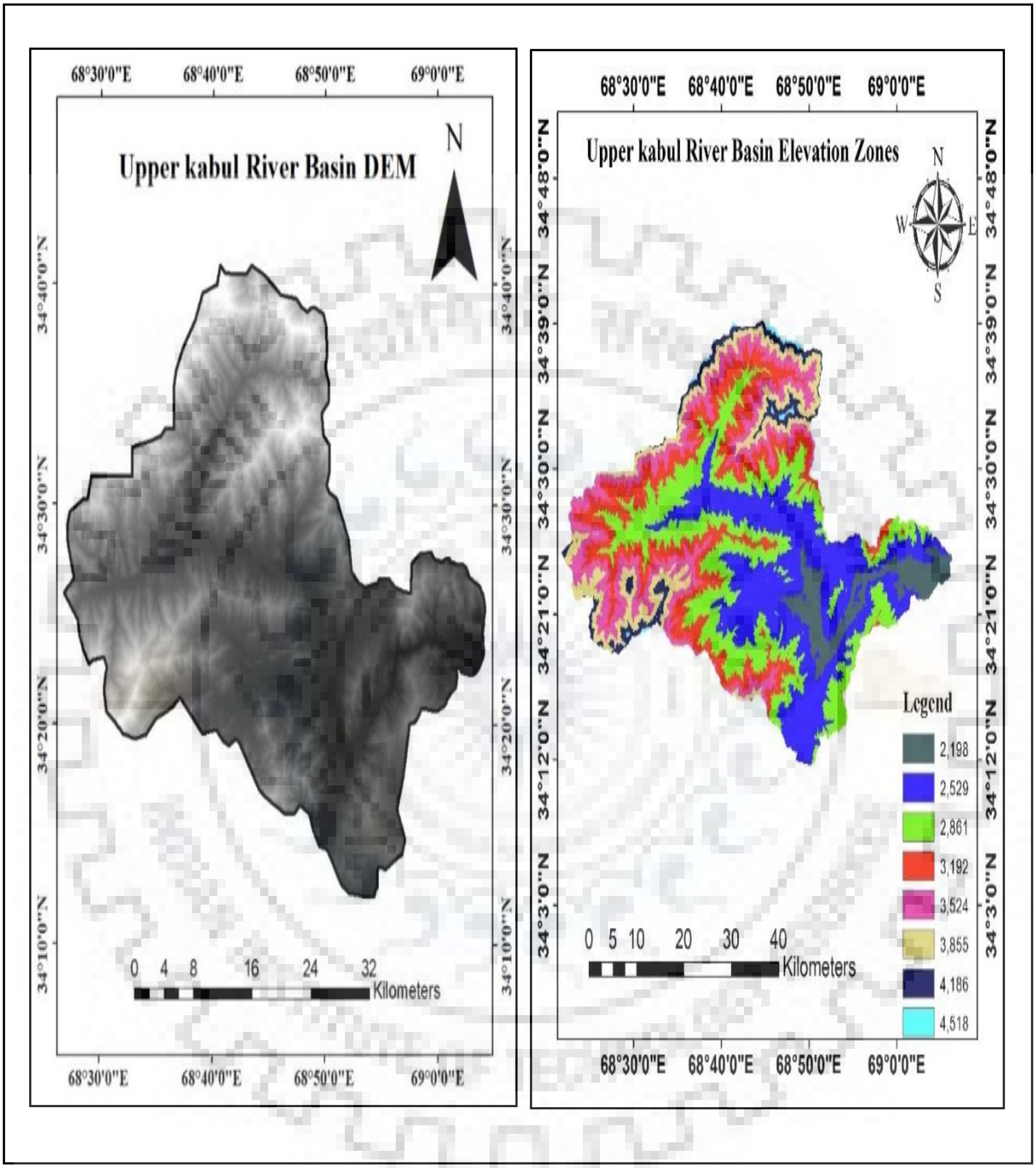

Figure 4.3: DEM and the elevation zones of the Upper Kabul river basin

4.8.2. AREA- ELEVATION CURVE OF THE BASIN

For the preparation of the area-elevation curves, the basin should be divided into number of elevation bands. By using the zone boundaries and other selected contour lines in the basin, the areas enclosed by the various elevation contours are determined either by palnimeter or by image processing system if the digital-elevation data is available. the area of the catchment falling in each elevation band is measured and the area elevation curve of the basin is prepared. The zonal mean hypsometric elevation (h) is then determined from this curve, by balancing the areas above and below the mean elevation. The (h) value is used as the elevation to which the base station temperatures are extrapolated for the calculation of zonal degree-days.

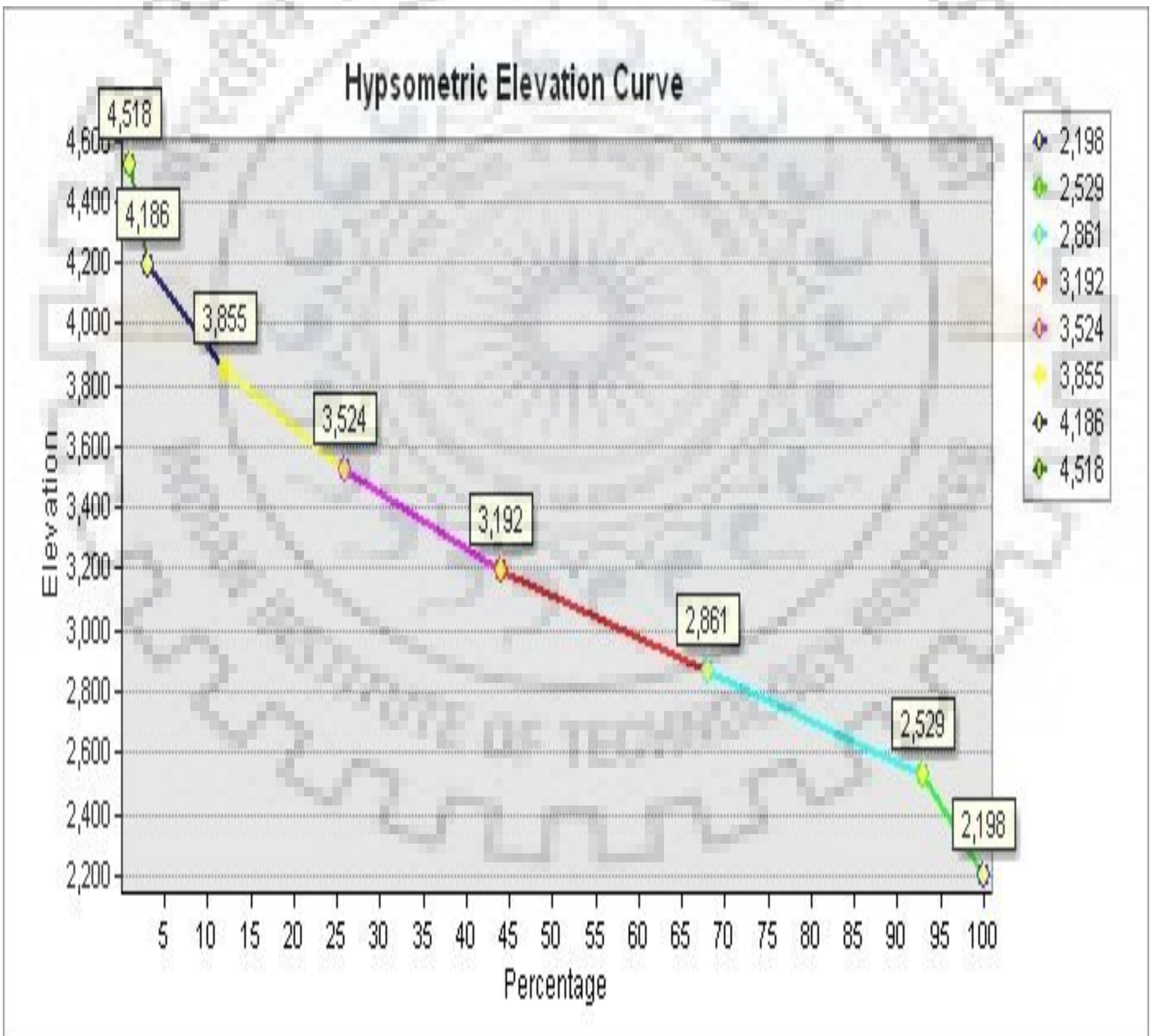

Figure 4.4: Hypsometric elevation curve of the basin.

4.8.3. DISTRIBUTION OF TEMPERATURE ON ELEVATION ZONES.

At least one temperature station is required in the basin, but if more temperature stations are available in different elevation zones, the condition will be more ideal as it will minimize the vertical distance to which the temperature would have to be extrapolated and will be applied to the model. Generally, in mountainous basins, due to the rugged relief of topography and other limiting conditions there are very few hydro-meteorological stations. Therefore, the data available at only one temperature station is to be extrapolated or interpolated to the mid-elevation of each zones, using predefined temperature laps rate in the model. Lapse rates are known to be quite variable. The lapse rate during a continuous rainstorm will approximate the saturated adiabatic lapse rate, whereas under clear sky and dry weather conditions, the lapse rate during the warm part of the day will tend to be quite low, as the radiation cooling will cause the temperatures to fall to the dew point temperature and at times even zero lapse rates may occur.

The daily temperature of the upper Kabul river basin has been calculated in the various elevation zones by using the temperature lapse rate approach, by extending data from the base station by the following equation,

$$\boldsymbol{\Delta T = \gamma * \left(h_{st} - \bar{h}\right) * \frac{1}{100}} \quad \textbf{(4.3)}$$

Where,

$\boldsymbol{\gamma}$	=	temperature lapse rate (^{0}C per 100 m)
$\boldsymbol{h_{st}}$	=	altitude of the temperature station (m)
$\bar{\mathbf{h}}$	=	hypsometric mean elevation of a zone (m)

Whenever the degree-day numbers (T + ΔT) in equation (1) become negative, they are automatically set to zero so that no negative snowmelt is computed.

The temperature lapse rate must be carefully selected and it should be indicative of the mountainous region where the basin is located.

- **Temperature input Program options:**

0 = basin wide

1= by zone

The snowmelt runoff model (SRM) accepts either temperature data from a single station (option0, basin wide) or from several stations (option1, by zone). With option 0, the altitude of the station is entered a temperature data are extrapolated to the hypsometric mean elevations of all zones using the lapse rate. So Upper Kabul River basin has one station, we selected option 0, the altitude of the station is (1870m).

4.8.4. PRECIPITATION

The evaluation of real precipitation is particularly difficult in mountain basin. The distinction between snow and rain for the basin is very crucial for all snowmelt models because precipitation falling in the form of snow and rain and snow behaves differently in terms of contribution to the stream flow. The contribution of rain to the stream flow is faster than snow because snow is stored in the basin area until melted, whereas rain is immediately processed. On any elevation zones, the precipitation may fall as snow or rain. Rain on zones or a part of the zones is added directly to the moisture input. Snow is added to the previously accumulated snow, if any. The new snow that falls on the previously snow covered area becomes part of the seasonal snowpack and its effect depends on the condition of the snowpack. The rain falling over a cold snow pack in the early melt season, will be frozen in the snow pack and would not be available immediately to the runoff. it would melt when favorable atmospheric and snow pack conditions are available. however, if the rain is falling over the ripe snow pack, it is transferred through the snow layer and contributes to runoff. when precipitation occurs in the basin, a critical temperature, t_c is to be selected to determine whether the precipitation is rain or snow. t_c is generally selected slightly above the freezing point and may vary from basin to basin.

