II. LITERATURE REVIEW

2.1 Drought Indicator

Drought indicator is divided in three types (Voght, et. al., 1999). First, is meteorological drought indicators based on meteorological parameters as recorded at meteorological stations. An example is the Standardized Precipitation Index (SPI). The SPI is a statistical indicator evaluating the lack or surplus of precipitation during a given period of time as a function of a long term normal precipitation to be expected during that period. It is calculated using a continuous, long term (more than 30 years) series of historic monthly precipitation records. A moving window is selected (1, 3, 6, 12, 24 months, depending on the purpose of the analysis) and new series generated. Because rainfall is not normally distributed for aggregation periods of less than 12 months a gamma distribution is fitted to the frequency distribution. The SPI for a given rainfall amount is then given by the precipitation deviation from the mean of an equivalent normally distributed probability distribution function with a zero mean and standard deviation of one resulting in units of standard deviation. This is an advantage since the SPI is normalized so that wetter and drier climates can be represented in the same way. In addition, wet periods can be monitored as well.

Second, is satellite based indicators calculated from satellite derived surface parameter. Examples are various vegetation indices such as the Normalized Difference Vegetation Index (NDVI), the Vegetation Condition Index (VCI) and the Temperature Condition Index (TCI). Vegetation indices can be efficient indicators of water stress in relatively homogenous terrain. However, in more heterogeneous regions their interpretation becomes more difficult. The VCI
is an indicator of the status of the vegetation cover as a function of the NDVI minimum and maximum encountered for a given ecosystem over many years. It normalizes the NDVI (or any other vegetation index) and allows for a comparison of different ecosystems. It is an attempt to separate the short-term climatic signal from long-term ecological signal and in this sense it is a better indicator of water stress condition than the NDVI. The significance of the VCI is strongly related to the relation between the vegetation index and the vitality of the vegetation cover under investigation. In addition, it depends on the number and quality of images available for the calculation of the absolute minimum and maximum. The Temperature Condition Index (TCI) is an equivalent indicator based on the surface skin temperature derived from Landsat data. Both, the VCI and the TCI are dimensionless and vary between the values of zero and one. Zero indicates the worst condition ever encountered over the period of available images, one indicates the best condition encountered during the same period of time. If the covered period includes dry and wet years and under the assumption that the vegetation condition is mainly related to the water availability, these indicators have a high potential for monitoring water stress.

Third, a process base indicator is the result of modeling of energy and matter transfer between the atmosphere and the surface. An example is evaporative fraction, EF. EF is defined as the part of the available energy used for evapotranspiration, i.e., the latent heat flux. This quantity is regarded as an indicative of the moisture status of the surface cover.

The remote sensing, especially satellite remote sensing, can observe land, ocean and atmosphere from the space. In most applications, remote sensing data are converted into physical parameters or indices. The quantitative and qualitative
analysis based on the same criteria can be implemented by using remote sensing data. Moreover, most of remote sensing data are processed as images, and then area-based information can be derived. Area-based information is quite effective to drought indicator.

Monitoring by satellite remote sensing techniques is most appropriate for detecting the status of soil moisture, evapotranspiration, crop growth, land cover change and drought. Using a surface observation station network and remote sensing techniques, the development and spread of drought conditions can be monitored in a routine and cost-effective manner. A method based on energy balance was developed to estimate evapotranspiration (ET) and evaporative fraction (EF) to monitor drought using Landsat-7 ETM+ digital images and meteorological data. The Landsat-7 ETM+ data were used to compute reflectance and temperature after atmospheric correction. ET was estimated by combining remotely sensed reflected solar radiation and surface temperature with ground station meteorological data to calculate net radiation and sensible flux.

