ROLE OF REMOTE SENSING IN SURFACE WATER ASSESSMENT INCLUDING HYDROLOGICAL MODELLING

Introduction

Water is the most spread substance in the natural environment and it is the most abundant substance on earth, the principal constituent of all livings, and a major force constantly shaping the surface of the earth. Water exists in three states: liquid, solid, and invisible vapour. It forms oceans, seas, lakes, rivers, and ground waters in the top layers of Earth's crust and soil cover. In a solid state, it exists as ice and snow cover in polar and alpine regions. Some amount of water is contained in the air as water vapour, water droplets, and ice crystals as well as in the biosphere. Throughout history water has been considered as a natural resources critical to human survival. Increasing population and higher levels of human activities, including effluent disposals into surface and ground water resources, have made suitable management of water resources a complex task throughout the world. To reliably assess water storage on the Earth is a complicated problem because water is very dynamic. It is in a permanent motion converting from liquid to solid or gaseous phase, or vice versa.

Surface Water Resources of the world

The Earth's hydrosphere (free water being in state in liquid, solid or gaseous state in the atmosphere) contains, an amount of about 1386 million km^3 of water. However, 97.5percent of this amount is saline water and only 2.5percent fresh water. The greater portion of the fresh water (68.7%) is in the shape of ice and permanent snow cover in the Antarctic, the Arctic, and mountainous regions. Next 29.9percent are fresh ground waters. Only 0.26percent of the total amount of fresh waters on the Earth is concentrated in lakes, reservoirs, and river systems.

Quantitative indices of different components of the global hydrological cycle are given here. Every year the water turnover on Earth involves 577,000 km^3 of water. It is the water that evaporates from the oceanic surface (502,800 km^3) and from land (74,200 km^3). The same water amount falls as atmospheric precipitation (on the ocean 458,000 km^3 and on land 119,000 km^3). The difference between precipitation and evaporation from land surface (119,000 - 74,200 = 44,800 km^3/year) represents the total runoff of Earth's rivers (42,600 km^3/year), and a direct groundwater runoff to the ocean (2200 km^3/year). Though the fresh water on the earth surface is only 0.26percent, it is the principal source of life necessities and man's economic activities.

Remote Sensing technology plays an important role, in understanding these fresh waters, as it can help in quantification of water stored in snow and ice crests, in most inaccessible areas and also surface water in liquid form can be carried out with remote sensing technology. Next few paragraphs will give incite into these aspects.

Spectral Characteristics of Snow/Ice and Surface Water

Remote sensing offers a valuable tool for obtaining snow data for predicting snowmelt runoff. Conventionally obtaining these data is extremely labour intensive, expensive and potentially dangerous. Quantification of the surface water assessment using remote sensing data, it is necessary to understand the interaction of light with ice and surface water on the earth surface.

However due to ever increasing technological developments, it is necessary to understand the interaction of the light with snow/ice and surface water.

Snow and Ice

All regions of the electromagnetic spectrum can provide useful information about the snow pack and its condition. Albedo of the snow surface is the property most easily measured by remote sensing. Typically, new snow will have an albedo of 90percent or more, where as older snow that has been weathered and has accumulated dust and litter cab have an albedo as low as 40percent (Foster et.al., 1987). The reflectivity depends upon snow properties such as grain size and shape, water content, surface roughness, depth and presence of impurities. The reflectivity of new snow decreases as it ages in both visible and infrared region of spectrum, however the decrease is more pronounced in the infrared as shown in Figure 1.

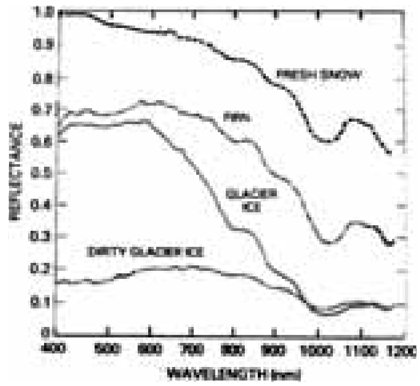

Figure1:Spectral reflectance curves of snow and ice (Hall and Martinec, 1985).