- **Rainfall input program options:**

0 = Basin wide

1= by zone

The snowmelt runoff model accepts either a single, basin-wide precipitation input (from one station or from several stations, that is, option 0) or different precipitation inputs zone by zone (option 1). If the program is switched to option 1 and only one station happens to be available, for example in the zone A, precipitation data entered for zone A must be copied to all other zones. Otherwise no precipitation from these zones is taken into account by the program. In basin with a great elevation range, the precipitation input may be underestimated if only low altitude precipitation stations are available. it is recommended to extrapolate precipitation data to the mean hypsometric altitudes of the respective zones by an altitude gradient, for example 3% or 4% per 100 m.

Acritical temperature is used to decide whether a precipitation event will be treated as rain ($T \geq T_{CRIT}$) or as new snow ($T < T_{CRIT}$). When the precipitation event is determined to be snow, its delayed effect on runoff is treated differently depending on whether it falls over the snow-covered area is assumed to become part of the seasonal snowpack and its effect is included in the normal depletion curve of the snow coverage. The new snow falling over the snow- free area is considered as precipitation to be added to snowmelt, with this effect delayed until the next day warm enough to produce melting. this precipitation is stored by SRM and then melted as soon as a sufficient number of degree-days has occurred.

4.8.5. SNOW COVER AREA, S

It is a typical feature of mountain basins that the areal extent of the seasonal snow cover gradually decreases during the snowmelt season. Snow cover area has to be estimated from the satellite imageries / digital data as information is not available on the extent of the snow covered area in the different months during the snowmelt period. In the snowmelt runoff computations, attention is usually focused on determining the snowmelt rate. However, the continuously changing cover in the watershed may be overshadow the improved accuracy. The changing snow may be of little importance in flat areas with uniform snow cover or for forecasting the total seasonal snowmelt volume. On the other hand, it plays dominant role in the mountainous catchments where snow cover area gradually decreases from 100% of the watershed to zero. In watersheds with rugged terrain and sharp mountain ridges, the evaluation by planimetering is impossible as the consists of numerous dispersed snow patches. The snow cover area in each elevation zone is plotted against the elapsed time to construct the depletion curves for the various elevation zones in the basin. Depletion curves of the snow cover can be interpolated form periodical snow cover mapping so that the daily values can be read off as an important input variable to SRM. The minimum area which can be mapped with an adequate accuracy depends on the spatial resolution of the remote sensor. Table 4.1 indicates some of the possibilities of remote sensing for snow cover mapping.

Table 4.1: Various Satellites for the snow cover area mapping.

Platform Sensor	Spectral Bands	Spatial Resolution	Minimum area size	Repeat period
Aircraft orthophoto	Visible/NIR	2 m	1 km^2	Flexible
IRS				
Pan	Green to NIR	5.8 m	2 km^2	24 days
LISS-II	1-3 Green to NIR	23 m	2.5-5 km^2	24 days
WiFS	1 Red / 2 NIR	188 m	10-20 km^2	5 days
SPOT				
HRVIR	1-3 Green to NIR	2.5-20 m	1-3 km^2	26 days
Landsat				
MSS	1-4 green to NIR	80 m	10-20 km^2	16-18 days
TM	1-4 Green to NIR	30 m	2.5-5 km^2	16-18 days
ETM-Pan	Visible to NIR	15 m	2-3 km^2	16-18 days
Terra/ Aqua				
ASTER	1-3 visible to NIR	15 m	2-3 km^2	16 days
MODIS	1 Red / 2 NIR	250 m	20-50 km^2	1 day
	3-8 Blue to MIR	500 m	50 – 100 km^2	1 day
NOAA				
AVHRR	1 Red / 2 NIR	1.1 km	10-500 km^2	12 hr
Meteosat				
SEVIRI	1-3 Red to NIR	3 km	500-1000 km^2	30 min
	12 Visible	1 km	10-500 km^2	30 min

Acronyms:

- ASTER= Advanced space borne thermal Emission and Reflection Radiometer.
- AVHRR = Advanced Very High Resolution Radiometer.
- HRVIR = high Resolution Visible and Near Infrared
- IRS = Indian Remote Sensing
- LISS = Linear imaging Self- scanning Sensor
- MIR = Middle Infrared
- MODIS = Moderate Resolution Imaging spectroradiometer
- MSS = Multi- Spectral Scanner
- NIR= Near Infrared
- Pan Panchromatic
- SEVIRI = Spinning Enhanced Visible and Infrared Imager
- SPOT = Satellite Pour Observation da La Terre

- TM = Thematic Mapper
- WiFS= Wide Field Sensor   ETM-Pan = Enhanced Thematic Mapper - Panchromatic

4.9.METHODOLOGY

The methodology followed in this study consisted of preparing the DEM of the study area, preparing snow cover map, details of the methodology follow:

4.9.1. PREPARATION OF SNOW COVER MAP

MODIS snow cover map are used as inputs for mapping seasonal snow cover. Terra MODIS snow cover 8- day 1.3 global 500m grid (MOD10A2) data from December to March between 2015 and 2017 were acquired.

The Normalized Difference Snow Index (NDSI) uses the spectral characteristics f snow and is based on the concept of Normalized Difference Vegetation Index (NDVI) used in vegetation mapping from remote sensing data (Dozier, 1989: Hall et al., 1995, Gupta et all., 2005). The NDSI is defined as the difference of reflectance observed in a visible band and the short-wave infrared band divided by the sum of the two reflectance (Gupta et al., 2005). It can be computed by:

$$\mathrm{NDSI} = \frac{\text{Visible Band} - \text{SWIR Band}}{\text{Visible Band} - \text{SWIR Band}}$$

the Modis Snow cover product is a classified image, the pixel values of the MOD10A2 data in the study area include 1 (No decision), 25 (snow free land), 50 (Cloud obscured), 200 (snow) and 254 (Detector saturated) (Riggs, Hall, and Salomonson 2015). The surface land cover on pixels denoted (No decision) and detector saturated cannot be determined, so these pixels are with those of (cloud obscured). Thus, the surface land cover of snow cover maps is recoded into two classes in this study: snow and snow-free land. Using the classified snow maps, the total percentage of snow cover in this study area has been estimated for different dates.

Snow cover mapping is a difficult task in this research. Because eight days' satellite images are downloaded and classified to snow and non-snow cover classes. This leads to processed of almost 50 satellite images. Modis images are available in JPEG format as subsets of different regions of the world. The following steps are involved in getting snow cover from an image.

- Downloading MODIS subsets (Upper Kabul river basin, Afghanistan) from the MODIS satellite websites.
- Rectification of the satellite image using the available corner coordinates form info file with the image.
- Classification of each image into snow and snow free classes.
- Converting the classified to thematic map.
- Converting the elevation zones of the DEM to the thematic map
- Intersect the classified snow covered and non-snow covered thematic map to the DEM elevation zone thematic map.
- Extraction of snow covers for each elevation zones.

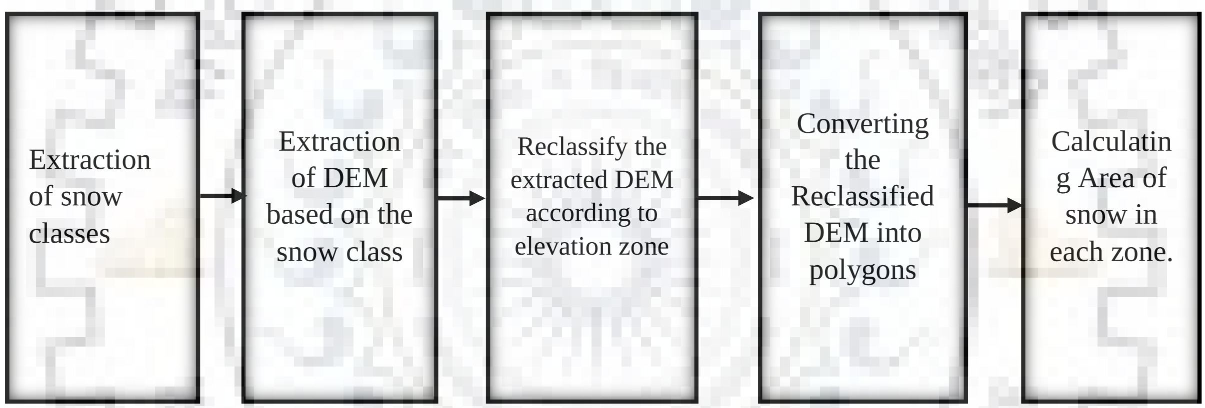

Figure 4.5: The main steps for obtaining the snow cover area using ERDAS and Arc GIS tools.

4.9.2. EXTRACT BY ATTRIBUTE TABLE

After classification of the image of the catchment into snow and snow-free classes, only snow class is extracted from the image using the option extract by attribute table. Extract by attribute extract the cells of a raster, based on a logical query e.g. where “Class name” = ‘snow'

4.9.3. RESAMPLE

The extracted snow class map has a 500*500 m spatial resolution, whereas the Digital Resolution Model has 90 m spatial resolution. It is very important to extract thematic DEM covered only with snow on a specific day, because the extracted DEM then can be used to extract snow for each elevation zone. Because of that resample tool is used. Resample tool alters the proportions of a raster dataset by changing the cell size. The cell size will be changed but the extent of the raster would remain same.

4.9.4. EXTRACT BY MASK

Now the resampled image can be used to extract DEM covered only with snow. extract by mask too in ArcGIS extracts the cells of a raster that correspond with the areas defined by a mask which in this case is the resample image extraction of snow. this way DEM where snow exists is extracted and which will be used for extracting snow in each elevation zone.

