CHAPTER 3

METHODOLOGY TO ESTIMATE EVAPOTRANSPERSION

3.1 Study Area

Hisar and Bhiwani, two districts of Haryana (India) were taken to estimate evapotranspiration through Surface Energy Balance Algorithm using satellite data. Detail description of these study areas has been given in Chapter 1.

Two types of methods have been used to calculate the evapotranspiration

1. Penman Monteith Method
2. Surface Energy Balance Algorithm for Land (SEBAL) Model

To calculate evapotranspiration by these two methods, the important prerequisites are weather data, satellite data and softwares. All these three parameters has been discussed below:

3.1.1 Weather Data

Weather data such as Minimum and Maximum Temperature, Wind Speed, Average Relative Humidity, Sunshine hours for Hisar and Bhiwani districts was taken from Indian Meteorological Department (IMD), Chandigarh. Average monthly weather data for 11 years (1998-2008) was taken from IMD Gurgaon. Monthly weather data of Hisar and Bhiwani districts has been shown in Table 3.1 to 3.2 for the year 1998 and rest of weather data is given in Appendix-1.
Table 3.1 Monthly Weather Parameters for the year 1998 for Bhiwani district

<table>
<thead>
<tr>
<th>1998 Month</th>
<th>Min Temp (°C)</th>
<th>Max Temp (°C)</th>
<th>Relative Humidity (%)</th>
<th>Sunshine Hours</th>
<th>Radiation MJ/m²/day</th>
<th>Rainfall (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>2.5</td>
<td>22.5</td>
<td>63</td>
<td>10.4</td>
<td>16.5</td>
<td>0</td>
</tr>
<tr>
<td>February</td>
<td>5.4</td>
<td>26.3</td>
<td>63</td>
<td>11.3</td>
<td>20.1</td>
<td>16.2</td>
</tr>
<tr>
<td>March</td>
<td>9.5</td>
<td>33.2</td>
<td>62</td>
<td>13.2</td>
<td>25.9</td>
<td>18.6</td>
</tr>
<tr>
<td>April</td>
<td>14.3</td>
<td>41.2</td>
<td>60</td>
<td>15.5</td>
<td>31.8</td>
<td>0</td>
</tr>
<tr>
<td>May</td>
<td>25</td>
<td>46.6</td>
<td>65</td>
<td>14.2</td>
<td>31.1</td>
<td>1.2</td>
</tr>
<tr>
<td>June</td>
<td>20.8</td>
<td>40</td>
<td>67</td>
<td>13.6</td>
<td>30.4</td>
<td>31.2</td>
</tr>
<tr>
<td>July</td>
<td>24.2</td>
<td>40.5</td>
<td>70</td>
<td>12</td>
<td>27.8</td>
<td>63.5</td>
</tr>
<tr>
<td>August</td>
<td>23.3</td>
<td>39.7</td>
<td>70</td>
<td>11.7</td>
<td>26.5</td>
<td>82.6</td>
</tr>
<tr>
<td>September</td>
<td>24.4</td>
<td>37.8</td>
<td>73</td>
<td>9.4</td>
<td>21.4</td>
<td>107</td>
</tr>
<tr>
<td>October</td>
<td>22.9</td>
<td>36.4</td>
<td>73</td>
<td>8.8</td>
<td>17.9</td>
<td>66</td>
</tr>
<tr>
<td>November</td>
<td>15</td>
<td>31.6</td>
<td>68</td>
<td>9.6</td>
<td>16.2</td>
<td>0</td>
</tr>
<tr>
<td>December</td>
<td>2.4</td>
<td>28</td>
<td>65</td>
<td>10.4</td>
<td>15.7</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 3.2 Monthly weather Parameters for the year 1998 for Hisar district

<table>
<thead>
<tr>
<th>1998 Month</th>
<th>Min Temp (°C)</th>
<th>Max Temp (°C)</th>
<th>Relative Humidity (%)</th>
<th>Sunshine Hours</th>
<th>Radiation MJ/m²/day</th>
<th>Rainfall (Mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>3</td>
<td>23.3</td>
<td>63</td>
<td>10.5</td>
<td>16.5</td>
<td>0</td>
</tr>
<tr>
<td>February</td>
<td>6.3</td>
<td>28.6</td>
<td>62</td>
<td>11.8</td>
<td>20.6</td>
<td>21</td>
</tr>
<tr>
<td>March</td>
<td>9.1</td>
<td>34.7</td>
<td>61</td>
<td>13.9</td>
<td>26.7</td>
<td>9</td>
</tr>
<tr>
<td>April</td>
<td>17.5</td>
<td>43.6</td>
<td>61</td>
<td>15.2</td>
<td>31.3</td>
<td>0</td>
</tr>
<tr>
<td>May</td>
<td>17.8</td>
<td>48.8</td>
<td>59</td>
<td>18.3</td>
<td>37.2</td>
<td>0.8</td>
</tr>
<tr>
<td>June</td>
<td>20.8</td>
<td>47.1</td>
<td>62</td>
<td>17.1</td>
<td>35.6</td>
<td>36.5</td>
</tr>
<tr>
<td>July</td>
<td>24.7</td>
<td>42.2</td>
<td>69</td>
<td>12.8</td>
<td>29</td>
<td>84.5</td>
</tr>
<tr>
<td>August</td>
<td>25</td>
<td>42.6</td>
<td>69</td>
<td>12.4</td>
<td>27.5</td>
<td>33</td>
</tr>
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<td>September</td>
<td>21.5</td>
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<td>96</td>
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<td>86.5</td>
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<td>9.3</td>
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<td>62</td>
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<td>16.4</td>
<td>0</td>
</tr>
</tbody>
</table>
3.2 Software used

RoltaGeometica and ILWIS, remote Sensing softwares were used to process the satellite images to estimate the Evapotranspiration by SEBAL model. To estimate the Evapotranspiration by Penman Montieth method, CROPWAT software was used.