In principle, remote sensing can detect the electromagnetic wave radiated or reflected from atmosphere, land or ocean. However, due to limitation in budget, we focus on the usage of medium resolution data (Landsat-7 ETM) which are available free of charge through the Internet. Landsat-7 ETM+ images were used for the analysis in the present research. Especially, those images were used for the extraction of drought indicator. Three parameters were selected, i.e. albedo, Normalized Difference Vegetation Index (NDVI), and land surface temperature (Ts). In the following subsection, methods to derive parameters are described.
2.1.1 Albedo (α)

Land surface albedo is a physical parameter that describes the optical reflectance of the land surface. Albedo is commonly defined as the reflectance of a surface integrated with respect to both wavelength (usually between 0.3 µm and 3.0 µm) and angle (i.e. for all directions within the hemisphere above the surface). Examples of the albedo applications include global and regional climatic models for computing the surface energy balance. Three types of albedos, i.e. total-shortwave, total-visible and total-near-infrared albedos, are available. The procedure to estimate albedo from digital number (DN) value is described as follows:

Firstly, according to USGS (2002) DN value was converted into radiance by applying equation (1). A digital number or DN is the value stored within a pixel or cell of an image. Typically, the DN of the pixel represents the amount of light reflected back to the satellite/sensor. However, this is dependent upon the type of data stored in the image. Digital data acquired from satellites are provided to the user in the form of quantified and calibrated values (QCal) for individual picture elements (pixels). These post-calibration QCal values are in units of digital numbers, which have a full range of 8 bits.

\[
L = \frac{(L_{\text{max}} - L_{\text{min}})}{(Q\text{Cal}_{\text{max}} - Q\text{Cal}_{\text{min}})}(Q\text{Cal} - Q\text{Cal}_{\text{min}}) + L_{\text{min}} \tag{1}
\]

where \(L_{\text{min}}\) and \(L_{\text{max}}\) are the spectral radiances for each band at digital numbers 0 and 255, QCal is calibrated and quantified scaled radiance values in digital numbers 0-255, QCal_{min} and QCal_{max} are the minimum and maximum quantized calibrated pixel value. Values for \(L_{\text{max}}\) and \(L_{\text{min}}\) vary for each of the Landsat satellites and for different period of their use. Often the \(L_{\text{max}}\) and \(L_{\text{min}}\) and \(L\) values
published for a given sensor are expressed in units of watts per square meter per steradian per micrometer (W m\(^{-2}\) ster\(^{-1}\) µm\(^{-1}\)). Table 1 shows \(L_{\text{min}}\) and \(L_{\text{max}}\) for all bands of Landsat 7 ETM+, or \(L_{\text{min}}, L_{\text{max}}, QC\alpha_{\text{min}}\) and \(QC\alpha_{\text{max}}\) available in the metadata of data.

Table 1. Landsat-7 ETM+ postcalibration dynamic ranges. Unit of \(L_{\text{min}}\) and \(L_{\text{max}}\) are W m\(^{-2}\)ster\(^{-1}\)µm\(^{-1}\)

<table>
<thead>
<tr>
<th>Band</th>
<th>(L_{\text{min}})</th>
<th>(L_{\text{max}})</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-6.2</td>
<td>191.6</td>
</tr>
<tr>
<td>2</td>
<td>-6.4</td>
<td>196.5</td>
</tr>
<tr>
<td>3</td>
<td>-5.0</td>
<td>152.9</td>
</tr>
<tr>
<td>4</td>
<td>-5.1</td>
<td>241.1</td>
</tr>
<tr>
<td>5</td>
<td>-1.0</td>
<td>31.06</td>
</tr>
<tr>
<td>6</td>
<td>0.0</td>
<td>12.65</td>
</tr>
<tr>
<td>7</td>
<td>-0.35</td>
<td>10.80</td>
</tr>
<tr>
<td>8</td>
<td>-4.7</td>
<td>158.3</td>
</tr>
</tbody>
</table>

Source: USGS (2002)

Radiance was converted into reflectance. According to USGS (2002) the reflectance can be calculated from Landsat-7 data using equation as follows:

\[
\rho_i = \frac{\pi L_{\lambda_i} d^2}{ESUN_{\lambda} \cos \theta} \tag{2}
\]

where,

- \(\rho_i\) = Effective at satellite planetary reflectance composed the combined surface and atmospheric reflectance of the earth (unitless).
- \(L_{\lambda_i}\) = Spectral radiance at the sensor's aperture (W m\(^{-2}\)ster\(^{-1}\)µm\(^{-1}\))
- \(d\) = Earth-Sun distance in astronomical units from nautical handbook (Table 2)
- \(ESUN_{\lambda}\) = Mean solar exoatmospheric irradiances (Table 3)
- \(\theta\) = Solar zenith angle in degrees
Table 2. Earth-sun distance in astronomical units