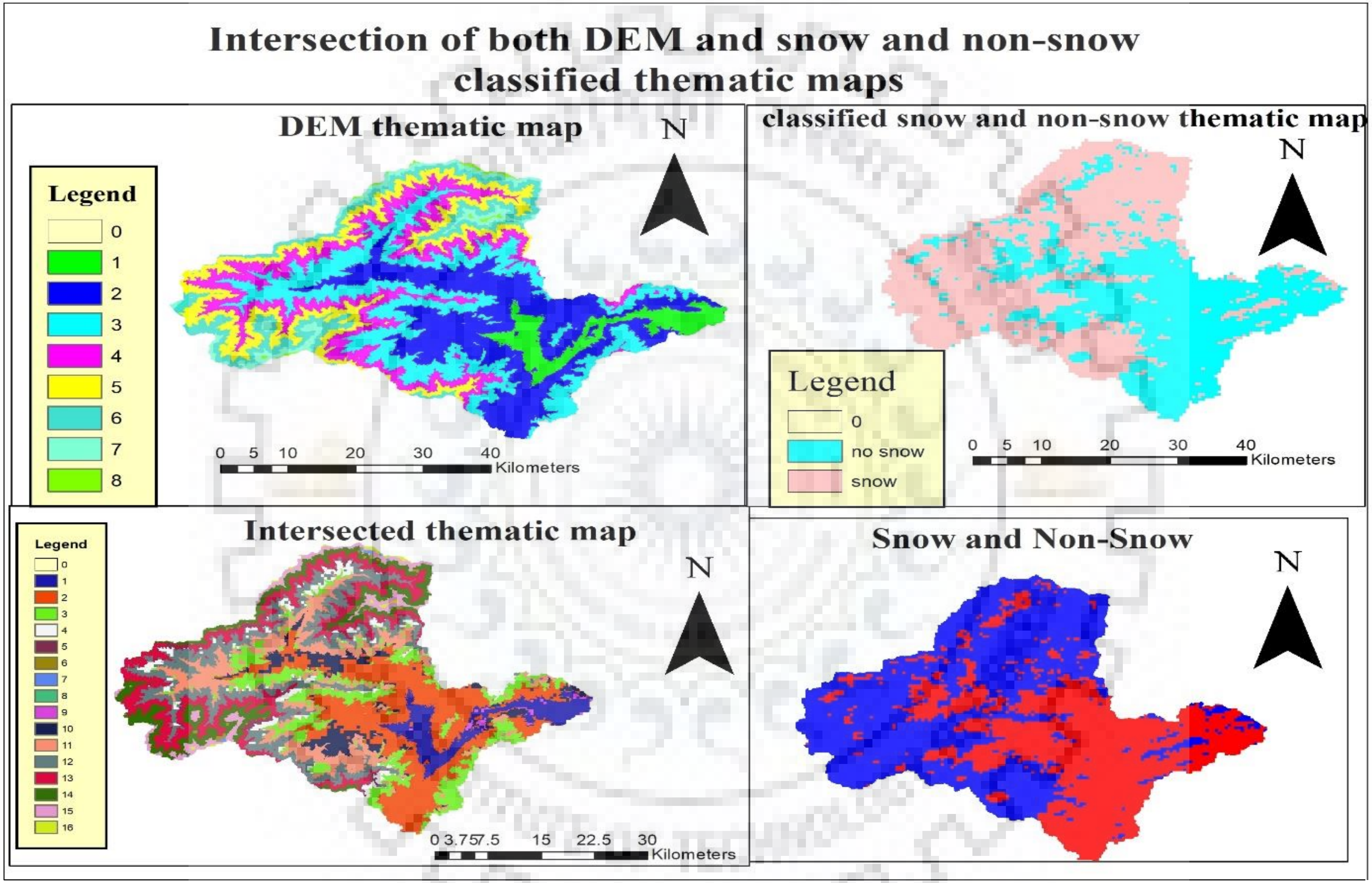

Figure 4.6: Intersected thematic maps of snow cover area.

Table 4.2: Snow in different elevation zones of the Upper Kabul River basin

No.	Elevation Zones of the DEM	Classified map (to non-snow=1) & (snow=2)	Number of cells covered by snow and Non- snow area	Color of the pixel which shown in the above Fig
1	1	1	434810	
2	2	1	242952	
3	3	1	76402	
4	4	1	23021	
5	5	1	6602	
6	6	1	2708	
7	7	1	179	
8	8	1	124670	
9	1	2	9275	
10	2	2	100144	
11	3	2	248814	
12	4	2	292356	
13	5	2	257012	
14	6	2	171702	
15	7	2	66249	
16	8	2	7355	

4.10. DEPLETION CURVES

In this study, SCA was computed for different elevation zones, for this purpose, DEM and SCA maps were processed for all the dates. The basin was divided into 8 elevation bands with an altitude difference of 322.2 m for convenience: these bands are >4518 m, 4518-4186, 4186-3855, 3855-3524, 3524-3192, 3192-2861, 2861-2529, 2529-2198m. these 8 eight elevation zones are also called band as band 1, band 2, band 3, band 8. The snow covers in each elevation zones were plotted against the time to construct the depletion curves for all the various elevation bands in the basin for all the years. The snow cover depletion curves vary significantly from years to years and hence separate curves have been drawn for each year under consideration. In order to simulate runoff on daily scale from the basin, daily SCA for each band was used as input to the model.

The analyses for each band were carried out. It was observed that the SCA in band 8, 7 remain 100% throughout the snow accumulates season. While bands 6,5,4 in winter season snow covered this area but in summer season these bands are snow free bands. Melting starts from the month of April and there is reduction in snow cover in the whole basin.

4.11. DETERMINATION OF MODEL PARAMETER

4.11.1. RUNOFF COEFFICIENTS

The runoff coefficients accounts for the losses that are the difference between the available water volume (snowmelt + rainfall) and the outflow from the basin or watershed. The average value of runoff coefficient for a basin is given by the ratio of annual runoff to annual precipitation. In fact, comparison of historical precipitation and runoff ratios provide a starting point for the runoff values. However, these ratios are not always easily obtained in view of the precipitation gauge catch deficit which particularly affects snowfall and of inadequate precipitation data from mountain region. At the start of the snowmelt season losses are usually very small because they are limited to evaporation from the snow surface, particularly at high elevations. Second, when some part of soil becomes exposed and vegetation grows, more losses must be expected due to evapotranspiration and interception. Until the end of snowmelt season. the SRM model uses different runoff coefficient for snow and rain (C_S and C_r), which can be varied throughout the season, the runoff coefficient also different from zone to zone in a basin.

The knowledge about basin and its hydrologic behavior under different hydro-meteorological condition it can ease the selection of C value for the Basin. The Upper Kabul River basin receive snow in winter a rain in summer season. But the contribution of the snowmelt is more than rainfall.

4.11.2. DEGREE- DAY FACTOR, (A)

The degree- day factor a [cm $^{0}C^{-1}$ d^{-1}] is used to covert the number- days T [^{o}C d] into the daily snowmelt expressed in depth of water M [cm]:

$$\mathbf{M = a.T} \qquad \textbf{(4.4)}$$

The degree-day factor is variable throughout the melt period because of the changing properties of the snow. the degree-day ratios can be evaluated by comparing degree-day values with the daily decrease of the snow water equivalent which is measured by radioactive snow gauge, snow pillow, or a snow lysimetric. These measurement has been done by Martinec, 1960. Show a considerable variability of degree day ratios from day to day.

The degree-day factor can be obtained from an empirical relation (Martinec, 1960):

$$\mathbf{a = 1.1 * \frac{\rho_s}{\rho_w}} \qquad \textbf{(4.5)}$$

Where

a = the degree-day factor [cm $^{0}C^{-1}$ d^{-1}]

$\boldsymbol{\rho_s}$ = density of snow

$\boldsymbol{\rho_w}$ = density of water

Generally, when the snow density increases (ρ_s), the albedo decreases, and the liquid water content in snow increases. Thus the snow density is an index of the changing properties which favor the snowmelt. Large variations can be expected if the melt season is long and there is large difference of elevation in the basin.

4.11.3. TEMPERATURE LAPSE RATE, Γ

If several temperature station at different altitudes are available, the lapse rate can be predetermined from historical data. otherwise it can be evaluated by analogy form other basins with regard to climatic condition.

The computer program accepts either a single or a basin wide lapse rate (option 0) or different rates for each zone (option 1). If the temperature station is located near the mean elevation of the basin, possible errors in the lapse rate are to some extent canceled out because the extrapolation of temperature takes place upwards as well as downwards. If the temperature station is at low altitude, SRM becomes sensitive to the laps rate.

4.11.4. CRITICAL TEMPERATURE, TCRIT

The purpose of critical temperature is to determine the measured or forecasted precipitation whether is rain or snow. model which simulate the runoff heavily depend on this parameter not only in the ablation period, but particularly, in the accumulation period. Snowmelt Runoff required critical temperature in order to decide whether precipitation immediately contributes to runoff as snowfall or rainfall. If the $T < T_{CRIT}$ snowfall took place. In this case, SRM automatically keeps the newly fallen snow in storage until it is melted on subsequent warm days.

4.11.5. RAINFALL CONTRIBUTING AREA, RCA

If rainfall happens in the same basins, it can be computed in two ways. In the initial stage, (option 0), it is assuming that rain falling on the snowpack early in the snowmelt season retained by the snow which is usually dry and deep, rainfall is added to snowmelt runoff only from the snow-free area, and the rainfall depth is decreased by the ratio snow-free area / zone area. At later stage. The snow cover becomes full of snow and the computer program should be switched to option 1. If rain falls on the snow cover, it is assumed that the same amount of water is released from the snowpack so that rain from the entire zone area is added to snowmelt. The melting effect of rain is neglected because the additional heat supplied by the liquid precipitation is considered to be small.

4.11.6. RECESSION COEFFICIENT, K

The recession coefficient (K) is a significant feature of SRM model since (1-k) is the proportion of the daily meltwater production which instantly appeared in the runoff stream. Analysis of historical discharge data is usually a good way to determine K value.

4.11.7. TIME LAG, L

The time lag between the temperature and the discharge can be determined from the observed data of discharge and temperature. The daily fluctuations in the snowmelt runoff enable the determination of the time lag directly from the hydrographs of the past years. if, for example, the discharge starts rising each day around noon, it lags behind the rise of temperature by about 6 hours. Consequently, temperature measure on the nth day correspond to discharge between 1200 hrs. on the nth day and 1200 hrs. on the n+1 day. Discharge data, however, are normally published for midnight-to- midnight intervals and need adjustments in order to be compared with the simulate values. Conversely, the simulated values can be adjusted to refer to the midnight-to-midnight periods. In larger basins with multiple elevation zones, the time lag changes during snowmelt season as a result of the changing spatial distribution of the snow cover with respect to the basin outlet.

5.1. DIVISION OF CATCHMENT INTO ELEVATION BANDS

Digital elevation model of the Upper Kabul River basin area was generated by different process through ArcGIS. The sinks are filled in the Digital elevation model and the process of flow directions and flow accumulation points were carried out before the delineation of main rivers and their watersheds up to confluence point at Upper Kabul river basin. the basin area is divided to 8 elevation zones of 322,2 m each starting from 1822m and ending at 4500m. The elevation zone map of the Upper Kabul river basin is shown in Fig 5.1. Zone wise and percentage of basin area is given in table 5.1.