3.2.1 RoltaGeometica

It is a software, which contains integrated operations of remote sensing and image processing. After receiving satellite data, all types of radiometric, geometric or atmospheric corrections can be performed in this software with high level of accuracy. Raster spatial analysis can also be done through this software for conducting urban, agricultural and utility mapping and change detection studies. It is suitable for various other image processing operations.

3.2.2 ILWIS Software

ILWIS (The Integrated Land and Water Information System) initially developed by ITC in 2005 and it is a good Remote Sensing and GIS software. Studies related with image processing, spatial analysis and digital mapping can be successfully and accurately performed using this software. It is very user friendly and is freely available after July 2007 and known as ILWIS open.

3.2.3 CROPWAT

CROPWAT is a decision support tool developed by the Land and Water Development Division of FAO. In order to calculate crop water requirement and irrigation requirements based on crop, soil and climate data, CROPWAT 8.0 for Windows was established. This program has also applications for developing irrigation schedules for varying management conditions. It can also be used to estimate crop performance under both irrigated and rainfed conditions. The procedure used for calculation in CROPWAT 8.0 is based on the FAO publication 56 i.e. guidelines for computing crop water requirements. It has one more major advantage that is, if climate data is not available, it can be retrieved from CLIMWAT which contains data of 5000 stations worldwide and it is associated with CROPWAT climatic database.
3.3 Satellite Data

LANDSAT-7 ETM+ data has been used for the present study.

3.3.1 Landsat-7ETM+ Satellite Data

LANDSAT-7 is the most accurately calibrated Earth-observing satellite because measurements done by this are extremely accurate when compared to the same measurements made on the ground. It was launched on April 15 1999 and equipped with ETM+ (Enhanced Thematic Mapper Plus) instrument. This instrument contains eight bands, having capability of multispectral scanning and provides high-resolution image of the Earth planet. Landsat-7ETM+ is improved version of TM (Thematic Mapper) because it has good image resolution (60m) as compared to the Thematic mapper (120 m). LANDSAT-7’s sensor is considered as the most suitable and best characterized Earth observation instrument ever placed in orbit. Extensive validation has been done for this sensor.

3.3.2 Downloading Landsat-7 ETM+ Data

The USGS Global Visualization Viewer (http://glovis.usgs.gov/) managed by the Center for Earth Resources Observation and Science (EROS) constitutes the main browser for ordering Landsat-7 ETM+ data. For the present research work, 7 images of paths/rows 146/40 (Figure 3.1) were downloaded to validate SEBAL Model and 15 cloud free images for Landsat-7 ETM+ scene for paths/rows 147/40 (Figure 3.2) were downloaded to map the Evapotranspiration by SEBAL model.

![Figure 3.1 Image downloading from USGS for paths/rows 146/40](image-url)
3.4 Estimation of Evapotranspiration by Penman Monteith Method

After acquiring weather data, Evapotranspiration was estimated by Penman Monteith Method and detail methodology is explained below:

In 1948, Penman combined the energy balance with the mass transfer method and derived an equation to compute the evaporation from an open water surface using standard climatological records of sunshine, temperature, humidity and wind speed. This so-called combination method was further developed by many researchers and extended to cropped surfaces by introducing resistance factors.

\[
ETo = \left( \frac{\Delta (R_a - G) + \rho_a c_p (e_s - e_a) r_a}{\Delta + \rho (1 + \frac{r_s}{r_a}) \lambda} \right) \lambda
\]

(3.1)

Where

\begin{align*}
ETo & \quad \text{Reference evapotranspiration [mm day}^{-1}] ,
R_a & \quad \text{net radiation at the crop surface [MJ m}^{-2} \text{ day}^{-1}] ,
G & \quad \text{soil heat flux density [MJ m}^{-2} \text{ day}^{-1}] ,
T & \quad \text{mean daily air temperature at 2 m height [Â°C] ,}
e_s & \quad \text{saturation vapour pressure [kPa] ,}
e_a & \quad \text{actual vapour pressure [kPa] .}
\end{align*}
\(e_s - e_a\) saturation vapour pressure deficit [kPa],
\(\Delta\) slope vapour pressure curve [kPa °C\(^{-1}\)],
\(\gamma\) psychrometric constant [kPa °C\(^{-1}\)]
\(\rho_a\) mean air density at constant pressure [kg m\(^{-3}\)]
\(r_s\) bulk surface resistance [s m\(^{-1}\)]
\(r_a\) aerodynamic resistance [s m\(^{-1}\)]
\(\lambda\) latent heat of vaporization [MJ kg\(^{-1}\)]

From the original Penman-Monteith equation and the equations of the aerodynamic and surface resistance, the FAO Penman-Monteith method to estimate ETo can be derived.