<table>
<thead>
<tr>
<th>Julian Day</th>
<th>Distance</th>
<th>Julian Day</th>
<th>Distance</th>
<th>Julian Day</th>
<th>Distance</th>
<th>Julian Day</th>
<th>Distance</th>
<th>Julian Day</th>
<th>Distance</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>.9832</td>
<td>74</td>
<td>.9945</td>
<td>152</td>
<td>1.0140</td>
<td>227</td>
<td>1.0128</td>
<td>305</td>
<td>.9925</td>
</tr>
<tr>
<td>15</td>
<td>.9836</td>
<td>91</td>
<td>.9993</td>
<td>166</td>
<td>1.0158</td>
<td>242</td>
<td>1.0092</td>
<td>319</td>
<td>.9892</td>
</tr>
<tr>
<td>32</td>
<td>.9853</td>
<td>106</td>
<td>1.0033</td>
<td>182</td>
<td>1.0167</td>
<td>258</td>
<td>1.0057</td>
<td>335</td>
<td>.9860</td>
</tr>
<tr>
<td>46</td>
<td>.9878</td>
<td>121</td>
<td>1.0076</td>
<td>196</td>
<td>1.0165</td>
<td>274</td>
<td>1.0011</td>
<td>349</td>
<td>.9843</td>
</tr>
<tr>
<td>60</td>
<td>.9909</td>
<td>135</td>
<td>1.0109</td>
<td>213</td>
<td>1.0149</td>
<td>288</td>
<td>.9972</td>
<td>365</td>
<td>.9833</td>
</tr>
</tbody>
</table>

Source: USGS (2002)

The astronomical unit, the AU, is a unit of distance equal approximately to the average distance between the earth and sun. More precisely stated, one astronomical unit is approximately the value of the semi major axis of the orbit of the earth. (For the purists, the AU is actually a tiny bit less than the semi major axis.) This represents a distance of about 93 million miles or 150 million kilometers.

Table 3. Mean solar exoatmospheric irradiances

<table>
<thead>
<tr>
<th>Band</th>
<th>Wm⁻² μm⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1969.000</td>
</tr>
<tr>
<td>2</td>
<td>1840.000</td>
</tr>
<tr>
<td>3</td>
<td>1551.000</td>
</tr>
<tr>
<td>4</td>
<td>1044.000</td>
</tr>
<tr>
<td>5</td>
<td>225.700</td>
</tr>
<tr>
<td>6</td>
<td>82.07</td>
</tr>
<tr>
<td>7</td>
<td>1368.000</td>
</tr>
</tbody>
</table>

Source: USGS (2002)

Finally, albedo was estimated for Landsat-7 ETM+ using equation as follows:

\[
\rho_{BS} = 0.443 \rho_1 + 0.317 \rho_2 + 0.240 \rho_3 \tag{3}
\]

\[
\rho_{NIR} = 0.693 \rho_4 + 0.212 \rho_5 + 0.116 \rho_7 - 0.003 \tag{4}
\]

\[
\rho_{SW} = 0.356 \rho_1 + 0.130 \rho_3 + 0.373 \rho_4 + 0.085 \rho_5 + 0.072 \rho_7 - 0.0018 \tag{5}
\]
where $\alpha_{\text{vis}}$, $\alpha_{\text{nir}}$ and $\alpha_{\text{sw}}$ denote total visible (0.4-0.7 µm), near-infrared (0.7-2.5 µm) and shortwave albedo (0.25-2.5 µm) for ETM+, respectively. $\rho_n$ denotes reflectance of band $n$ (band 1-5 and 7).

### 2.1.2 Surface Temperature (Ts)

Thermal band data (band 6) from Landsat ETM+ can also be converted from spectral radiance to effective at-satellite temperature. According to USGS (2002) surface temperature can be calculated from Landsat data divided two steps. First step calculation brightness temperature with equation:

$$T(K) = \frac{K_2}{\ln\left(\frac{K_1}{L_\lambda} + 1\right)}$$

Where,

- $T(K)$ = Effective at satellite temperature in Kelvin
- $K_1$ = calibration constant 1 in Wm$^{-2}$ster$^{-1}$µm$^{-1}$
- $K_2$ = calibration constant 2 in Kelvin
- $L_\lambda$ = Spectral Radiance in Wm$^{-2}$ster$^{-1}$µm$^{-1}$
- DN=Digital number of each channel (0-255)

Table 4 shows calibration constants $K_1$ and $K_2$ for Landsat TM and ETM+.

