This chapter is devoted to important aspects of the remote sensing technology utilised in the study.

The literature survey presented in this chapter gives the state-of-art of the studies carried out by various researchers to investigate various land surface characteristics.
2.0 INTRODUCTION

Remote sensing is the acquisition of data i.e. deriving information about objects or materials (targets) located at the earth’s surface or in its atmosphere using sensors mounted on platforms located at a distance from the targets. Here measurements are made in different spectral regions on interactions between the targets and electromagnetic radiation (EMR). The field of remote sensing encompasses techniques that obtain precise information about earth’s surface from a distance (Gupta et al. 2003).

Remote sensing, a technique for making measurements on the object at a distance, to a great extent relies on the interaction of EMR with matter. Different stages in remote sensing can be broadly enlisted as (1) a source of electromagnetic energy (Sun/self-emission), (2) transmission of energy from a source (Sun) to the surface of the earth wherein it also undergoes absorption and scattering during passage through atmosphere, (3) interaction of EMR with earth’s surface (reflection, scattering, absorption and re-emission), (4) transmission of reflected / scattered / emitted energy from the object /features of the earth’s surface to the remote sensors (with due modifications due to atmospheric effects), and sensors data output.

Macroscopically, the interactions are absorption, transmission and reflection of radiation from the features of the earth’s surface. Microscopically, these are due to atomic and molecular absorption and scattering. Earth emits energy with peak emission in Thermal infrared (8-12 μm) region and this is used for estimating temperatures of various features on the surface of the earth.

2.1 ELECTROMAGNETIC SPECTRUM

Information from an object to the sensor is carried by the electromagnetic energy and could be encoded in the form of frequency, intensity or polarization of the electromagnetic wave. The information is propagated by electromagnetic waves with velocity of light from the object
directly through free space as well as indirectly by reflection, scattering and re-radiation by aerosols to the sensors. The interaction of electromagnetic waves with natural surfaces and atmosphere is strongly dependent on the frequency of the waves. Electromagnetic spectrum is divided into a number of spectral regions as illustrated in Figure 2.1.

Figure 2.1: Electromagnetic spectrum (Canada Centre for Remote Sensing, Natural Resources Canada).

The radio band covers the region of wavelengths that is longer than 100 cm (frequency less than 300 MHz). Microwave band covers the neighbouring region, having wavelength up to 1 mm (300 GHz frequency). The sensors operating in this region are RADAR, Microwave Radiometer, Altimeter, Scatterometer etc. The spectral region from 0.7 μm to 1 mm is
subdivided into near; middle; thermal; far infrared and sub millimetre sub regions. The sensors operating in this region of EMR are imaging spectrometers; radiometers, polarimeters and laser based active sensing system.

Table: 2.1 RADAR wavelength and frequencies used in Microwave Remote sensing (source: Jensen, 2003).

<table>
<thead>
<tr>
<th>RADAR</th>
<th>Wave length (λ) in cm</th>
<th>Frequency (ν) in GHz</th>
</tr>
</thead>
<tbody>
<tr>
<td>K_a (0.86cm)</td>
<td>0.75-1.18</td>
<td>40.0-26.5</td>
</tr>
<tr>
<td>K</td>
<td>1.19-1.67</td>
<td>26.5-18.0</td>
</tr>
<tr>
<td>Ku</td>
<td>1.67-2.4</td>
<td>18.0-12.5</td>
</tr>
<tr>
<td>X (3.0 and 3.2 cm)</td>
<td>2.4-3.8</td>
<td>12.5-8.0</td>
</tr>
<tr>
<td>C (7.5, 6.0 cm)</td>
<td>3.9-7.5</td>
<td>8.0-4.0</td>
</tr>
<tr>
<td>S (8.0, 9.6, 12.6 cm)</td>
<td>7.5-15.0</td>
<td>4.0-2.0</td>
</tr>
<tr>
<td>L (23.5, 24.0, 25.0cm)</td>
<td>15.0-30.0</td>
<td>2.0-1.0</td>
</tr>
<tr>
<td>P (68.0cm)</td>
<td>30.0-100</td>
<td>1.0-0.3</td>
</tr>
</tbody>
</table>

Visible light (0.4 - 0.7 μm) is only one of many form of electromagnetic energy. Radio waves, thermal, ultraviolet and X-rays are other familiar forms of energy that propagate in accordance with basic wave theory. Table 2.1 shows RADAR wavelength and frequencies used in Microwave Remote sensing Investigations.