The aerodynamic resistance for neutral stability condition is

\[
r_a = \frac{\ln\left(\frac{z_w}{z_{om}}\right)\ln\left(\frac{z_h}{z_{oh}}\right)}{k^2u_z} \tag{3.2}
\]

Where
\(r_a\) = aerodynamic resistance
\(z_w\) = height of wind measurements (m)
\(z_h\) = height of humidity and or air temperature measurements (m)
\(z_{om}\) = roughness length governing momentum transfer
\(z_{oh}\) = roughness length governing transfer of heat and vapour
\(k\) = von Karman’s constant
\(u_z\) = wind speed at height \(z\) (ms\(^{-1}\))
\(h\) = mean height of vegetation, m

Bulk surface resistance is

\[
r_s = \frac{r_s}{LAI_{active}} \tag{3.3}
\]

Where
\(r_s\) = bulk surface resistance [s m\(^{-1}\)]
\(r_1\) = bulk stomatal resistance of a well illuminated leaf [s m\(^{-1}\)]
\(LAI_{active}\) = active (sunlit) leaf area index [m\(^2\) (leaf area) m\(^{-2}\) (soil surface)]

Generally with reference crops that have full ground cover
\(LAI_{active} = 0.5\ LAI\)
For clipped grass:

\[ \text{LAI} = 24 \text{ h} \]

Where \( h \) is vegetation height (m)

In Ref-ET, \( \text{LAI} = 23.8 \) is used so that \( r_s \) for 0.12 m grass calculations as \( r_s = 70 \text{ s m}^{-1} \)

For alfalfa = \( 5.5 + 1.5 \ln(h) \)

After incorporating the aerodynamic and bulk resistance, this method became FAO standard method[27] which is widely used and most accurate point method.

\[
\text{ETo} = \frac{0.408 \Delta (R_n - G) + \gamma \frac{900}{\Delta + \gamma(1+0.34u_2)}}{ \Delta + \gamma(1+0.34u_2)} \tag{3.4}
\]

Where

- \( \text{ETo} \): Reference evapotranspiration [mm day\(^{-1}\)],
- \( R_n \): net radiation at the crop surface [MJ m\(^{-2}\) day\(^{-1}\)],
- \( G \): soil heat flux density [MJ m\(^{-2}\) day\(^{-1}\)],
- \( T \): mean daily air temperature at 2 m height [°C],
- \( u_2 \): wind speed at 2 m height [m s\(^{-1}\)],
- \( e_s \): saturation vapour pressure [kPa],
- \( e_a \): actual vapour pressure [kPa],
- \( e_s - e_a \): saturation vapour pressure deficit [kPa],
- \( \Delta \): slope vapour pressure curve [kPa °C\(^{-1}\)],
- \( \gamma \): psychrometric constant [kPa °C\(^{-1}\)]

The equation uses standard climatological records, measured at well maintained weather stations. To ensure the integrity of computation, the weather measurements should be made at 2m above an extensive surface of green grass, shading the ground and not short of water.

**Data:** Calculation of \( \text{ETo} \) with the Penman-Monteith equation on 24-hour time scales will generally provide accurate results. The required meteorological data consist of:

- Air temperature: maximum (Tmax) and minimum (Tmin) daily air temperatures.
- Air humidity: mean daily actual vapour pressure (ea) derived from psychrometric, dewpoint temperature or relative humidity data.
• Wind Speed: daily average for 24 hours of wind speed measured at 2m height (u_2)
• Radiation: Net radiation (Rn) measured or computed from solar and longwave radiation or from the actual duration of sunshine (n). The extraterrestrial radiation (Ra) and daylight hours (N) for a specific day of the month should be computed. As the magnitude of daily soil heat flux (G) beneath the reference grass surface is relatively small, it may be 24-hour time steps.

3.5 ET estimation using SEBAL (Surface Energy Balance Algorithm for Land) model

After acquiring satellite data, Evapotranspiration was estimated using SEBAL model. Detail methodology for estimating the ET is illustrated below:

SEBAL: The Surface Energy Balance Algorithm for Land (SEBAL) is an image-processing model comprised of 25 computational steps, which help in the estimation of the actual (ET_{act}) as well as other energy exchanges between land and atmosphere [81].

3.5.1 Surface energy budget equation

Applied the residual approaches of surface energy balance for estimating the ET at spatial and different temporal scales by using the surface energy budget equation (SEBAL). The total solar energy coming from the sun and atmosphere in the form of longand shortwave radiation is transformed and used for (a) soil heating, (b) surface environment heating (sensible heat flux to the atmosphere), and (c) transforming water into vapour (latent heat flux from the crop or soil surfaces). All the energy involved in the soil vegetation atmosphere interface can be given as the Energy Balance equation:

$$Rn = G + H + \lambda ET$$  \hspace{1cm} (3.5)

Where Rn is the net radiation at the surface (W/m²), G is the soil heat flux (W/m²), H is the sensible heat flux to the air (W/m²), \lambda ET is the latent heat flux associated with ET (instantaneous value for the time of the satellite overpass, W/m²). Flow chart to calculate various parameters is shown in Figure 3.
3.5.2 Theoretical Basis of SEBAL

After receiving satellite data through remote sensing techniques visible, near-infrared and thermal infrared radiation were measured, SEBAL uses this data and evapotranspiration is computed as a residual of the energy balance on a pixel-by-pixel basis:

\[
LE_{\text{pixel}} = R_{n_{\text{pixel}}} - H_{\text{pixel}} - G_{\text{pixel}}
\]  

(3.6)
Where $LE_{\text{pixel}}$ is the latent heat flux for the pixel, and $R_{n\text{pixel}}$, $H_{\text{pixel}}$ and $G_{\text{pixel}}$ are the net radiation, sensible heat flux and soil heat flux for each pixel, respectively. A general description of the SEBAL model is presented in Figure 3.3

### 3.6 Components of SEBAL

Components of the original SEBAL model developed by Bastiaanssen et al. (1998) [59] are described below

#### 3.6.1 Net Radiation (Rn)

The net Radiation is the main term in energy balance equation because it derives all the process of evapotranspiration. Net radiation itself can be represented by a budget equation. The net radiation is the sum of the incoming and outgoing short and long wave components.