<table>
<thead>
<tr>
<th>Units</th>
<th>Wm$^{-2}$ster$^{-1}$µm$^{-1}$</th>
<th>Kelvin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Constant</td>
<td>$K_1$</td>
<td>$K_2$</td>
</tr>
<tr>
<td>Landsat 5 TM</td>
<td>607.76</td>
<td>1260.56</td>
</tr>
<tr>
<td>Landsat 7 ETM+</td>
<td>666.09</td>
<td>1282.71</td>
</tr>
</tbody>
</table>

Source: USGS (2002)

Second step is to calculate a kinetic temperature. The effective satellite temperature values $T(K)$ are referred to a black body. Therefore, corrections for
spectral emissivity $\varepsilon$ are necessary. The kinetic temperature is the variable needed for subjects like heat transfer, because it is the translational kinetic energy which leads to energy transfer from a hot area (larger kinetic temperature, higher molecular speeds) to a cold area (lower molecular speeds) in direct collisional transfer. Kinetic temperature is derived from emissivity correction with equation follows:

$$K = \varepsilon^{0.25}T_{\text{kinetic}}$$

where $\varepsilon$ is emissivity constant (vegetation 0.95, non vegetation 0.92 and water body 0.98).

The last step is to calculate a land surface temperature with equation below:

$$T_s = T_{\text{kinetic}} - 273.16$$

where $T_s$ is surface temperature ($^\circ$C).

1.3 Normalized Different Vegetation Index (NDVI)

NDVI has been widely recognized useful for the studies of the land biosphere characteristics and dynamics at regional to global scales. NDVI is more sensitive to chlorophyll and less contaminated by atmospheric water vapor. NDVI is obtained through calculation of reflectance’s of the red and near infrared (NIR), expressed as the following equations.

$$NDVI = \frac{\rho_{\text{NIR}} - \rho_{\text{RED}}}{\rho_{\text{NIR}} + \rho_{\text{RED}}}$$

where $\rho_{\text{RED}}$ and $\rho_{\text{NIR}}$ denote reflectance’s at red band and NIR band, respectively.

For Landsat TM/ETM+, band 3 and 4 are used.

Vegetation indices (VI) are commonly used to calculate and map vegetation characteristics. The NDVI has been the most widely used method to explore spatial and temporal variation in vegetation properties. The index value
has a range from -1.0 to 1.0, in which higher index values are associated with higher level of healthy vegetation.

2.1.4 Energy Balance

Solar radiation is the largest energy source and is able to change large quantities of liquid water into water vapor. The potential amount of radiation that can reach the evaporating surface is determined by its location and time of the year. Due to differences in the position of the sun, the potential radiation differs at various latitudes and in different seasons. The actual solar radiation reaching the evaporating surface depends on the turbidity of the atmosphere and the presence of clouds which reflect and absorb major parts of the radiation. When assessing the effect of solar radiation on evapotranspiration, one should also bear in mind that not all available energy is used to vaporize water. Part of the solar energy is used to heat up the atmosphere and the soil profile.

A simple modeling methodology rooted in climatology – called ‘energy balance’ modeling – is available to study the role of land cover energy consumption rates. Energy balance refers to the physical fact that energy cannot be created nor destroyed so that the solar and longwave radiation energy received by a land cover layer during any time interval must exactly equal, or ‘balance,’ the energy gained by that layer minus that is lost from the layer during the same time interval. The physical equations that describe these gains and losses are widely used in climate studies.

The radiation coming from the sun can be split into longwave and shortwave. The longwave radiation does heat particular ground features that will eventually be released after a certain amount of time. The shortwave radiation is
instantly reflected by ground features according to their albedo characteristics.

The components of the net radiation (Rn) have been derived as follows (W/m²):

\[ R_n = R_{s_{in}} + R_{l_{in}} - R_{s_{out}} - R_{l_{out}} \]

The equation above is the energy budget concept of the surface at noon. This equation explained that the net radiation is the accumulation from incoming shortwave radiation (Rs_{in}) plus incoming longwave radiation (Rl_{in}) minus outgoing shortwave radiation (Rs_{out}) and outgoing longwave radiation (Rl_{out}).