This increased sensitivity in the infrared region is caused by the increasing grain size of the snow which results from melting and refreezing (Engman and Gurney, 1991). For the most, decreasing reflectivity in the visible region can be attributed to the contaminants such as dust, pollen and aerosols. Properties of Snow in different bands of electromagnetic spectrum are given in table 1.

Table 1: Properties of Snow in different bands of Electro Magnetic spectrum

Property	Sensor Band: Visible/ Near Infrared	Thermal Infrared	Microwave
Snow-Covered Area	Yes	Yes	Yes
Depth	If Very Shallow	Weak	Moderate
Snow Water Equivalent	If Very Shallow	Weak	Strong
Stratigraphy	No	Weak	Strong
Albedo	Strong	No	No
Liquid Water Content	Weak	Weak	Strong
Temperature	No	Strong	Weak
Snow Soil Boundary	No	No	Weak (high frequency to strong low frequency)
All Weather capability	No	No	Yes
Current Best resolution from Space platform	Meters	Tens of meters	Passive: 30(high frequency) to 150 km (low frequency); Active 10 of meters

Surface Water

The majority of radiant flux incident upon water is either absorbed or transmitted and little is reflected. In visible wavelength of EMR, little light is absorbed, a small amount (less than 5 percent) is reflected and rest is transmitted. Water absorbs NIR and MIR strongly (Figure 2) leaving little radiation to be either reflected or transmitted. This results in sharp contrast between any water body and surrounding land surface.

Longer wavelength visible and near infrared radiation is absorbed more by water than shorter visible wavelengths. Thus water typically looks blue or blue-green due to stronger reflectance at these shorter wavelengths, and darker if viewed at red or near infrared wavelengths. If there is suspended sediment present in the upper layers of the water body, then this will allow better reflectivity and a brighter appearance of the water. The apparent colour of the water will show a slight shift to longer wavelengths. Suspended sediment can be easily confused with shallow (but clear) water, since these two phenomena appear very similar. Chlorophyll in algae absorbs more of the blue wavelengths and reflects the green, making the water appear more green in colour when algae is present. The topography of the water surface (rough, smooth, floating materials, etc.) can also lead to complications for water-related interpretation due to potential problems of specular reflection and other influences on colour and brightness.

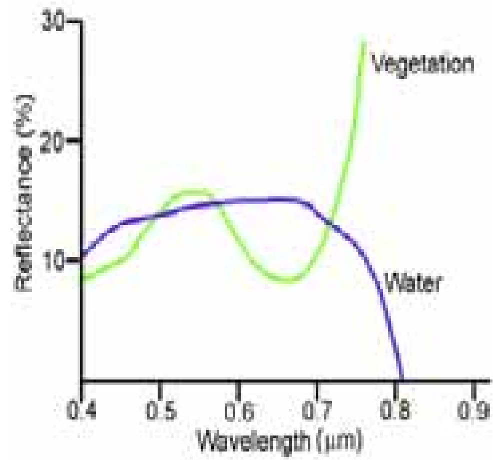

Figure 2: Spectral Reflectance curve of water and vegetation.

Mapping of Snow/Ice and Surface Water using remote sensing data

Mapping of snow and Ice and surface water using remote sensing data is essential to estimate the amount of water available in solid and liquid forms.

Satellite Data (Optical)

Snow/Ice. Snow cover can be detected and monitored with a variety of remote sensing devices in cost effective method. The greatest application has been found in the visible and near infrared region of electro magnetic spectrum. The red band (0.6-0.7 µm) was used extensively for snow cover mapping because of strong contrast with snow free areas. In the visible (VIS) and near infra red (NIR) region the senor signals gets saturated because of very high reflectance. A spectral channel (1.55 – 1.75 µm) can be used to assist in mapping snow cover when clouds partially cover a drainage basin. In this situation, a method has been developed to estimate the snow over in the cloud-obscure parts of the basin (Lichtenegger et.al, 1981; Baumgartner et. al.; Ehrler et. al., 1997). The method uses digital topographic data and assumes that pixels with equal elevation, aspect and slope have the same relative snow coverage over the entire basin.