An effective method to calculate net radiation by summing up net shortwave (Rns) and net longwave radiation (RnL) at the surface, which can be represented as:

$$R_n = R_{ns} + R_{nL} \quad (3.7)$$

Net shortwave radiation can be further represented by the sum of incoming shortwave and outgoing shortwave radiations

$$R_{ns} = R_s\downarrow + R_s\uparrow \quad (3.8)$$

Short wave radiation lies between the wavelength of 0.3- 3 µm. Arrows show the direction of the flux entering ↓ or leaving ↑ the system. Pyranometers are the instruments, which are used to measure incoming shortwave radiation. These instruments usually work in all visible broadband ranges (usually 0.305 – 2.4 µm). This range comprises almost 96% of the spectral interval of the solar irradiance. Only the outgoing radiations can be estimated by remote sensing which are associated with reflective properties of terrestrial bodies.

$$R_{nL} = R_L\downarrow + R_L\uparrow + (1-\epsilon_o)* R_L\downarrow \quad (3.9)$$

Longwave radiations emitted by the surface are represented by $R_L\uparrow$, means outgoing longwave radiations and longwave radiation represented by $R_L\downarrow$ are radiations emitted by atmosphere that reaches the earth surface. Amount of
\( R_L \downarrow \) which are reflected back to atmosphere represented by \((1-\varepsilon_o)^* R_L \downarrow\), where \( \varepsilon_o \) is broad band surface emissivity.

Reflected net shortwave radiations can also be represented as \((1-\alpha)^* R_s \downarrow\) and \( \alpha \) is the albedo of shortwave radiation and values[97] are given in Table 3.3. Thus final net radiation flux can be calculated by equation 3.10 and Figure 3.4 below shows the net radiation balance at the earth surface.

\[
R_n = (1-\alpha)^* R_s \downarrow + R_L \downarrow - R_L \uparrow - (1-\varepsilon_o)^* R_L \downarrow
\]  

(3.10)

![Figure 3.4 Radiation Balance at the Earth Surface](image)

Table 3.3 Albedo Values for Different Surfaces[82]

<table>
<thead>
<tr>
<th>Surface</th>
<th>Albedo</th>
</tr>
</thead>
<tbody>
<tr>
<td>Green grass and other short vegetation</td>
<td>0.15 - 0.25</td>
</tr>
<tr>
<td>Coniferous forest</td>
<td>0.10 - 0.15</td>
</tr>
<tr>
<td>Dry soils; deserts</td>
<td>0.20 - 0.35</td>
</tr>
<tr>
<td>Gray soils; bare fields</td>
<td>0.15 - 0.25</td>
</tr>
<tr>
<td>White sand; lime</td>
<td>0.30 - 0.40</td>
</tr>
<tr>
<td>Moist dark soils</td>
<td>0.05 - 0.15</td>
</tr>
<tr>
<td>Deep water</td>
<td>0.04 - 0.08</td>
</tr>
<tr>
<td>Fresh dry snow</td>
<td>0.80 - 0.90</td>
</tr>
</tbody>
</table>

For calculating the net radiation flux, equation 3.10 was solved. Each component of this equation has been explained below:
3.6.1.1 Incoming Short wave Radiation \( (R_s \downarrow) \): The incoming short wave radiation \( R_s \downarrow \) is estimated from the radiation received at the top of the atmosphere as:

\[
R_s \downarrow = R_a \tau_{sw} \tag{3.11}
\]

\( R_a \) represents the radiations received at the top of atmosphere. \( \tau_{sw} \) is the transmittance for the atmosphere and it takes into the consideration effects of scattering and absorption by the atmosphere.

The amount of radiation receive at the top of the atmosphere, \( R_a \), is estimated as:

\[
R_a = G_{sc} \cos \theta \cdot dr \tag{3.12}
\]

\( G_{sc} \) is the solar constant values 1367 W/m\(^2\) and \( \cos \theta \) is the cosine of solar Zenith angle and \( \text{and} \cdot dr \) is the inverse relative squared distance of Earth to Sun.

\[
dr = 1 + 0.33 \cos \left( \text{DOY} \cdot \frac{2\pi}{365} \right) \tag{3.13}
\]

\( \text{DOY} \) is the day of the year.

For a flat topography, the cosine of the solar incident angle is calculated from the solar elevation angle as:

\[
\cos \theta = \cos \left( \frac{\pi}{2} - \phi \right) \tag{3.14}
\]

Where \( \phi \) is the angle in radians.