This rather simple system of radiation balance is considering the ground elements as a layer of a given height, responding uniformly to a radiation stimulus. This concept has two direct advantages, the first one is to simplify layer structural ground element interactions, and the second one is that it is very well fitting the ideal description of a satellite remote sensing and its ground sampling unit: the pixel.

The net radiation is the difference between incoming and outgoing radiation of both short and longwave lengths. It is the balance between the energy absorbed, reflected and emitted by the earth's surface or the difference between the incoming net shortwave and the net outgoing longwave radiation. Rn is normally positive during the daytime and negative during the nighttime. The total daily value for Rn is almost always positive over a period of 24 hours, except in extreme conditions at high latitudes.

The Energy Balance partitioning is summarized at an instant time \( t \) (at the time of satellite overpass) by the following equation:

\[ R_n = G + H + \lambda E \]
where $R_n$ is the net radiation emitted from the Earth surface ($W/m^2$), $G$ is the soil heat flux ($W/m^2$), $H$ is the sensible heat flux ($W/m^2$), $\lambda E$ is the latent heat flux, being the energy necessary to vaporize water ($W/m^2$).

The equation to calculate the net radiation is given by the following equation:

$$R_n = \epsilon \sigma (T_s + 273.16)^4 - (1 - \alpha) R_s + R_l$$

Where $\alpha$ is the albedo, $R_s$ is the downward solar radiation, $R_l$ is the downward longwave radiation, $\epsilon$ is the emissivity of the surface, $\sigma$ is the Stefan-Boltzmann constant, and $T_s$ is the surface temperature.

If the solar radiation, $R_s$, is not measured, it can be calculated with the Angstrom formula which relates solar radiation to extra terrestrial radiation and relative sunshine duration:

$$R_s = Ra \left( \frac{as + bs \frac{n}{N}}{as + bs} \right)$$

Depending on atmospheric conditions (humidity, dust) and solar declination (latitude and month), the Angstrom values as and bs will vary. Where no actual solar radiation data are available and no calibration has been carried out for improved as and bs parameters, the values $a_s = 0.25$ and $b_s = 0.50$ are recommended. The extraterrestrial radiation, $Ra$, and $N$ is the daylight hours or maximum possible duration of sunshine. The actual duration of sunshine, $n$, is recorded with a Campbell Stokes sunshine recorder. $Ra$ is extraterrestrial radiation constitute function from altitude, time angle, zenith angle and sun declination angle depending on date.

The calculation of the clear-sky radiation, $R_s$, when $n = N$, for as and bs the data is not available, is required for computing net shortwave radiation.
The rate of longwave energy emission is proportional to the absolute temperature of the surface raised to the fourth power. This relation is expressed quantitatively by the Stefan-Boltzmann law. The net energy flux leaving the earth's surface is, however, less than that emitted and given by the Stefan-Boltzmann law due to the absorption and downward radiation from the sky.

\[
R_s = (0.75 + 2 \times 10^{-5} z) R_a \quad \text{(14)}
\]

where \( z \) is station elevation above sea level.

\[
\varepsilon \sigma (T_a + 273.16)^4 \times 0.7(1 + 0.17N^2) \quad \text{(15)}
\]

where \( T_a \) is air temperature from climate station and \( N \) is percentage of cloud for satellite data.

Generally lateral fluxes are not considered when dealing with remote sensing images because of their spatial cover capturing the instantaneous energy balance system. Even when transforming the energy balance components for a daily extrapolation of the values, lateral exchanges between pixels are found either in one pixel or in the neighboring ones, extrapolation does not expose lateral values yet encompasses them. The instantaneous soil heat flux (\( G \)) is approximated by fraction on the net radiation, as a function of the NDVI (Normalized Difference Vegetation Index). The soil heat flux is the energy that is utilized in heating the soil. \( G \) is positive when the soil is warming and negative when the soil is cooling. Although the soil heat flux is small compared to \( R_n \) and may often be ignored, the amount of energy gained or lost by the soil in this process should theoretically be subtracted or added to \( R_n \) when estimating evapotranspiration.