Another easy method to derive snow cover as using the two band data of SWIR and VIS bands it is possible to derive an index, Normalised Difference Snow Index (NDSI), which is an analogue to normalised difference vegetation index (NDVI),

$$\text{NDSI} = \frac{(\text{Reflectance in Green} - \text{Reflectance in SWIR})}{(\text{Reflectance in Green} - \text{Reflectance in SWIR})} \qquad (1)$$

Using Landsat TM, Band 2 (0.52–0.60 µm) and Band 5 (1.55–1.75 µm) and using TERRA MODIS, Band 4 (0.545–0.565 µm) and Band 6 (1.628–1.652 µm) are used for deriving NDSI. Using MODIS data, SNOWMOD algorithm (Hall, 1995) is developed to derive global snow cover products. A pixel in a non-densely forested region will be mapped as snow if the NDSI is > 0.4 and reflectance in MODIS band 2 (0.841–0.876 µm) is >11%.

However, if the MODIS, band 4 (0.545 – 0.565 µm) reflectance is < 10%, then the pixel will not be mapped as snow even if the other criteria are met. The NDVI and NDSI are used together to improve snow mapping in dense forests.

If the NDVI =~0.1, the pixel may be mapped as snow even if the NDSI is < 0.4 (Klein et. al., 1998).

Using MODIS infrared bands 31 (10.78–11.28 µm) and 32 (11.77–12.27 µm), a split-window technique (Key et.al., 1997) is used to estimate ground temperature. If the temperature of a pixel is >277 ^{o}K, then the pixel will not be mapped as snow.

Surface Water. In liquid form**,** the surface water is stored in lakes, reservoirs and rivers. Satellite sensors does not measure hydrological data directly, the hydrology is obtained only after interpretation of the measured electro magnetic radiation. Locating and delineating surface waters are most easily using remote sensing data in the near-infrared and visible wavelength. Since water absorbs most energy in the near and middle-infrared wavelengths (>0.8 µm) there is energy available for reflection in these wavelengths. Vegetation and soil, on the other hand, have lower reflectance in the visible bands (0.4 – 0.8 µm) and high reflectance in the near and middle infrared wave lengths. Thus, on gray scale infrared images or multi-spectral scanner images in the reflective infrared portion of the spectrum, water bodies appear dark and stand out in stark contrast to surrounding vegetative and soil features (Swain and Davis, 1978). Regression equations are developed between surface water body and available pixel reflectance. Such regressions are not applicable beyond the time and space constraints of the original data. Various bands which are useful for surface water body detection is in the range of 0.79 – 0.89 µm. This band is available in NOAA AVHRR (0.72 – 1.1 µm), Band 4 of Landsat TM (0.76 – 0.90 µm), MODIS band 2 (0.841–0.876 µm), band 3 of ASTER (0.78 – 0.86 µm), band 3 of Advanced Wide field Sensor (AWIFS) (0.77 - 0.86 µm) of Indian Remote Sensing RESOURCESAT (IRS- P6) satellite. The same band is available in Linear Imaging Scanner Sensors (LISS III and LISS IV) on IRS P6 satellite also. Lake and reservoir area estimates can possible with all these sensors.

Satellite data (Microwave)

Optical remote sensing data has constraints, such as presence of clouds will obstruct the satellite signals from the ground. Especially, during monsoon presence of clouds will limit the optical remote sensing data. In those circumstances, microwave data will be the only viable solution in obtaining information about the snow/ice and also surface water bodies. In microwave range of electro magnetic spectrum both active and passive sensors exists. Active sensor exist on board European Resource Satellite (ERS) – 1 and 2, Japanese Earth Resources Satellite (JERS) – 1 and Radarsat satellites. Back scatter coefficient is the parameter which is calculated to understand the properties of the objects. A passive sensor exists on board are Scanning Multi channel Microwave Radiometer (SMMR) and Special Sensor Microwave Imager (SSM/I). Brightness temperature is the property which is used to determine the various properties of snow and ice as well as surface water in passive microwave.