3.6.1.2 Incoming Longwave Radiation \( (R_L \downarrow) \): Stephan-Boltzman equation is used to calculate the incoming longwave radiation, emitted by the atmosphere:

\[
R_L \downarrow = \varepsilon_a \sigma T_a^4 \tag{3.15}
\]

Where \( \varepsilon_a \) is the atmospheric emissivity (dimensionless), \( \sigma \) is the Stefan-Boltzmann constant \( (5.67 \times 10^{-8} \text{ W/m}^2/\text{K}^4) \), and \( T_a \) is the air temperature in K.

3.6.1.3 Outgoing Longwave Radiation \( (R_L \uparrow) \): The Stephan-Boltzman equation is used to estimate the long wave radiation emitted by the surface:

\[
R_L \uparrow = \varepsilon_o \sigma T_s^4 \tag{3.16}
\]

Where \( \varepsilon_o \) is the broadband surface emissivity, \( \sigma \) is the Stefan-Boltzmann constant \( (5.67 \times 10^{-8} \text{ W/m}^2/\text{K}^4) \), and \( T_s \) is the surface temperature in K.

3.6.1.4 Surface Albedo (\( \alpha \)): Surface Albedo means reflection coefficient and it is the ratio of reflected radiations to the incident radiations.

\[
\alpha = \frac{R_s \uparrow}{R_s \downarrow} \tag{3.17}
\]

Where \( \alpha \) is the albedo, \( R_s \downarrow \) is the incoming short wave radiation, and \( R_s \uparrow \) is the outgoing short wave radiation.
In the SEBAL procedure, the hemispherical surface albedo ($\alpha$) is obtained from the broadband directional planetary reflectance ($\alpha_{toa}$). The surface albedo\cite{46} can be estimated with the following equation:

$$\alpha = \frac{\alpha_{toa} - \alpha_{path\_radiance}}{\tau_{sw}^2} \quad (3.18)$$

Where $\alpha$ is the surface albedo, $\alpha_{toa}$ is the clear-sky shortwave hemispherical albedo at the top of the atmosphere, $\alpha_{path\_radiance}$ is the albedo path radiance, and $\tau_{sw}$ is the broadband shortwave atmospheric transmittance.

The part of the incoming radiation that is reflected by the atmosphere prior to it reaches the surface is known as path radiance which can be sensed by satellite causing in the increase of the amount of measured radiation. Approximate value of path radiance lies between 0.025 to 0.04.

The albedo at the top of the atmosphere, $\alpha_{toa}$ is estimated by integrating all reflectance of bands in the visible and near-infrared region of the spectrum:

$$\alpha_{toa} = \int_{0.3}^{3.0} \rho(\lambda) d\lambda = \sum_{i=1}^{n} w(\lambda) \rho_{\lambda i} \quad (3.19)$$

Where “n” represents the total number of spectral bands “i” of the sensor corresponding to the 0.3 to 3.0$\mu$m region of the spectrum, $w(\lambda)$ is a weighting factor that accounts for the uneven distribution of the extraterrestrial radiation for each narrow band region of the spectrum, and $\rho_{\lambda i}$ is the narrow band spectral reflectance ($\rho_{\lambda}$) corresponding to band i.

**3.6.1.5 Surface Emissivity, $\epsilon_o$:** According to Plank’s law the emissivity of an object is the ratio of the energy radiated by that object at a given temperature to the energy radiated by a blackbody at the same temperature. Surface temperature from thermal band can be calculated if the emissivity of land surface is known.

An empirical method derived is used to estimate surface emissivity using NDVI in SEBAL model

$$\epsilon_o = 1.009 + 0.047 \ln \text{(NDVI)} \quad (3.20)$$

NDVI is Normalised Difference Vegetation Index which has been explained later in this section.
3.6.1.6 Surface Temperature $T_s$

Band 6 from satellite LANDSAT7 ETM+ images is converted from spectral radiance to surface temperature. Using Plank's law and pre-launch calibration constants for satellite imagery, equation 3.21 has been used for deriving the surface temperature.

$$T = \frac{k^2}{\ln\left(\frac{kT}{\lambda} + 1\right)}$$  \hspace{1cm} (3.21)

3.6.2 Soil Heat Flux ($G$)

Due to conduction, the rate of heat storage into the soil and vegetation is known as Soil Heat Flux. This is the parameter, which cannot be directly estimated from the satellite sensors so its estimation needs empirical formulas. Thus by considering influence of some parameters (Bastiaanssen and Robeling, 1993) [98] like surface temperature, albedo and incoming shortwave radiation, empirical relations are derived. The evaluation of $G$ is usually presented as a ratio $G/R_n$. The instantaneous $G/R_n$ function depends on pixel size, location and time. Over a period of one day, the integrated soil heat flux is usually considered as negligible.

Bastiaanssen and Roebeling (1993) [98] introduced an equation based on their own research (Egypt and applied in HAPEX-EFEDA experiment, 1992) and other soil researchers.

$$\frac{G}{R_n} = \frac{(T_s - 273.15)}{100\alpha} \left(0.32\alpha + 0.62\alpha^2\right)(1 - 0.98NDVI^4)$$  \hspace{1cm} (3.22)

Where $\alpha$ is the average albedo (approx 1.1 $\alpha$) when the soil heat flux is directly downward. Note that $T_s$ is the surface temperature in Kelvin. Equation 3.22 explains reasonably well the wide range of variability of the soil heat fluxes on clear days even in rich relief terrain, mainly for low NDVI values.