Based on (Khomarudin, et. al., 2005) the calculation of \( G \) is as follows:
where,

\[
\frac{G}{Rn} = Ts(0.0038 + 0.007\alpha)(1 - 0.98\text{NDVI}^4) \quad \text{.................................}(16)
\]

NDVI = Normalized Different Vegetation Index (determined from Landsat)

Heat flow into the soil, is driven by a thermal gradient in the uppermost topsoil. It is a conduction flux through the soil matrix. This gradient varies with the state of the vegetation covering the soil that is influencing the light interception by the soil surface. The radiative heating of the topmost layer is then directly modifying the surface temperature and thermal gradient in the top layer.

The sensible heat flux \((H)\) is a convection flux through the atmosphere layers, coming from the surface skin boundary layer with the topmost soil/vegetation layer. The sensible heat flux has been estimated from the difference between radiometric surface temperature \(\text{(Ts)}\) and surface-measured air temperature \(\text{(Ta)}\), and the formulation of a bulk aerodynamic resistance \((W/m^2)\).

The difference between the known radiometric surface temperature and the unknown aerodynamic surface temperature, which actually should be applied, is referred to the following formulation of aerodynamic resistance.

\[
H = \gamma \frac{900}{Ta + 273} \times U_2(Ts - Ta) \quad \text{.................................}(17)
\]

where,

- \(Ts\) = Surface temperature \((^\circ \text{C})\) (Landsat)
- \(Ta\) = Air temperature \((^\circ \text{C})\)
- \(\gamma\) = Psychometrics constants \((\text{kPa} \cdot ^\circ \text{C}^{-1})\)
- \(U_2\) = Wave length of radiation emission \((11.5 \, \text{µm})\)
- \(U_2\) = Wind velocity at 2 m above ground surface \((\text{m s}^{-1})\)

The psychrometric constant, \(\gamma\), is given by:
\[ \gamma = \frac{c_p P}{\varepsilon \lambda} = 0.665 \times 10^{-3} P \text{ (FAO, 1998)} \] 

(18)

where,

\( P = \text{Atmospheric pressure (kPa)} \)

\( \lambda = \text{Latent heat of vaporization (2.45 MJ kg}^{-1} \) \)

\( c_p = \text{Specific heat at constant pressure, 1.013 \times 10^{-3} (MJ kg}^{-1} °C^{-1} \) \)

\( \varepsilon = \text{Ratio molecular weight of water vapour/dry air = 0.622} \)

The specific heat at constant pressure is the amount of energy required to increase the temperature of a unit mass of air by one degree at constant pressure. Its value depends on the composition of the air, i.e., on its humidity. For average atmospheric conditions a value \( c_p = 1.013 \times 10^{-3} \text{ MJ kg}^{-1} °C^{-1} \) can be used. As an average atmospheric pressure is used for each location (Equation 19), the psychrometric constant is kept constant for each location.

\[ zP = 101.3 \left( \frac{293 - 0.006z}{293} \right)^{5.26} \text{ (FAO, 1998)} \] 

(19)

(FAO, 1998)...

where,

\( z = \text{Elevation above sea level (m)} \)

Wind speeds measured at different heights above the soil surface are different. Surface friction tends to slow down wind passing over it. Wind speed is slowest at the surface and increases with height. For this reason anemometers are placed at a chosen standard height, i.e., 10 m in meteorology and 2 or 3 m in agrometeorology. For the calculation of evapotranspiration, wind speed measured at 2 m above the surface is required. To adjust wind speed data obtained from instruments placed at elevations other than the standard height of 2 m, a
logarithmic wind speed profile may be used for measurements above a short grassed surface:

\[ U_2 = U_z \frac{4.87}{\ln(67.8z - 5.42)} \]  

(FAO, 1998)  

(20)

where,

- \( U_2 \) = Wind velocity at 2 m above ground surface (m s\(^{-1}\))
- \( U_z \) = Wind velocity from measurements at above ground surface (m s\(^{-1}\))
- \( z \) = Height of measurement above ground surface (m)

The energy necessary to vaporize water under given atmospheric conditions is especially ruled by the resistance to vaporization parameter. The latent heat flux (\( \lambda E \)) is energy that used to evaporate water. The evapotranspiration process is determined by the amount of energy available to vaporize water.

\[ \lambda E = (R_n - G - H) \]  

(21)

where,

- \( \lambda E \) = Energy for Evapotranspiration (W/m\(^2\))
- \( R_n \) = Net Radiation (W/m\(^2\))
- \( H \) = Energy for Sensible Heat Flux (W/m\(^2\))
- \( G \) = Energy for Soil Heat Flux (W/m\(^2\))

Finally, the evaporative fraction (EF) is expressed as:

\[ EF = \frac{\lambda E}{R_n - G} \]  

(22)

EF indicates how much of the available energy is used for evapotranspiration, that for transpiration of the vegetation and evaporation of the soil and EF will be close to one (no water stress). As long as moisture is available, energy will be used for its evaporation. With little or no moisture left, all available energy will
be directed into the sensible heat flux and EF will approach zero (serious water stress).