Table 5.1: Division of Upper Kabul river basin into different elevation zones

Elevation Bands or zones	Zones	Zone area (Sq.km)	Percentage zone area to total catchment area
1867-2198	Zone-1	107.51	6.5
2198-2529	Zone-2	426.62	25.78
2529-2861	Zone-3	393.92	23.75
2861-3192	Zone-4	294.62	17.80
3192-3524	Zone-5	224.66	13.57
3524-3855	Zone-6	143.72	8.7
3855-4186	Zone-7	57.12	3.45
4186-4518	Zone-8	6.56	0.45

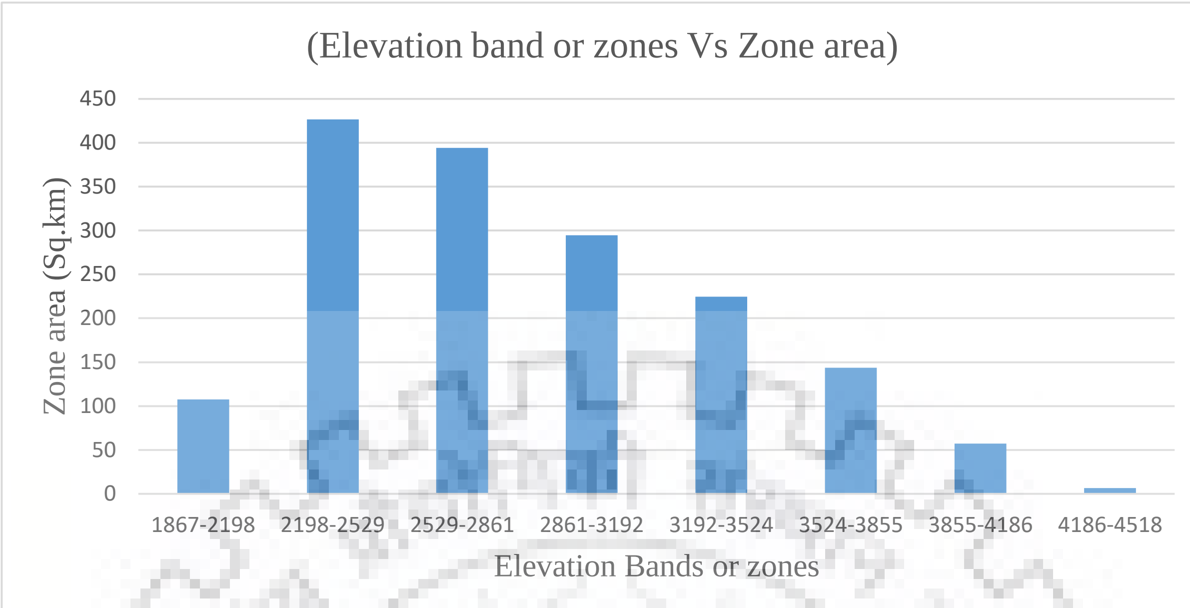

Figure 5.1: Area covered in each elevation zone of the upper kabul river basin

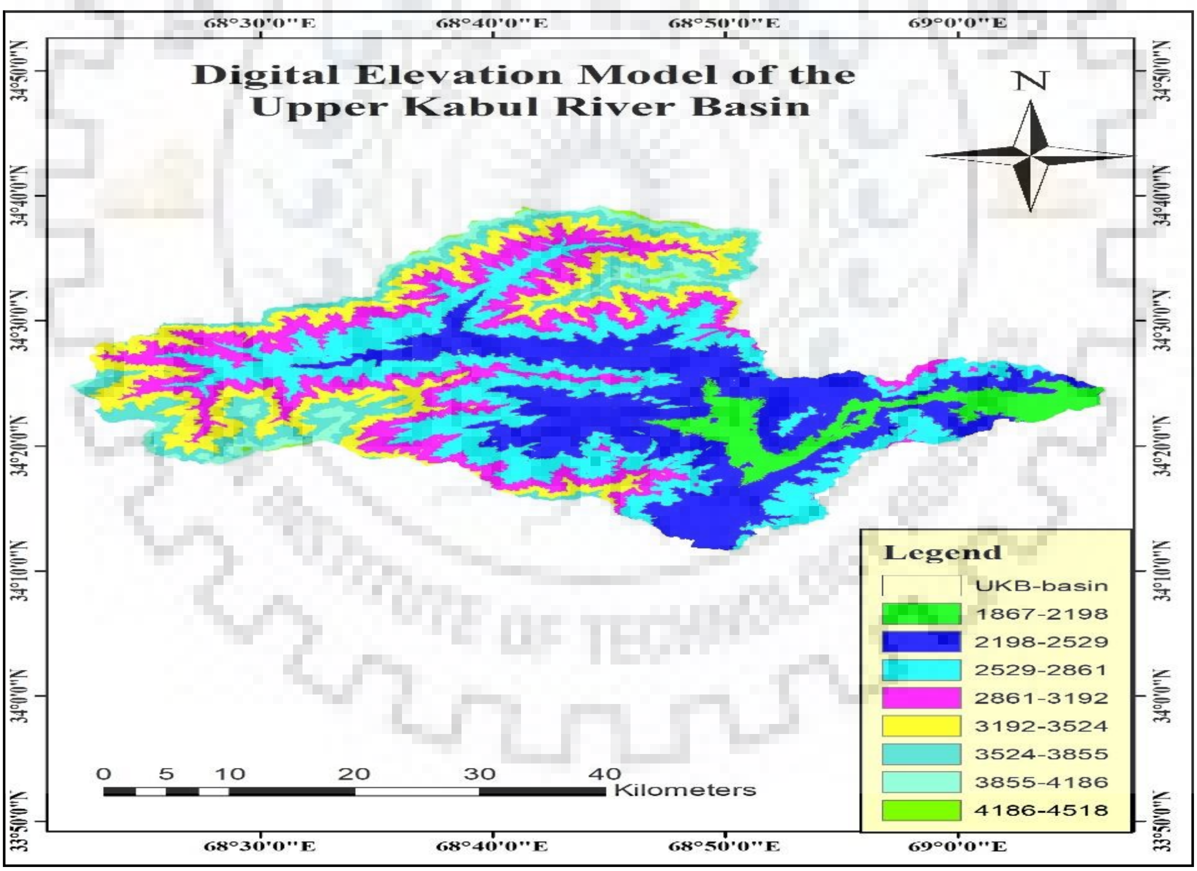

Figure 5.2: Classification of Upper Kabul River Basin into 8 elevation zones

5.2. SNOW-COVERED AREA

Snow cover area was estimated for the years2015 to 2017 from MODIS data. the snow covered area maps in different years are shown in figure 5.3 to 5.4. MODIS image classification was used for calculating the snow covered in the Upper Kabul river basin, the total percentage of snow cover in the study area was estimated for different dates and presented in table 5.2. 1st of December up to end of March, approximately when the accumulation of snow is started, the snow covered area was (49%, 61,25%, 70,625%, 76,18%, 80%) of the basin and the average snow covered area is in the year 2015. And during the accumulation the snow covered areas in the year 2016 was (57.01%, 71.75%, 79.09%,81.14%, 84,19%). And for the year 2017 the snow covered was (29%, 61%, 63%, 81.41%, 96,00%). but depletion of the snow covered area starts from April on ward till August.

In the present study, the snow curve has been used for different elevation zones for further application in the model. For this purpose, DEM and SCA maps have processed for all the dates.

Table 5.2: SCA in Different elevation zones in upper Kabul-river basin

Dates	Snow covered area in Different Elevation Bands (km^2)							
	1867-2198	2198-2529	2529-2860	2860-3191	3191-3522	3522-3853	3853-4184	4184-4515
Nov-2015	58.48	294.16	737.34	911.66	780.72	518.05	201.46	22.79
Dec-2015	90.87	435.78	579.95	816.06	617.69	438.89	151.96	20.83
Jan-2015	88.10	662.79	1032.41	1002.87	778.01	494.07	192.54	22.14
Feb-2015	86.41	615.60	940.14	841.38	621.86	387.26	150.13	17.30
Mar-2015	101.37	563.04	938.83	1010.03	804.40	516.00	191.44	19.08
Nov-2016	19.88	115.72	223.95	385.49	369.57	276.11	143.84	20.96
Dec-2016	28.39	177.7213	423.43	760.04	731.45	512.51	208.26	23.66
Jan-2016	132.38	926.14	1155.31	1039.30	815.60	527.67	208.54	23.21
Feb-2016	34.82	456.83	945.91	982.41	827.39	543.86	213.11	23.29
Mar-2016	67.91	336.93	751.22	917.04	721.02	416.52	139.32	13.38
Nov-2017	33.34	237.81	474.35	487.24	373.01	238.30	115.39	18.76
Dec-2017	76.53	530.41	942.41	979.07	810.09	516.91	201.55	23.03
Jan-2017	103.86	733.95	968.03	872.58	750.55	516.05	205.54	22.88
Feb-2017	227.63	1180.01	1256.96	1086.01	857.52	547.99	209.72	22.39
Mar-2017	91.04	878.72	1267.45	1090.91	868.49	560.75	216.32	22.94