Wave theory describes electromagnetic energy travelling in a harmonic sinusoidal fashion at the velocity of light c which is $3\times10^8$ m/s. The distance
between two consecutive points having phase difference of $2\pi$-radian is called wavelength $\lambda$ and the number of peaks passing through a fixed pointing space per unit time is the wave frequency. All the waves obey the general equation

$$c = f\lambda.$$  \hspace{1cm} (2.1)

Particle theory offers useful insight into how much electromagnetic energy interacts with matter during transfer of energy and momentum.

The particle theory suggests that EM energy is composed of many discrete units called the photons or Quanta. The energy of quantum is given by

$$E = hf.$$  \hspace{1cm} (2.2)

Where $E$ is the energy in joule;

$h$ is a Plank's constant ($6.626 \times 10^{-34}$J$s$);

$f$ is the frequency.

Where $f = c/\lambda$, hence,

$$E = hc/\lambda.$$  \hspace{1cm} (2.3)

Thus energy of quanta is inversely proportional to its wavelength. The longer the wavelength involved, the lower is its energy content. This has important implication in remote sensing.

The naturally emitted longer wavelength radiation, such as microwave emission from terrain features is more difficult to sense than radiation from shorter wavelength such as thermal infrared radiation. The low energy content of longer wavelength radiation means that, a system operating at longer wavelength must collect energy from large area of the earth at any given time in order to obtain detectable energy signal so that it maintains satisfactory signal to noise ratio.
2.2 ATMOSPHERIC EFFECTS IN REMOTE SENSING

As sunlight initially enters the atmosphere, it encounters gas molecules, suspended dust particles, and aerosols. These materials tend to scatter a portion of the incoming radiation in all directions, with shorter wavelengths experiencing the strongest effect. Although most of the remaining light is transmitted to the surface, some atmospheric gases are very effective at absorbing particular wavelengths. As a result of these effects, the illumination reaching the surface is a combination of highly filtered solar radiation transmitted directly to the ground and more diffused light, scattered from all parts of the sky, which helps illuminate shadowed areas (Figure 2.2).

[Image: Interaction of EMR with Earth's objects. (Source: Oza, 2006)]

In remote sensing technique the electromagnetic radiation emitted or reflected from the objects of interest has to pass through atmosphere before the receiver of remote sensor detects it. Remote sensing by satellites (at an altitude above 700 KM) involves atmospheric degradation from the entire atmospheric column. Thus, the characteristics of the atmosphere significantly determine the effective use of electromagnetic spectrum for remote sensing.

The absorption and re-emission process in the atmosphere are through changes in electronic, vibrational and rotational quantization levels. The
spectral line absorption in visible and near-IR regions are due to changing electronic quantization levels of atoms as it is in the case of absorption of ultraviolet light by ozone. The transition of electron in an atom or molecule can occupy certain energy levels, but not all possible energy levels. In certain wavelengths of incident radiation, the photon energy is just sufficient to cause a transition from one permissible energy level to another, thereby getting selectively absorbed by molecules of certain gases. Vibrational energy of the molecules is related with the to and fro movement of atoms. The kinetic energy that causes rotation of molecule as a whole is called rotational energy.

The most important atmospheric constituents that influence the incident radiation are water vapour (H₂O), oxygen (O₂), Ozone (O₃), carbon dioxide (CO₂) and aerosols. The following spectral regions are atmospheric windows with very little attenuation that could be used for remote sensing.

The wavelengths shorter than 0.3 μm are completely absorbed (with exception of 1216 Å by Ozone layer in upper atmosphere. The atmospheric attenuation at selective wavelengths in the reflected IR (Near and Mid IR) is mainly due to water vapour present in the atmosphere.

The CO₂ absorption bands around 4.3 and 15 μm have been in use to estimate temperature of atmosphere up to 50 km altitude while 6.7 μm absorption band has been in use to estimate water vapour distribution up to 10 km altitude. Figure 2.3 depicts the generalized absorption spectrum of the earth’s atmosphere. The attenuation of incoming solar and outgoing earth’s radiation is the consequence of the atmospheric absorption and scattering.

Scattering of electromagnetic radiation within the atmosphere reduces the image contrast and changes the spectral signature of ground objects as seen by the sensor. The scattering of electromagnetic radiation depends on the relative size of the gas molecule or particle with reference to the interacting wavelength. The gas molecules are of the order of 10⁻⁴ μm size. The size of haze particles, which are water droplets formed by condensation
around particles of soluble substances, varies from $10^{-2}$ to $10^2 \mu m$ depending upon the relative humidity.