Very simple approaches neglect the value of $G$ for very dense canopies or give a fixed percentage of the incoming solar global radiation. Seguin et. al (1983) [99] has revealed the simplest approaches used by several authors based on this fixed relation.

$$G = 0 \text{ for very dense vegetation.}$$  \hspace{1cm} (3.23)

$$G/R_n = 0.1 \text{ for normal vegetation}$$  \hspace{1cm} (3.24)

$$0.2 < G/R_n \leq 0.3 \text{ for bare soil}$$  \hspace{1cm} (3.25)
Different empirical studies have shown that the daytime ratio $G/R_n$ is related to other factors, also like the amount of vegetation present[100]. Thus, an approximation of $G$ can be achieved by assuming that it is a fraction of $R_n$ dependent on the spectral estimates of surface vegetation cover. Jackson et al, (1987) [101] used an exponential relation (equation 3.26).

$$\frac{G}{R_n} = 0.58 \exp[-2.13 \text{NDVI}]$$  \hspace{1cm} (3.26)

Where NDVI is a spectral index that estimates the amount of vegetation present based on the normalized difference between near-infrared (NIR) and red (R) bands reflectance. Kustas and Daughtry (1990) [102] proposed two linear expressions equation 3.27 and 3.28.

$$\frac{G}{R_n} = 0.325 - 0.208 \text{NDVI}$$  \hspace{1cm} (3.27)

$$\frac{G}{R_n} = 0.294 - 0.0164 \text{NDVI}$$  \hspace{1cm} (3.28)

Clothier et al. (1986) [103] suggested a similar relation between $G$, $R_n$ and NIR/Red reflectance given by equation 3.29.

$$\frac{G}{R_n} = 0.295 - 0.0133 \text{NIR/RED}$$  \hspace{1cm} (3.29)

### 3.6.3 Sensible Heat Flux (H)

After calculating the above two fluxes viz Net radiation Flux and Soil heat flux, the crucial component of SEBAL model is the estimation of sensible heat flux. Estimation of sensible heat flux will lead to the estimation of evapotranspiration as a residual form of energy balance. The sensible heat flux ($H$) is the flow of energy through air as a result of the temperature gradient. The mathematical formulation of the sensible heat flux is based on the theory of mass transport of heat and momentum between the surface and the near surface environment. The expression can be written as

$$H = \rho_a C_p \frac{dT}{r_{ah}}$$  \hspace{1cm} (3.30)

Where $\rho_a$ represents the air density (Kg/m$^3$), $C_p$ represents air specific heat (1004J/kg/K), $dT$ (K) represents the temperature difference ($T_1$-$T_2$) between two heights ($Z_1$ and $Z_2$) and $r_{ah}$ represents the aerodynamic resistance to heat transport.

$$r_{ah} = \frac{\ln(Z_2/Z_1) - \psi_h(Z_2) - \psi_h(Z_1)}{u_* \times k}$$  \hspace{1cm} (3.31)
$Z_1$ and $Z_2$ are the two heights, $u^*$ represents friction velocity, $k$ is Von Karman constant (0.41) and $\psi_h$ is the integral stability correction factor, the corresponding factor between surface and $Z_1$ is $\psi_h(Z_1)$ and the corresponding factor between surface and $Z_2$ is $\psi_h(Z_2)$.

In sensible heat flux equation, there are two unknown factors, thus solution of this equation is only possible through iteration procedure. Scientist suggested that this iteration procedure is based on Monin-Obukov, similarity-stability theory to compute the sensible heat flux. Detailed procedure to calculate the sensible heat flux is described below:

### 3.6.3.1 Iterative Procedure for Sensible Heat Flux:

1. The input needed for this procedure is the wind speed corresponding to the time for which satellite image of the study area with time of satellite overpass from the location under study was used. By using equation 3.32 wind speed is calculated by the formula
   \[
   u^* = \frac{u_x \times k}{\ln \left( \frac{Z_x}{Z_{om}} \right)} \tag{3.32}
   \]
   $u_x$ is wind speed measured at height $Z_x$ at the weather station.
   $Z_{om}$ is empirically estimated from the vegetation height
   \[
   Z_{om} = 0.123 \times h \tag{3.33}
   \]
   $h$ is vegetation height in meters.

2. Wind speed is calculated at some blending height in the atmosphere, where it is assumed that wind velocity is not affected by surface roughness elements. This blending height is assumed to be at 200 meter and $u_{200}$ is calculated as
   \[
   u_{200} = u^* \frac{\ln \left( \frac{200}{Z_{om}} \right)}{k} \tag{3.34}
   \]

3. $u_{200}$ (blended wind speed) in the model considered to be unaffected by surface features and it is equal for all pixels. But it still can be extrapolated to specific pixels and land cover conditions using equation 3.34, using the specific value of $Z_{om}$ for particular pixel.

4. Firstly by considering neutral conditions ($\psi_h(Z_1) = \psi_h(Z_2) = 0$), and using equation 3.31 the initial value of aerodynamic resistance is calculated.
To calculate $dT$ for equation 3.30 (surface temperature difference) for each pixel, in SEBAL model a linear relationship is considered between $dT$ and surface temperature $Ts$.

$$dT = a Ts + b$$  \hspace{1cm} (3.35)

Where $dT$ is the near-surface air temperature difference, $Ts$ is the surface temperature, and "a" and "b" are empirical coefficients.

SEBAL model considers an assumption that hot areas cause higher vertical differences in the air temperature $dT$ than cold surfaces and that this relationship is linear. Even this relationship has been also validated by field experiments in China[104-105] and Kenya [106].