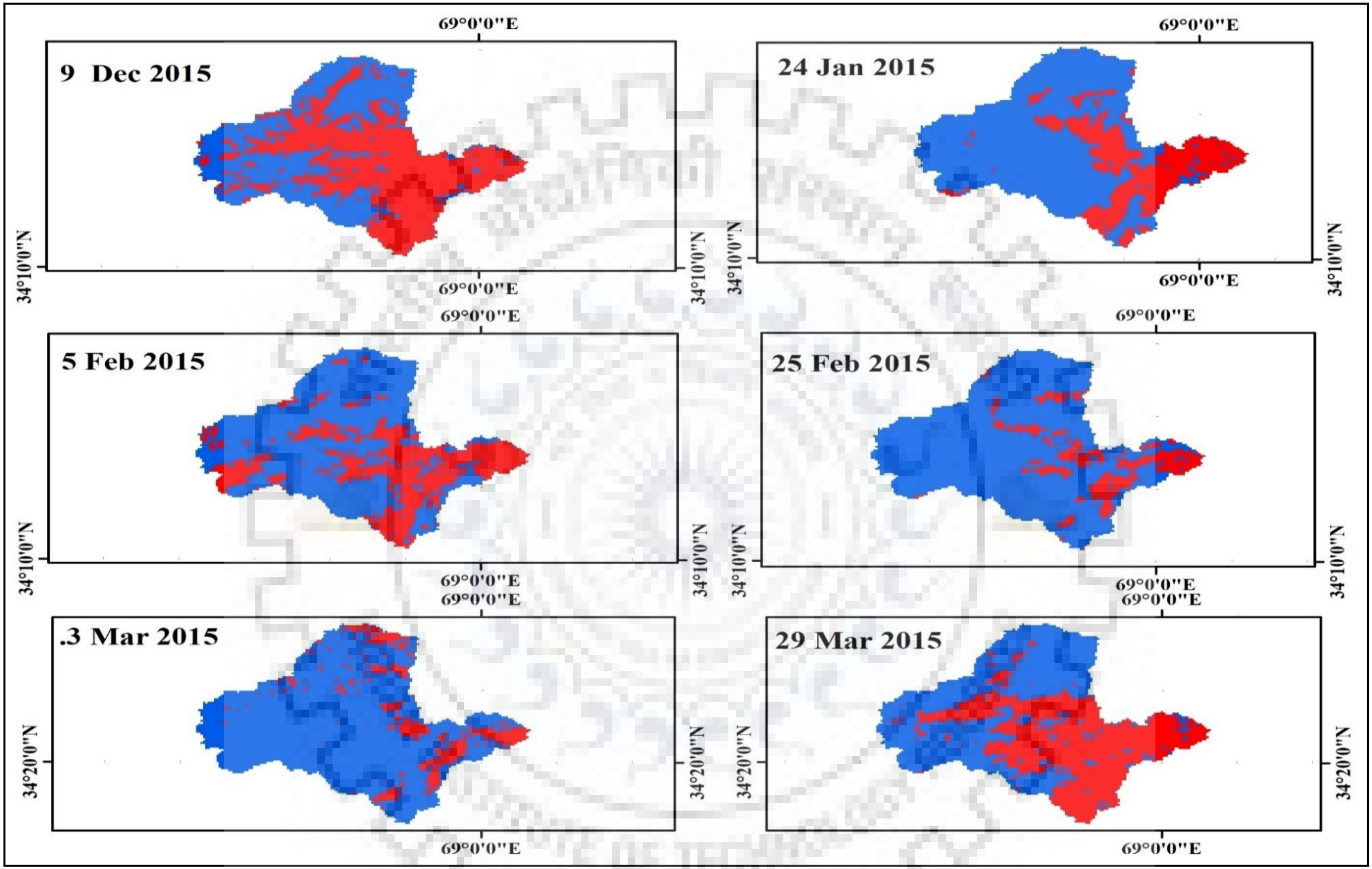

Figure 5.3: Snow cover area for the year 2015

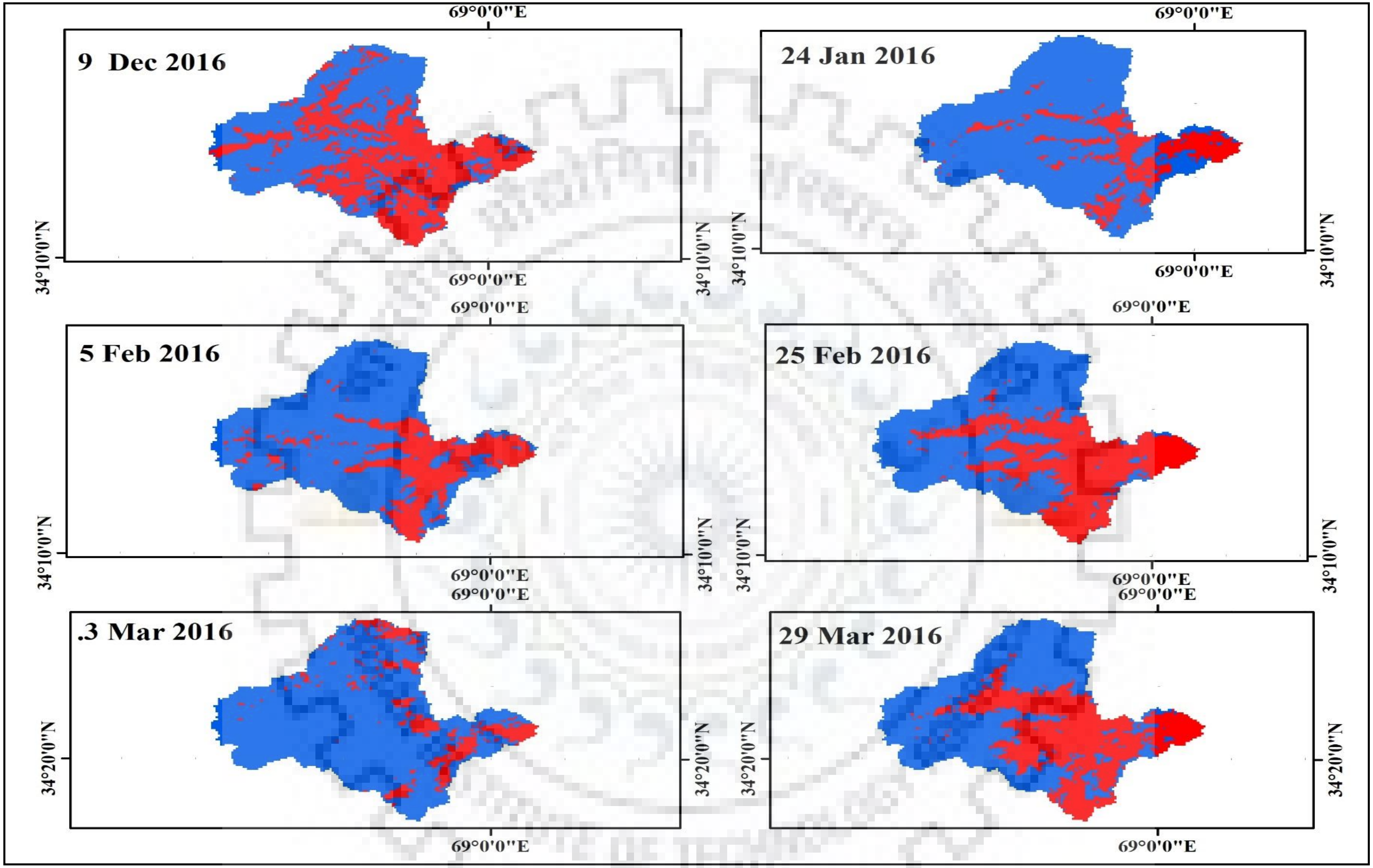

Figure 5.4: Snow cover area for the year 2016

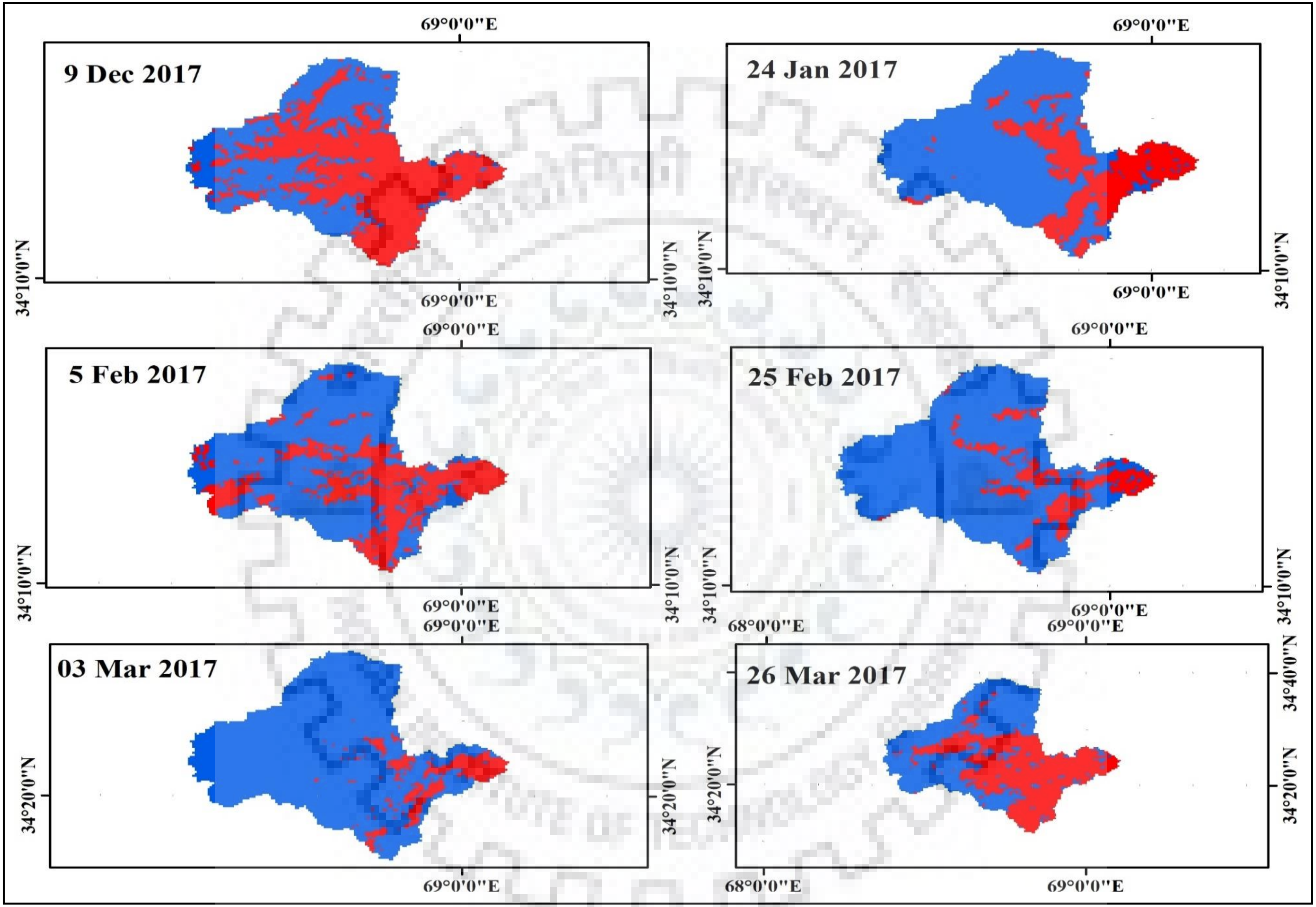

Figure 5.5: Snow cover area for the year 2017

5.3. SNOW DEPLETION CURVE

Quantification of snow-covered area and its decline throughout the snowmelt season is an important input for snowmelt runoff models in the prediction of runoff and simulation of streamflow. Several remote sensing methods of snow cover mapping exist today that can be used in determining the progressive reduction of snow cover during snowmelt; however, some of these methods of snow mapping, such as the Moderate-Resolution Imaging Spectroradiometer (MODIS) satellite sensor, are relatively new. The relative regency of MODIS which launched in 1999 and other new remote sensing methods, limit their direct use in developing historical snow depletion curves. Nevertheless, historical depletion curves are important for general model validation purposes, and also for studying impacts of varying or changing climate on snowmelt and surface runoff. For the latter purpose, historical depletion curves are useful both for establishing a baseline, and for testing impacts of perturbations to climate such as associated with climate change scenarios.

These historical depletion curves, among many other uses, provide snow cover information for snowmelt runoff modeling in hydrologic models such as snowmelt runoff model (SRM). Based on numerous observations in the literature that snowmelt occurs in a repeating patterns, albeit shifted and/or accelerated or decelerated in time, from one year to the next, a method is presented in this study that makes use of the available remotely sensed MODIS data and concurrent ground based SNOTEL data to construct a single dimensionless snow depletion curve that is subsequently used with historical SNOTEL data to reconstruct snow depletion curves for base periods preceding the availability of current satellite remote sensing, and for future periods associated with altered climate scenarios; the method may also be used in any current snowmelt season to forecast the snowmelt and improve streamflow forecasts.

In the present research, the SCA was computed for different elevation zones. For this purpose, DEM and SCA maps were processed for the year 2017. The basin was divided into 8 elevation bands with and altitude difference 398 for convenience: these bands are 1800-2198 m, 2198-2529 m, 2529-2861 m, 2861-3192, 3192-3524, 3524-3855, 3855-4186, 4186-4518. These 8 elevation zones are defined as band 1, band 2, band3, band 4, band 5, band 6, band 7, band 8. The snow covered area in each bands were plotted against the elapsed time to derive the depletion curves for all the different elevation bands for the whole 2017 year. The snow cover depletion curves vary significantly from month to month and year to year as well and hence for each month of the year and hence separate curves have been drawn for each month of the year under consideration. In order to simulate runoff on daily scale from the basin, daily SCA for each band was used as input to the model. The analyses for each band and for every month of the year (2017) were carried out. It was observed that SCA in the higher elevation bands (5,6,7,8) during winter season (Dec, Jan, Feb, March), the depletion curve reached to 100%, but in lower elevation bands the snow depletion curves gradually decreased. As already illustrated, that the upper Kabul River basin received heavily snow fall in the winter season from Dec to March, melting of snow starts from the month of Apr and the reduction of snow cover in Upper Kabul River basin starts from April onwards till August, therefore, mainly there is variation in the snow depletion curves from month to month.