The atmospheric scattering can be categorized as Rayleigh / Mie or Non-selective scattering. The atmospheric scattering can be described in terms of Rayleigh scattering coefficient $\beta$ when the interacting wavelength is much larger than the particle size, Mie scattering, when the particle size is comparable to the interacting wavelength. Non-selective scattering, which is independent of wavelength, occurs when the particle size is very much larger than the interacting wavelength. The particles such as in cloud, fog and dust having dimension ranging from 2 to 20 $\mu m$ are responsible for non-selective scattering and these specially affect remote sensing in infrared region. The process of scattering causes reduction in image contrast.

2.3 INTERACTION OF EMR WITH MATTER

As this modified solar radiation reaches the ground, it may encounter soil, rock surfaces, vegetation, or other materials that absorb a portion of the radiation. The amount of energy absorbed varies in wavelength for each material in a characteristic way, creating a sort of spectral signature. Most of the radiation not absorbed is reflected (scattered) back up into the atmosphere. A part of this radiation is reflected back in the direction of the satellite. This upwelling radiation undergoes a further scattering and absorption as it passes through the atmosphere before finally being detected and measured by the sensor. If the sensor is capable of detecting thermal infrared radiation, it will also pick up radiation emitted by surface objects as a result of solar heating.

As discussed above, various mechanisms play a significant role when EMR interacts with matter. These are mainly depending on the frequency of the wave (its photon energy) and the energy level structure of the matter. During this interaction, energy is being exchanged between EMR and matter.
The main interaction mechanisms playing a role in wave-matter across the EM spectrum are summarized in table 2.2.

Table: 2.2 Interaction mechanisms for wave-matter across the EM spectrum.

(Source: Elachi, 1987)

<table>
<thead>
<tr>
<th>Spectral Region</th>
<th>Main Interaction Process</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gamma &amp; X-rays</td>
<td>Atomic processes</td>
</tr>
<tr>
<td>Ultraviolet</td>
<td>Electronic processes</td>
</tr>
<tr>
<td>Visible and Near Infrared</td>
<td>Electronic and vibrational molecular processes</td>
</tr>
<tr>
<td>Mid-IR</td>
<td>Vibrational, vibrational-rotational molecular processes</td>
</tr>
<tr>
<td>Thermal-IR</td>
<td>Thermal emission, Vibrational and rotational processes</td>
</tr>
<tr>
<td>Microwave</td>
<td>Rotational processes, thermal emission, scattering, conduction</td>
</tr>
<tr>
<td>Radio-frequency</td>
<td>Scattering, conduction, ionosphere effect</td>
</tr>
</tbody>
</table>

2.4 ATMOSPHERIC WINDOWS

Scattering and absorption of EM radiation by the atmosphere have significant effects. It has an impact on sensor design as well as the processing and interpretation of images. When the concentration of scattering agents is high, the scattering reduces contrast of the visual effect that we call, haze. Haze increases the overall brightness of the scene and reduces the contrast between different ground materials. A hazy atmosphere scatters some light upwards, so a portion of the radiation recorded by a remote sensor, called
path radiance, is the result of this scattering process. Since the amount of scattering varies with wavelength in general (barring by large aerosols), so does the contribution of path radiance to remotely sensed images. The path radiance effect is greatest for the shortest wavelengths, falling off rapidly with increasing wavelength. When images are captured over several wavelength ranges, the differential path radiance effect complicates comparison of brightness values at the different wavelengths. The atmospheric components that are effective absorbers of solar radiation are water vapour and ozone. Each of these gases tends to absorb energy in specific wavelength range. Some wavelengths are almost completely absorbed. Consequently, most broadband remote sensors have been designed to detect radiation in the “atmospheric windows”, those wavelength ranges for which absorption is minimal, and, conversely, transmission is high (Figure 2.3).

![Figure 2.3: Atmospheric Windows. Atmospheric windows can be found from higher transmittance (Source: Oza, 2006).](image)

### 2.5 OPTICAL SENSOR PROPERTIES

The information collected by the remote sensing sensor is meant to identify and map various earth surface objects. Therefore, the performance of the sensor can be evaluated based on its classification as well as its mapping accuracy requirements. This will depend on the instrument’s ability to detect small differences in the emittance/reflectance of the earth’s surface in an optimum number of spectral bands and recognition of smallest object. However, these sensor specifications are application dependent (Joseph, 2003).
In OIR region, sensor parameters can be divided in four domains (i) spatial, (ii) spectral, (iii) Radiometric and (IV) Temporal. Temporal resolution refers to the temporal frequency with which a given scene can be imaged, usually expressed in days. Temporal resolution of 24 days means that the sub-satellite track repeats every 24 days. Hence any part of the globe could be imaged every 24 days. Here, besides sensor characteristic (like swath), satellite orbit also plays a decisive role.