Snow/Ice. The physical characteristics of the snow pack determine its microwave properties: microwave radiation emitted from the underlying ground is scattered in many different directions by the snow grains within the snow layer, resulting in a microwave emission at the top of the snow surface being less than the ground emission. Properties affecting microwave response from a snow pack include: depth and water equivalent, liquid water content, density, grain size and shape, temperature and stratification as well as snow state (dry/wet) and land cover. These properties make it possible to monitor snow cover using passive microwave remote sensing techniques to derive information of the properties, which are affecting the signal. Information from 19, 37 and 85 GHz channels are used for snow cover mapping in Europe. The 85 GHz data were if use in mapping very shallow snow covers (<5 cm depth). Snow water equivalent (SWE) is also inferred from Synthetic Aperture Radar (SAR) data. Glacier snow line is also mapped using VV SAR data within 50-75 m of ground based measurements. The spatial and temporal patters of lake ice freeze-up can be monitored using passive microwave measurements over large lakes because of large difference between microwave emissivity of the lake ice and fresh water. The resolution is the only constraint in using passive microwave data such as SMMR and SSM/I (~ 25 km) data, which makes it possible to take studies in basins of the size of 10,000 km^2 (Schultz and Engman, 2000)

Surface Water. Microwave remote sensing platforms are also sensitive to water is discrimination and have the distinct advantage of nearly all-weather viewing. Active sensors such as ERS-1 and 2, JERS-1 and Radarsat have shown potential for estimating open water boundaries because of the secular reflection of the incident wave and very low return at the operating angle of the satellites. Low incidence angles will also increase the backscatter response from an open water target and large incidence angles are often recommended for surface water delineation. Polarization can also have large effect on radar back scatter. 'C' band data acquired in HH polarization and at large incidence angles (> 45 degrees) maximized the ability to detect flooded areas through reflection of the incident wave (Barber et. al, 1996).

Hydrologic Modelling

'Hydrological Modeling' is a collection of physical laws, physical based equations, empirical relationships written in the mathematical terms and combined in such a way as to produce a set of results, a set of known and or assumed conditions (Haan, 1982). The fundamental objective of hydrological modeling is to gain an understanding of the hydrological system in order to provide reliable information for managing water resources in a sustained manner to increase human welfare and protect the environment. In this lecture an attempt has been made to briefly describe the snowmelt runoff modeling and also surface runoff modeling.

Snowmelt Runoff Modelling

Snow is a form of precipitation, but in hydrology it is treated some what differently because of log between when it falls and when it produces runoff, ground water recharge and in other hydrologic processes. Snowmelt during the accumulation period is usually very little or nil. Precipitation falling as snow (and some time rain) is temporarily stored in the snow pack until the melt season begins.

Physics of Snowmelt. The physics of melting of snow and transformation of melt water into runoff are very important aspect of snow hydrology. Snowmelt is the overall result of different heat transfer processes to the snow pack. The sun is the ultimate source of energy responsible for the melting of the snow pack. There is a complex interaction between the incoming solar radiation, earth's atmosphere and terrain surface. Hence a number of intermediate steps in the process of energy transfer to the snow surface have to be considered to understand the process of snowmelt and also to make quantitative estimations of the melt (Singh and Singh, 2001). The principal fluxes of heat involved are:

- Absorbed solar radiation,
- Interchange of long wave radiation between the snowpack and the atmosphere and surrounding features,
- Convective heat transfer to the air,
- Latent heat of vaporization released by the vapor condensed from the snow surface,
- Conduction of heat from the ground from the snow pack,
- Heat transferred from the snow pack.

There are two basic approaches generally adopted for estimation of snowmelt from a snow pack. The first approach is known as Energy Budget or Energy Balance Approach and the second is the Temperature Index of Degree-Day Approach.