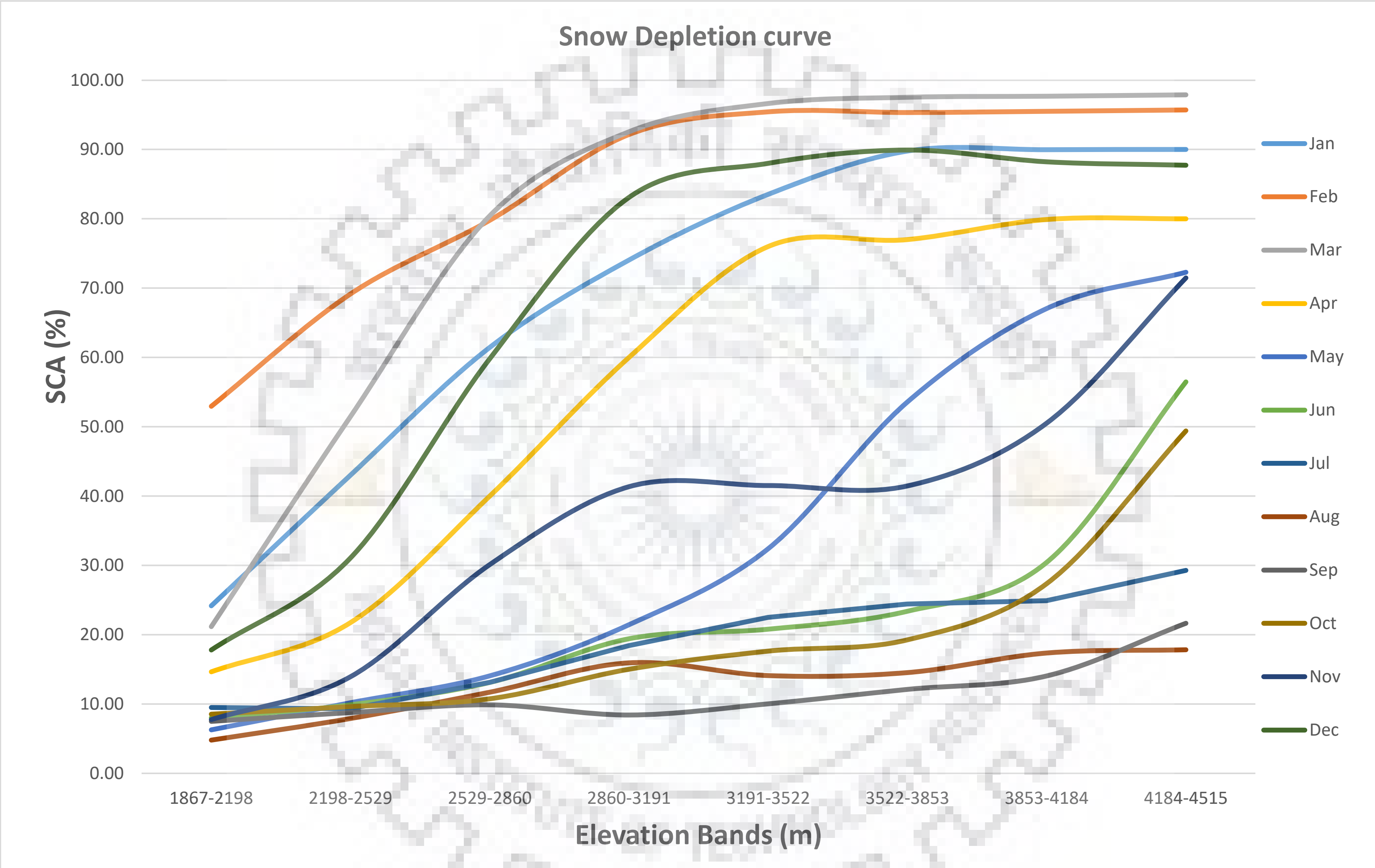

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Figure 5.6: Snow Depletion curve of the basin of the whole year.

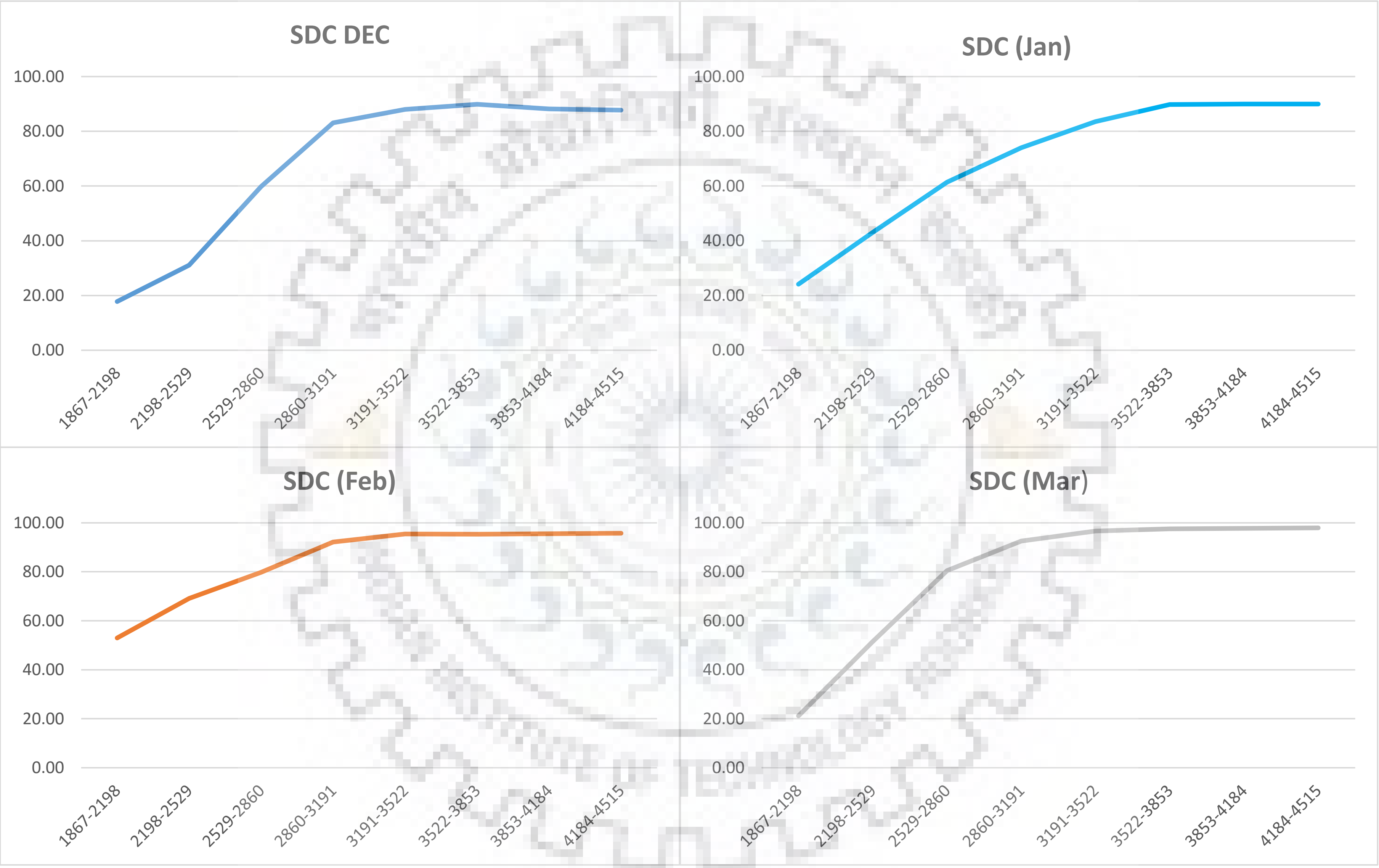


Figure 5.7: Snow Depletion curve for the winter season

5.4. APPLICATION OF SRM MODEL

5.4.1. SIMULATION OF STREAMFLOW

Daily runoff simulation for the snow melting season 2015, 2016, 2017 were carried out after calibrating the SRM model for the year 2015. The model requires seven parameters, three input variables and catchment or basin characteristics for its snowmelt runoff simulation. The basin receives snowfall in winter season during December to March, which melts in the summer period.

5.4.2. CALIBRATION OF MODEL

Generally, hydrological models are calibrated using observed and simulated data. The existing data set is divided into two parts, one is used for calibration purpose and the next validation to check the model performance in simulation mode. The values of the calibrated parameter were calculated according to the overall performance of the model and the reproduction of the flow hydrograph, in this study for the calibration the daily stream flow of one year (2015) have been considered. The model calibration values are given in table 5.3.

Table 5.3: Calibrated model Parameters for 2015

Date	Monthly lapse rate (^{o}C/100)	T_{CRIT} (^{0}C)	A_n (cm/^{0}C/day)	Lag time (Hour)	C_a	C_R	RCA	X coeffi.	Y coeffi
Jan	0.6	2	0.12	3	0.15	0.10.	0	0.9	0.001
Feb	0.7	2	0.12	3	0.15	0.10.	0	0.9	0.001
Mar	0.88	2	0.12	3	0.15	0.10.	0	0.9	0.001
Apr	0.65	2	0.12	3	0.15	0.10.	0	0.9	0.001
May	2.5	2	0.12	3	0.15	0.10.	0	0.9	0.001
Jun	2.9	2	0.12	3	0.15	0.10.	0	0.9	0.001
July	4	2	0.12	3	0.15	0.10.	0	0.9	0.001
Aug	5	2	0.12	3	0.15	0.10.	0	0.9	0.001
Sep	3	2	0.12	3	0.15	0.10.	0	0.9	0.001
Oct	3	2	0.12	3	0.15	0.10.	0	0.9	0.001
Nov	2.9	2	0.12	3	0.15	0.10.	0	0.9	0.001
Dec	0.8	2	0.12	3	0.15	0.10.	0	0.9	0.001

The efficiency of the model for the calibration year is giver in table 5.4.