To determine empirical coefficients "a" and "b", SEBAL model incorporates the selection of two extreme pixels, which are also known as "anchor" pixels. Anchor pixels involve the concept of hot and cold pixels.

**Hot Pixel** (dry pixel): To locate the hot pixel on satellite image, image has to be explored for highest or near highest surface temperature. Indeed the hot pixel should be dry or no moisture should be available for evaporation. ET for hot pixel is presumed to be Zero. In current study bare land with no vegetation is considered as hot pixel.

**Cold Pixel**: The pixel where all of the available energy ($Rn-G$) is converted into evaporation or evapotranspiration is considered as cold pixel. In current study highly water vegetation area is considered as cold pixel.

When hot and cold pixels are defined, then $dT$ function permits for the prediction of $dT$ for each pixel based on the corresponding pixel surface temperature. First approximation value of Sensible heat flux can be estimated using the predicted $dT$.

6. For atmospheric stability adjustment of aerodynamic resistance: In SEBAL model by assuming neutral atmospheric conditions, estimation of sensible heat flux is obtained first. In the lower atmosphere buoyancy effects, which cause surface heating is taken into account by applying the Monin-Obukov similarity theory. This theory is applied by iterative procedure as mentioned below:

6.1 Monin-Obukov length : This parameter is calculated as given by Monteith and
Unsworth (1990) [107]

\[ L = \frac{\rho_a C_p u_*^2 T_s}{k g H} \]  (3.36)

Where \( \rho_a \) is the air density, \( C_p \) is the specific heat, \( T_s \) is the surface temperature in Kelvin, \( g \) is gravitational acceleration (= 9.81 m/s\(^2\)), \( k \) is the von Karman constant = 0.41, and \( H \) is the sensible heat flux in W/m\(^2\).

The value of \( L \) defines the stability condition of the atmosphere. If \( L < 0 \) (\( H > 0 \)) the atmosphere is unstable.

\( L > 0 \) indicates stable conditions,

\( L = \infty \) (\( H = 0 \)), neutral conditions.

In equation 3.36 for estimation of \( L \) instead of \( T_a \), \( T_s \) is used because uncertainties in \( T_a \) give inaccurate results. Sensitivity analysis [108] performed also showed the variation in results if \( T_a \), used instead of \( T_s \).

6.2 The values of the stability parameters for momentum and heat transfer depending on the atmospheric stability condition are determined as follows [109]

If \( L < 0 \)

\[ \Psi_{m(200m)} = 2 \ln \left( \frac{1 + x_{(200m)}}{2} \right) + \ln \left( \frac{1 + x_{(200m)}^2}{2} \right) - 2 \arctan(x_{(200m)}) + 0.5 \pi \]  (3.37)

\[ \Psi_{h(z1)} = 2 \ln \left( 1 + \frac{x_{(z1)}^2}{2} \right) \]  (3.38)

\[ \Psi_{h(z2)} = 2 \ln \left( 1 + \frac{x_{(z2)}^2}{2} \right) \]  (3.39)

\( x_{(200)} = (1 - 16 ((200))/L )^{0.25} \)  (3.40)

\( x_{(z1)} = (1 - 16 ((Z1))/L )^{0.25} \)  (3.41)

\( x_{(z2)} = (1 - 16 ((Z2))/L )^{0.25} \)  (3.42)

and if \( L > 0 \)

\[ \Psi_{h(z1)} = -5 \left( \frac{Z_1}{L} \right) \]  (3.43)

\[ \Psi_{h(z2)} = \Psi_{m(200m)} = -5 \left( \frac{Z_2}{L} \right) \]  (3.44)

6.3 A new value of the friction velocity \( u_* \) is determined for each pixel as:

\[ u_* = \frac{k \times u_{200}}{\ln \left( \frac{200}{\delta_{om}} \right) - \Psi_{m(200m)}} \]  (3.45)
This equation considers that instability of the surface of each pixel impacts the wind profile. Due to this it is presumed that neighbouring pixels are similar in aerodynamic and energy balance characteristics. This is only possible when large agriculture fields exist.

\[
H = \rho C_p \frac{dT}{r_{ah}}
\]

\[
L = \rho a c \rho \frac{u_2^3 T_s}{k g H}
\]

\[
\Psi_{m(200m)} \Psi_{h(x1)} \Psi_{h(x2)}
\]

\[
u_* = \frac{u_2 \times k}{\ln(Z_{200}/Z_{om})}
\]

\[
r_{ah} = \frac{\ln(Z_2/Z_1)}{u_* \times k}
\]

\[
H_{cold} = Rn - G
\]

\[
dT_{cold} = 0
\]

\[
H_{hot} = Rn - G
\]

\[
dT_{hot} = H_{hot} \times (r_{ah}/(\rho C_p))
\]

\[
dT = aT_s + b
\]

Figure 3.5 Iterative procedure for determination of Sensible heat flux for SEBAL Model.
6.4 After taking this new $u^*$ by using equation 3.45, a stability corrected value of the aerodynamic resistance was computed.

6.5 By rearranging equation 3.30, using new value of $r_{ah} \cdot dT$ for hot pixel was determined.

This new value of $dT$ will change the relationship between $dT$ versus $T_s$, which further leads to new determination of sensible heat flux for each pixel. Until the values of $r_{ah}$ become stable, this iterative process for $r_{ah}$, $dT$, and $H$ continues. This iterative procedure is also shown in Figure 3.5.