Energy Budget OR Energy Balance Approach. The energy balance or heat budget of a snow pack governs the production of melt water. This method involves accounting of the incoming energy, outgoing energy, and the change in energy storage for a snow pack for a given period of time. The net energy is then expressed as equivalent of snowmelt. The seasonal variability in the energy inputs available for melt in general increases towards the poles. The energy balance (Singh and Singh, 2001) of the snow pack for any time interval can be expressed as

$$\mathbf{Q_m = Q_{nr} + Q_h + Q_e + Q_p + Q_g + Q_q} \qquad \textbf{(2)}$$

Where,

$\mathbf{Q_m}$ = Energy available for melting of snow pack

$\mathbf{Q_{nr}}$ = Net radiation

$\mathbf{Q_h}$ = Sensible or convective heat from the air

$\mathbf{Q_e}$ = Latent heat evaporation, condensation or sublimation

$\mathbf{Q_p}$ = Heat content of rainwater

$\mathbf{Q_g}$ = Heat gained through conduction from underground

$\mathbf{Q_q}$ = Change of internal energy of the snowpack

The positive value of $\mathbf{Q_m}$ will result in the melting of snow.

Degree-Day Method. In the Himalayan Mountains, the meteorological network for data collection is very poor. The most generally available data are daily maximum and minimum temperatures, humidity measurements and surface wind speed. Temperature indices are widely used in the snowmelt estimation because it is generally considered to be the best index of the heat transfer processes associated with snowmelt. Air temperature expressed in Degree-Day is used in snowmelt computations as an index of the complex energy balance leading to snowmelt.

A degree-day 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 simplest and the most common expression relating daily snowmelt to the temperature index is,

Snowmelt Runoff Model (SRM). The Snowmelt-Runoff Model (SRM: also referred to in the literature as the "Martinec Model" or "Martinec-Rango Model") is designed to simulate and forecast daily stream flow in mountain basins where snowmelt is a major runoff factor. SRM was developed by Martinec (1975) in small European basins. Thanks to the progress of satellite remote sensing of snow cover, SRM has been applied to larger basins. SRM can be applied in mountain basins of almost any size (so far it has been used from 0.76 to 122 000 km^2) and any elevation range (WMO, 1986). 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. 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 of course useful for facilitating the determination of the model parameters.

Model structure Each day, the water produced from snowmelt and rainfall is computed, by the using the equation (Martincec, et.al., 1994);

$$\mathbf{Q_{n+1} = [C_{Sn}\, \boldsymbol{a}_n\, (T_n + \Delta T_n)\, S_n + c_{Rn}\, P_n] * A*(10000/86400) * (1-k_{n+1}) + Q_n\, k_{n+1}} \qquad \textbf{(3)}$$

where:

Q = average daily discharge (m^3s^{-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 [$cm \cdot {}^oC^{-1} \cdot d^{-1}$] indicating the snowmelt depth resulting from 1 degree-day

T = number of degree-days [$^oC \cdot d$]

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

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

P = Precipitation contributing to runoff [cm]. A pre selected threshold temperature, T_{CRIT}, determines whether this contribution is rainfall and immediate. If precipitation is determined by T_{CRIT} to be new snow, it is kept on storage over the hitherto snow free area until melting conditions occur.

A = area of the basin or zone [km^2]

k = recession coefficient indicating the decline of discharge in a period without snowmelt or rainfall:

$$= \frac{Q_{n+1}}{Q_n}$$ (n, n+1 are the sequence of days during a true recession flow period).

N = sequence of days during the discharge computation period

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

The Beas basin up to Pandoh Dam is chosen for the study based on the availability of satellite data and meteorological data. The Beas River catchment up to Pandoh dam is 5406 km^2 out of which only 538 km^2 (from satellite image of October 1998) is under permanent snow. Remote sensing data, hydro-meteorological data and other data have been used. More details of the data used are given here. The snow cover area in the Beas basin during May 1998 - November 1999 was determined using Indian Remote Sensing (IRS) satellite 1C/1D WIde Field Sensor (WIFS) satellite data. WIFS sensor is having 2 bands one each in visible and infer-red bands.

Average monthly deviation of computed runoff with observed runoff is + 4.6 %. The coefficient of determination obtained is 0.854**.** Results of the present study is an improvement over the results of previous study by Kumar, et.al. (1993) in Beas basin in which a computed seasonal snowmelt differ by -5.4 % with observed snowmelt runoff and the coefficient of determination of 0.845 (Vajja & Roy, 2005).