Table 5.4: Efficiency of the model for the calibration period

Calibration year	R^2	Volume Difference (%)
2015	0.89	9.34

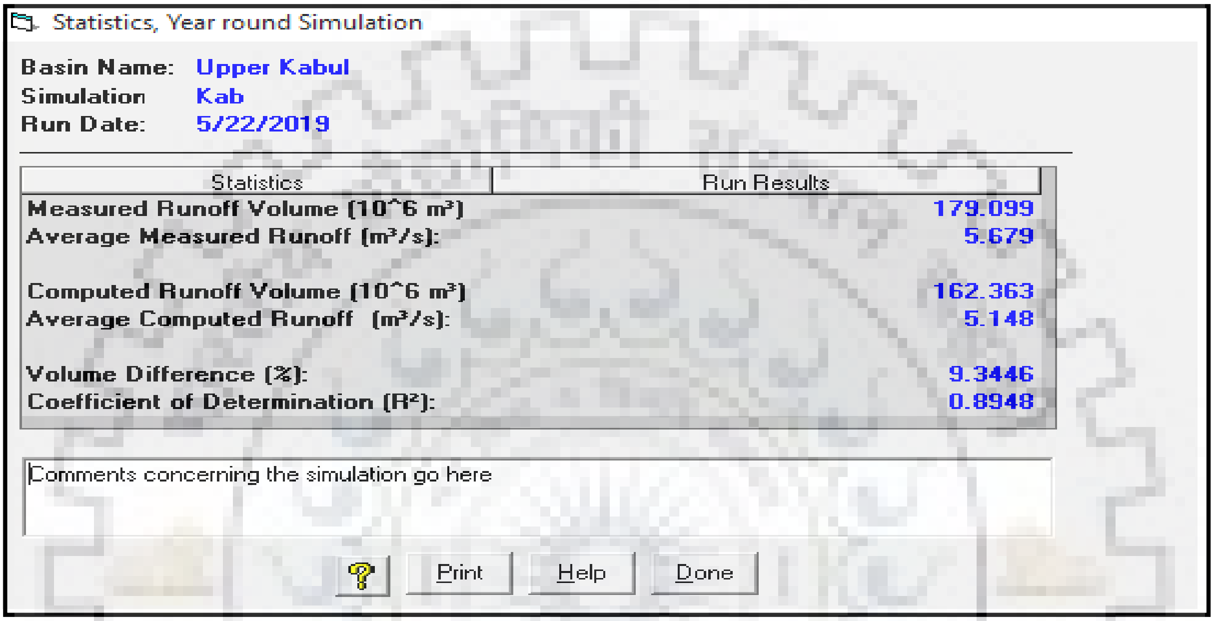


Figure 5.8: Statistics of the daily simulation for the year 2015

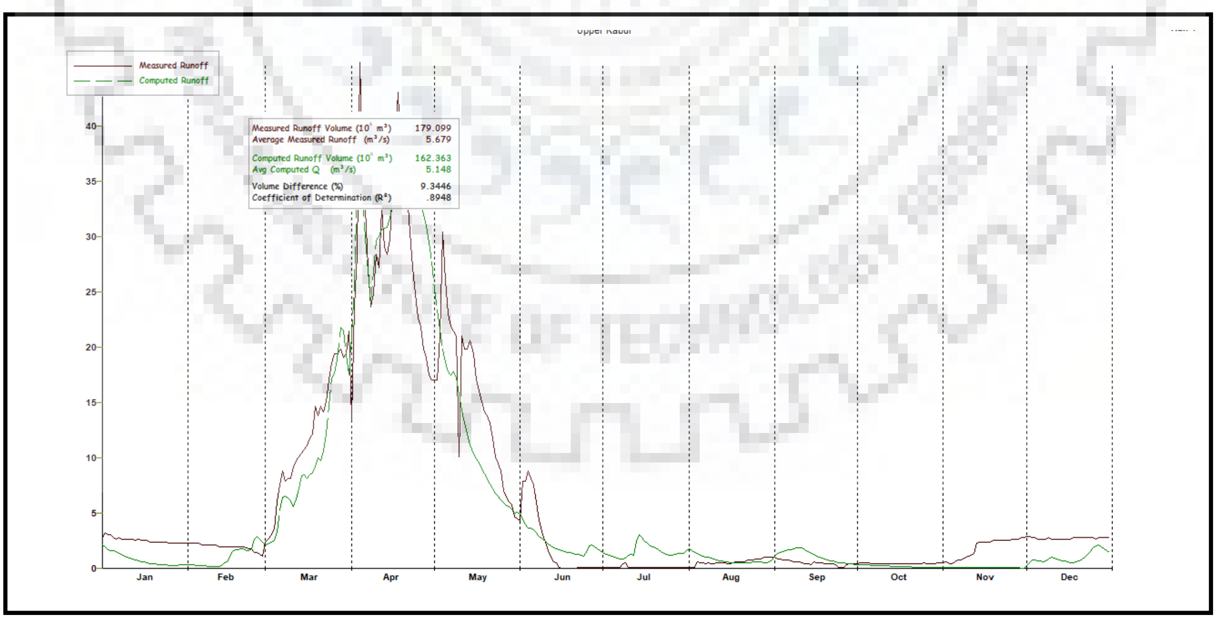
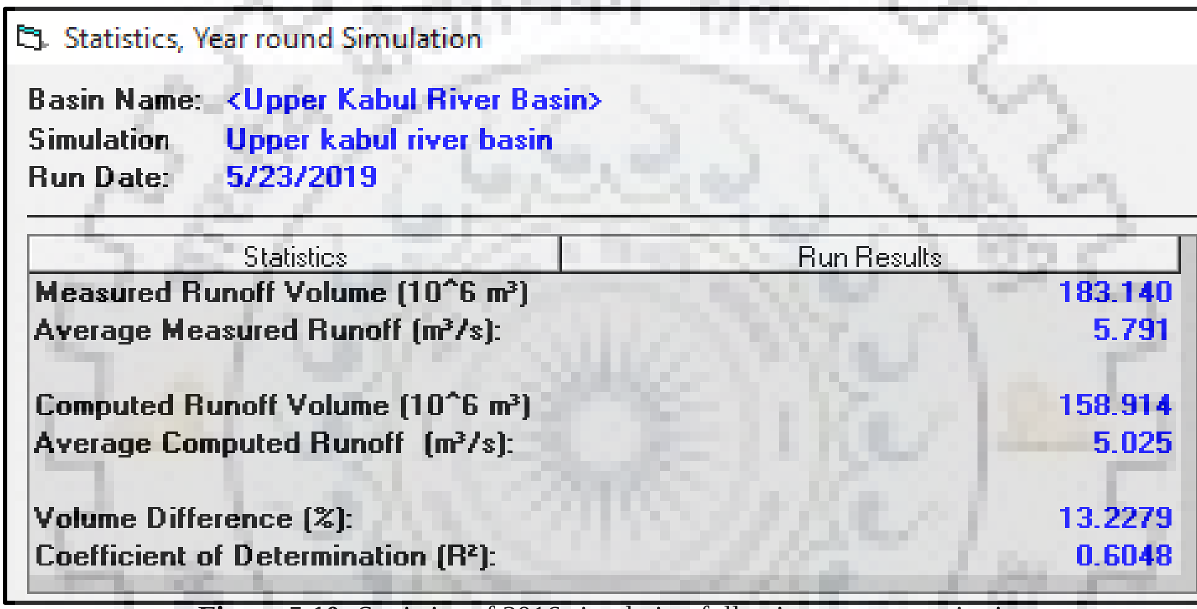


Figure 5.9: Runoff simulation for the melting season 2015 of Upper Kabul River basin. the red line is the measured discharge line at upper Kabul river basin whereas the green line is simulated runoff.

The statistics of this simulation show that the simulation is very well both for daily and for the whole melting season.

5.4.3. SRM VERIFICATION FOR 2016 AND 2017

Once the model is calibrated it should be applicable to another year only changing the precipitation (rain), snow cover, temperature values for the years without changing the parameters used in the calibrated year. This was done for the two years (2016 and 2017) with the same parameters used in 2015 simulation. For the validation period the result is given in Fig 5.6 and 5.7. For the year 2016 Volume Difference D_v = 11. 3776 whereas the coefficient of determination R^2 =0.60.


Statistics, Year round Simulation

Basin Name: <Upper Kabul River Basin>
Simulation Upper kabul river basin
Run Date: 5/23/2019

Statistics	Run Results
Measured Runoff Volume (10^6 m³)	183.140
Average Measured Runoff (m³/s):	5.791
Computed Runoff Volume (10^6 m³)	158.914
Average Computed Runoff (m³/s):	5.025
Volume Difference (%):	13.2279
Coefficient of Determination (R²):	0.6048

Figure 5.10: Statistics of 2016 simulation following accuracy criteria.

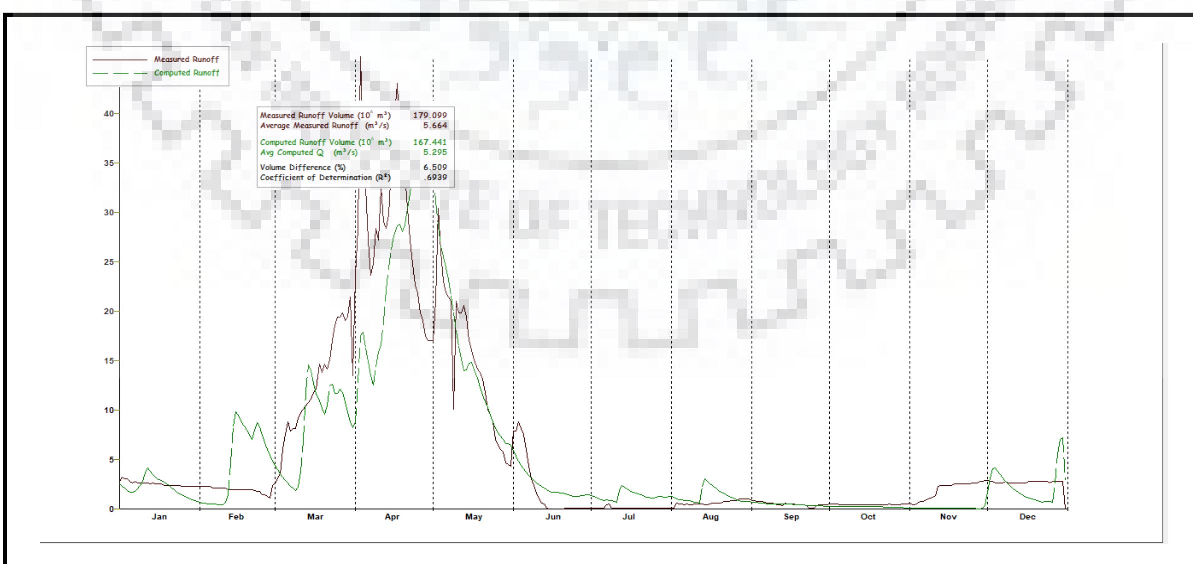


Figure 5.11: Runoff simulation for the year 2016 of upper Kabul river basin.

And the result for year 2017 for Q are given in Fig 5.9.. for 2017 Volume Difference Dv = 6.50 whereas the coefficient of determination R2 =0.69.