The above mentioned methods, however, require quite complex models to construct the case-specific algorithms which make direct use of remote measurements of spectral radiances. Presently a lot of effort is concentrated into increasing the accuracy of radiant fluxes, even if surface albedo can easily be estimated by common sensors (enabling the calculation of the shortwave net radiation), it takes more specific sensors to estimate the longwave component of the radiation balance. Surface albedo and temperature can also be the basis for estimates of the upwelling components, while the downwelling components are based on meteorological data. Soil heat flux can be estimated by the ratio G/Rn through spectral indices or by semi empirical equation including Rn, the surface albedo, the surface temperature, NDVI and the area average surface albedo. Even if the net radiation and soil heat flux parts of the energy balance equation are relatively well known and estimated from a remote sensing point of view, the remote sensing of the sensible heat flux especially its most critical parameter is still limited.

2.2 Application of Remote Sensing and GIS for Drought Prediction

Remote sensing is the science (and to some extent, art) of acquiring information about the earth's surface without actually being in contact with it. This is done by sensing and recording reflected or emitted energy and processing, analyzing, and applying that information. In much of remote sensing, the process involves an interaction between incident radiation and the targets of interest. This exemplified by the use of imaging systems where the following seven elements
are involved. Note, however, that remote sensing also involves the sensing of emitted energy and the use of non-imaging sensors.

Computer based systems are used to store and manipulate geographic information (Aronoff, 1991). A computer system is capable of holding and using data describing places on the earth’s surface. Any of various software applications, running on PCs or workstations, that store, analyzes, and displays multiple layers of geographic information (Lang, 1998).

Many types of information that are needed in natural drought management are important such as map, satellite imagery, GPS data, climate data, etc. Many of these data have different projection and coordinate system, and need to be brought to a common map basis in order to superimpose them (Westen, 2002). Remote sensing and GIS provide a historical database from which drought map may be generated, indicating which areas are potentially dangerous.

When drought occurs, the speed of information collection from air and space borne platforms and the possibility of information dissemination with a corresponding swiftness make it possible to monitor the occurrence of the drought. Simultaneously, GIS may be used to integrate satellite data with other relevant data (such as climate data), useful in combination with Global Positioning System (GPS) and classification of the spatial analysis functionality (selection, manipulation, exploration, and confirmation). The interaction between the various functions is schematically summarized in Figure 2 (Anselin, 1998).
2.3 Web Publication

The greatest problem for decision makers, resource managers and user who handle information is accessing the right information at the real time. Internet is a global, public collection of individual networks that is operated by private organizations, universities, and government agencies (Maududie, et al., 2002). GIS takes advantages of the internet as a media of GIS dissemination, so that the GIS data can be accessed by different computers (or servers) from anywhere across the internet. To be able to access and share remote GIS data, the system requires high interoperability (Peng, et al., 2003). The internet and web-based GIS is an effective medium for publication of spatial data. Web-based GIS is not a single technology, there are many combined components to build web based GIS, such as software and hardware. These technologies are related to web based GIS including Object-Oriented Language, GIS package and language,
HTML and web scripting, and the theories about GIS. It is important to define the different forms of web-based GIS, before selecting a specific location for any given application.

One type of web-based GIS applications is Map Generators. Map generators use a web-based browser form. The user enters specifications of a drought event such as location, thematic layers and symbols on the form. The form is passed to the web server. A gateway at the web server passes the request to a GIS server. For instance, the gateway could pass the request in the form of AML to an Arc/INFO server. The Arc/INFO server generates a graphic file, which is converted to a GIF image. The GIF image is sent back to the client and viewed using native browser capability. The advantage of map generators is creating custom maps on the fly. Disadvantages include lack of access to the raw spatial data, typically at a slower speed, limited predefined user choices, and involved setup.