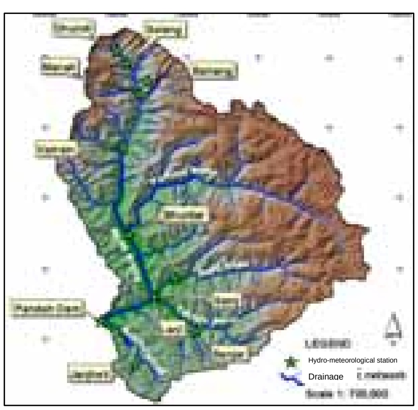

Figure 3 Beas Basin and meteorological Stations

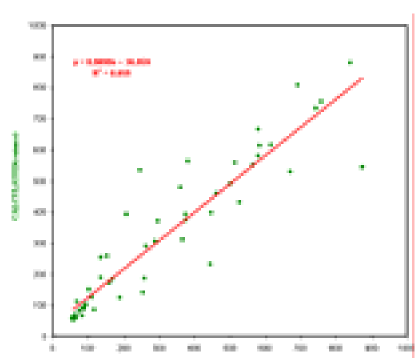

Figure 4 Comparison of observed and calculated discharge .

Surface Runoff Modelling

Remote sensing and Geographic information system is an excellent tool for mapping the earth surface for various thematic layers such as land use land cover, soil and hydrologic condition etc. Using these geospatial data, it is possible to use the existing hydrological models such as Hydrologic Engineering Centre (HEC) – 1 model, Soil Conservation Service (SCS) model, Soil and Water Assessment Tool (SWAT) etc. Among them the most easiest and simple model is SCS model, using which spatial surface runoff potential map can be prepared by integrating land use / land cover, hydrologic soil group maps with runoff curve numbers. Using the rainfall data antecedent moisture condition can be assessed based on 5 day rainfall, which will reflect the hydrologic condition of the area.

Similarly, the lumped model of HEC – 1 can be used in a distributed way to generate the parameters required for various hydrologic methods such as loss methods, surface runoff estimation and river routing for each of the sub –watersheds. Using remote sensing land use / land cover maps can be prepared and using geographic information system parameters such as digital elevation model and its derivatives such as slope, aspect etc can be derived. Applying these data in HEC-1 model will result into the hydrographs at various locations where observed discharge data can be compared. This type of study has been attempted by Prasad and Chakraborti, 1998 in Nagwan Watershed (92.46 km^2) in India, which provided a quite comparable hydrographs at the watershed outlet.

Seasonal water balance can be calculated using a popular models such Thorthwaite and Mather (Thorthwaite and Mather, 1957) and derivatives using remote sensing and GIS techniques. An attempt has been made in Nanakosi watershed (59.43 km^2) to estimate seasonal water balance and also effect of land use / land cover on surface runoff (Singh, et.al., 2004).

Interaction of Surface Water with Groundwater

The mechanism whereby surface water becomes ground water, and vice versa, is important in understanding hydrology of an area because they determine water balances and hydrologic safe yields of aquifers and ground water basins. Surface water becomes groundwater through infiltration of rain ad irrigation water, seepage from streams (called as loosing streams) and canals and artificial recharge with infiltration basins or injection wells. Artificially if water is allowed to move to ground water, it is called artificial recharge, which is used to reduce, stop or even reverse declines of ground water levels. Assuming that the amount of water that escapes evaporation and transpiration moves down below the root zone as "deep percolation" and eventually joins the groundwater, the rate of groundwater accretion can be calculated by the difference between infiltration and evaporation over a long period.

Ground water returns to the atmosphere by evaporation from soil or vegetation, and due to the surface drainage into streams or other surface water, flow from the springs or seeps, discharge of agricultural underground drainage systems, and, of course, flows from pumped or free flowing wells. Using remote sensing techniques it will be possible to identify ground water recharging zones, by identifying some of geomorphic features such as paleo channels, ox-bow lakes, etc. Nilantha & Prasad, 2004, made an attempt in this direction. In this study an attempt has been made to identify the suitable zones for ground water recharge in Chindwara district. More details can be had from the paper (Nilantha and Prasad, 2004).

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