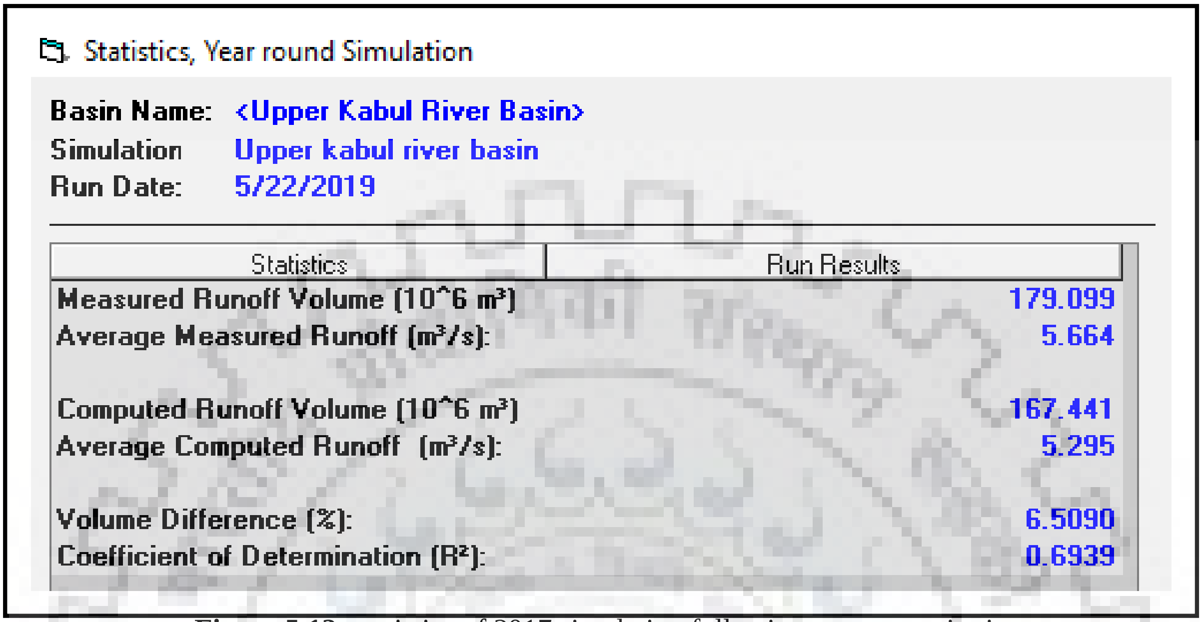


Figure 5.12: statistics of 2017 simulation following accuracy criteria.

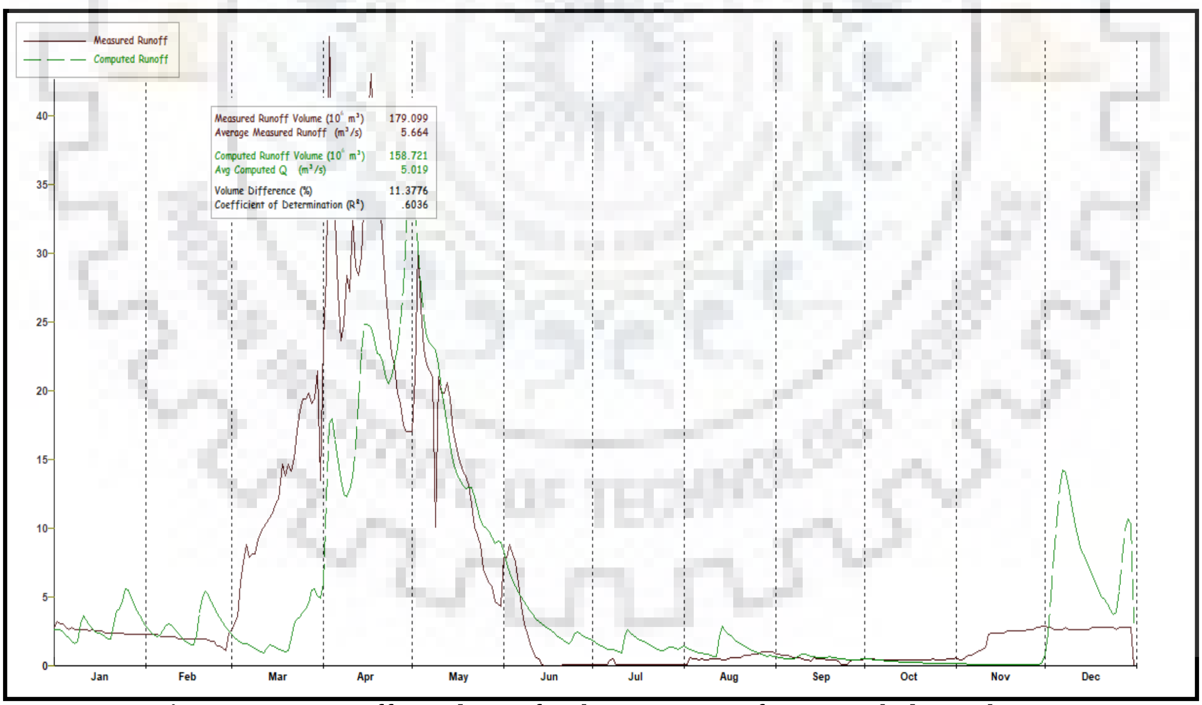


Figure 5.13: Runoff simulation for the year 2017 of upper Kabul river basin.

5.4.4. DISCUSSION ON MODEL SIMULATION

In SRM simulations of 2016 and 2017, the accuracy criteria show week good with $R^2 = 0.60$ and $R^2 = 0.69$. for getting the higher value of R^2. therefore, the runoff coefficient and the degree day factor have to be changed, keeping the lower and Upper limits of these parameters well within the extreme values used during the calibration year 2015, which means that runoff coefficients and degree day factor do not remain constant temporally i.e. from year to year. Heavy snowfall means more snow available for the melting season and for a much longer time. In fact, snow that remains for a longer time is denser and hence a higher degree day is needed. Similarly, in a particular year there are below average rains or above average high temperature the catchment will be dry and vegetation will be stressed.

Air temperature data recorded at Upper Kabul River station also confirms that temperature on the average from April to Nov were higher as compared to year 2015 (Fig 5.11) at the same time precipitation data revealed that there was higher variation during months in these years. and the other factor is also land use and land cover changes.

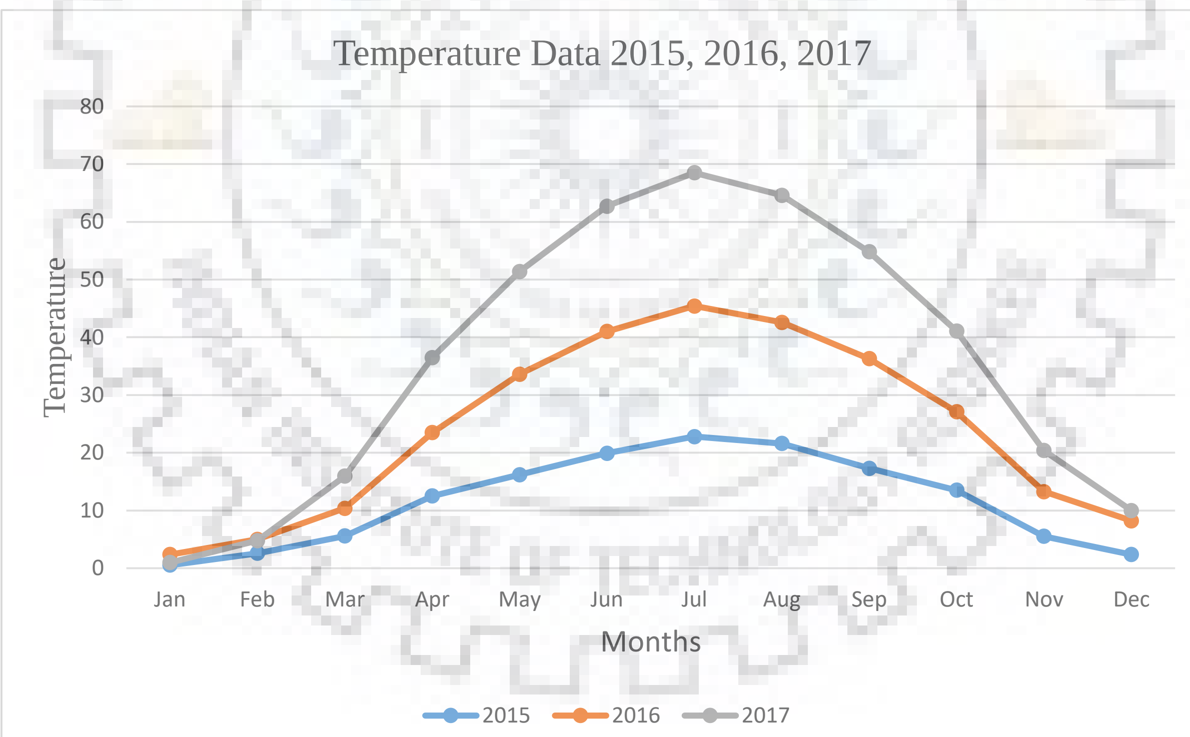


Figure 5.14: Variation of temperature data among the years.

As result, theses factor effect on the result of validation period. If we adjust these parameters, the simulation would give a better fitting hydrograph.

CHAPTER-6

Upper Kabul River basin is a snow fed river, starting with tributaries, Maidan river, Nerkh river, Durani river. Evaluation of snowmelt in Upper Kabul river basin is very important to understand the behavior of Kabul River. Although snow in the basin is an important water resource, it causes flooding at the lower part of Kabul river. Whenever snow starts melting abruptly flooding due to over flow in the melting season; it is not only the problem in Kabul province it's also effect the other provinces as well.

During winter season most of the precipitation is in the form of snow in the basin, followed by a little rainy season from April to August. The snow cover of the basin has been mapped using data from the MODIS sensor onboard the Terra and Aqua satellites. MODIS provide different band combination images free of cost

The present study has been carried out to estimate snowmelt runoff from upper Kabul basin up to Kabul. For simulation of snowmelt runoff from study catchment, SRM model has been used. The model requires snow cover extent, temperature, precipitation. A methodology to produce daily SCA directly from Terra and Aqua MODIS snow cover map products (MOD10A2) is presented. The basin has been divided into 8 elevation zones. A critical temperature was used for distinguishing the precipitation as rain or snow, and snow cover has been computed for these elevation zones using MODIS data. on the basis of SCA, snow cover depletion curves have been prepared. The geographical information system (GIS) software was used for the preparation of the digital elevation map, and for computation of the aerial extent of snow cover in different elevation bands ERDAS IMAGINE 2015 software was used. The daily snow cover data was extracted from the depletion curves which has been prepared form the snow cover information obtained from the satellite data.

It was observed that the snowmelt starts in the month of March while after December snow cover accumulation start. The upper part of the basin remains snow covered for a long time and the lower part remains snow covered for a short time.

Snowmelt runoff model (SRM) developed by Martinec 1975 has been used for the runoff which produced from snow or both snow and rain. The model requires snow extent, precipitation and temperature data as input variables, and some catchment parameters for calibrating the model.

After getting the input variables for the running of SRM model, several simulations in the snowmelt season have been done for different years i.e. 2015, 2016 and 2017. Graphical display and statistics can tell about the accuracy of the model's simulation. The model seems to be satisfactory the values of R^2 and D_v ae within the acceptable range. If the correct inputs are available for the models, the model can give good results in simulation.

Following conclusions are drawn from the present study:

- the characteristics of snow cover in the basin shows that the accumulation of snow at higher altitude starts form the December and the snowlines come down to lower elevation up to 2000 and sometimes the whole basin covered by snow in winter season.
- By the start of April, the snowmelt begins and the snowline gradually decreases day by day.
- Till the start of the melting season, more than 90% of the basin area is covered with snow and then reduces (approx. 20%) at the end of the melt season.
- It can be concluded form above facts that a large part of the study basin contributes actively to stream flow in the form of snowmelt runoff.
- The calibration of the model for computation of stream flow has indicated that the low flows and almost all the peaks in the stream flow are usually well reproduced.
- For calibration of the SRM model high value of R^2 = 0.89 and volume difference of 9.34% between observed and simulated runoff indicates good model fit.
- From the study it appears that snow runoff contribution in Kabul river in the beginning of summer season is more but laterally it gradually decreases.

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