3.6.4 Determination of Daily Evapotranspiration

After determination of Sensible heat flux for SEBAL Model using iterative procedure, the latent heat flux LE for each pixel is calculated from equation 3.6 incorporating the values of net radiation flux as well as Ground heat Flux. This latent heat flux is instantaneous evapotranspiration at the time of satellite overpass.

To determine the daily evapotranspiration, concept of evaporative fraction is applied, which is the ratio of latent heat flux and the available energy and assumed that it remains constant throughout the day [110]

So the evaporative fraction ($\Lambda$)

$$\Lambda_{inst} = \frac{LE}{R_n - G} \quad (3.46)$$

$$\Lambda_{inst} = \frac{R_n - G - H}{R_n - G} \quad (3.47)$$

When $\Lambda_{inst} \approx \Lambda_{24}$

$$LE_{24} = \Lambda_{24}(R_{n24} - G) = \Lambda_{inst}(R_{n24} - G) \quad (3.48)$$

Where $R_{n24}$ is the 24 hours net radiation and the 24 hour value of soil heat flux ($G$) is usually ignored in this equation for simplicity. Hence the expression for the daily evapotranspiration ($ET_{24}$) can be expressed as.

$$ET_{24} = \frac{LE_{24}}{\Lambda_v} = \frac{\Lambda_{inst} R_{n24}}{\Lambda_v} \quad (3.49)$$
The final equation that is used to evaluate the daily evaporation is based on the evaporative fraction:

\[ ET_{24} = \frac{8.64 \times 10^7 \Lambda_{ins} R_{n24}}{\lambda \rho_v} \]  

(3.50)

In above equation 3.50 daily net radiation is given in Wm\(^{-2}\), \(\lambda\) is 2.47\times10^6 JKg\(^{-1}\) and \(\rho_v\) is 1000 kgm\(^{-3}\). The final form of equation is given by equation 3.51

\[ ET_{24} = \frac{\Lambda_{ins} R_{n24}}{28.588} \]  

(3.51)

### 3.6.5 Vegetation Indices

There are three vegetation indices which have been used in SEBAL (Surface Energy Balance Algorithm for Land) model which are used further to determine all energy balance components that are explained above.

**3.6.5.1 Leaf Area Index**

**3.6.5.2 Soil Adjusted Vegetation Index**

**3.6.5.3 Normalised Difference Vegetation Index**

**3.6.5.1 Leaf Area Index (LAI):** Leaf Area index is dimensionless quantity and it is totally one sided area of tissue per unit ground surface area [111]. It characterises the canopy of ecosystem. It is very important and key component of ecosystem because it controls radiation extinction, canopy water interception, water and carbon gas exchange. If some factors like frost, defoliation, storm and management practice alters canopy leaf area index then it leads to change in stand productivity. It is also a key input to many more models, which do simulation for crop yield analysis. Thus this parameter is also used by ecologists, farm managers, ecophysiologists and global modellers need information about canopy leaf area index.

\[ LAI = \frac{SAVI - C_1}{C_2} \]  

(3.52)

**3.6.5.2 Soil Adjusted Vegetation Index (SAVI):** This index represents the characteristics for the two indices viz ratio based indices and perpendicular indices. This index is based on the concept that isovegetation lines do not converge at a single point and they are not parallel. Initially this index was developed on
measurements of cotton and range grass canopies with dark and light soil backgrounds. In its calculation a correction factor ‘L’ is needed which was initially determined by trial and error method until this factor gives equal vegetation index results for dark and light soils. The value of correction factor was lying between 0 and 1 viz 0 for high densities and 1 for very low densities. In most of the applications this value is taken as 0.5, which is for intermediate vegetation densities. In case of water body it shows negative value.

\[
SAVI = \frac{(1 + L)(NIR - R)}{NIR + R + L}
\]  

(3.53)

Where NIR and R are the reflectance values of near infrared band and red band of spectrum respectively and L is a correction factor

3.6.5.3 Normalized Difference Vegetation Index (NDVI): This index purely represents the vegetation. It measures the amount and vigor of vegetation at the surface. It is determined by using reflectance values of near infrared band and red band of the spectrum owing to the fact that healthy vegetation shows good reflection in the near infrared band. Green leaves have a reflectance of 20% or less in the 0.5 to 0.7 range (green to red) and about 60 percent in the 0.7 to 1.3 \( \mu \text{m} \) range (near infrared). Then this value is normalized to \(-1 \leq \text{NDVI} \leq 1\) to incorporates the effects of differences in illumination and surface slope. Due to this it is known as Normalized Difference Vegetation Index. It is determined by following formula.

\[
\text{NDVI} = \frac{\text{NIR} - \text{R}}{\text{NIR} + \text{R}}
\]  

(3.54)

3.7 Chapter Conclusion

• Monthly weather data for 11 years (1998-2008) of Hisarand Bhiwani districts has been presented for estimating the ET trend and associated components.
• Remote sensing and GIS softwares (RoltaGeometica and ILWIS) were used for image processing of LANDSAT7 ETM+ satellite data. Image corrections were performed before processing the data to minimise the errors.
• Different methodologies used for the estimation of evapotranspiration like Penman-Monteith Method and SEBAL model has been discussed in detail along with their limitations and advantages.
• Methodologies to calculate vegetation indices like NDVI, SAVI and LAI have important role in ET estimation and same has been illustrated in detail.