EMR when incident on an earth object, gets reflected, absorbed, transmitted or re-radiated depending upon the property of object and wavelength of the incident EMR. This interaction of object with various wavelengths (frequencies) of EMR is different for different materials. This basic property, which allows identification of an object, is called signature.

2.6 SIGNATURE OF VEGETATION IN OPTICAL AND NIR REGION

Various features of earth surface exhibit different signatures while interacting with the different portions of EM spectrum. These sections describe the interaction of vegetation with the OIR region. The main part of vegetation playing significant role in the interaction of reflected OIR region is leaves.

2.6.1 Intrinsic Indices

In remote sensing applications and research, more than 40 indices have been developed during the last three decades (Bannari et al., 1995). Among these indices, the Simple Ratio (SR) vegetation index and the Normalized Difference Vegetation Index (NDVI) have been widely employed for exploiting the spectral signature of vegetation.

SR is the ratio of reflectance of red in visible and NIR bands:

\[ SR = \frac{\rho_{NIR}}{\rho_{Red}} \]  

(2.4)
Here, $\rho_{\text{red}}$ and $\rho_{\text{NIR}}$ are spectral reflectances of red and near-infrared bands.

The NDVI is calculated from these individual measurements as follows:

$$NDVI = \frac{\text{NIR} - \text{VIS}}{\text{NIR} + \text{VIS}}$$  \hspace{1cm} (2.5)

Where, VIS and NIR stand for the spectral reflectance measurements acquired in the visible (red) and near-infrared regions. These spectral reflectances are themselves a ratio of the reflected to the incoming radiation in each spectral band individually; hence they take on values between 0.0 and 1.0. By design, the NDVI itself thus varies between -1.0 and +1.0. It should be noted that NDVI is functionally and not linearly, equivalent to the simple infrared/red ratio (NIR/VIS).

The advantage of NDVI over a simple infrared/red ratio is therefore generally limited to any possible linearity of its functional relationship with vegetation properties (e.g. biomass). The simple ratio (unlike NDVI) is always positive, which may have practical advantages, but it also has a mathematically infinite range (0 to infinity), which can be a practical disadvantage as compared to NDVI. Also in this regard, note that the VIS term in the numerator of NDVI only scales the result, thereby creating negative values. NDVI is functionally and linearly equivalent to the ratio NIR / (NIR+VIS), which ranges from 0 to 1 and is thus never negative or limitless in range (Crippen, 1990). But the most important concept in the understanding of the NDVI algebraic formula is that, despite its name, it is a transformation of a spectral ratio (NIR/VIS), and it has no functional relationship to the spectral difference (NIR-VIS).

In general, if there is much more reflected radiation in near-infrared wavelengths than in visible wavelengths, then the vegetation in that pixel is likely to be dense and may contain some type of forest. Subsequent work has shown that the NDVI is directly related to the photosynthetic capacity and
hence energy absorption of plant canopies (Sellers, 1985; Myneni at al., 1995).

- It can be seen from its mathematical definition that the NDVI of an area containing a dense vegetation canopy will tend towards positive values (say 0.3 to 0.8) while clouds and snow fields will be characterized by negative values of this index.

- Free standing water (e.g., oceans, seas, lakes and rivers) which have a rather low reflectance in both spectral bands (at least away from shores) and thus results in very low positive or even slightly negative NDVI value.

- Soils which generally exhibit a near-infrared spectral reflectance (somewhat larger than red,) and thus tend to generate rather small positive NDVI value (say 0.1 to 0.2).

NDVI Data can be used to estimate a large number of vegetation properties from the value of this index. Typical examples include the Leaf Area Index, biomass, chlorophyll concentration in leaves, plant productivity, fractional vegetation cover, accumulated rainfall, etc. Such relations are often derived by correlating space-derived NDVI values with ground-measured values of these variables, which are shown in Figures 2.4, 2.5 and 2.6.

The drought affected zones of northern Rajasthan, much of Gujarat, and parts of Madhya Pradesh clearly show signs of significant vegetation stress, however these are not the major crop producing areas. The major wheat production areas exhibited a noticeable increase in vegetation health during the critical late-March time period.

This would be consistent with a slightly delayed crop, likely resulting from the cool growing season. It is worth noting that cool temperatures during the early portion of the growing season are typically beneficial to wheat, as it creates an optimal tillering environment which often results in higher yields.
Yield potential is also boosted by reports of an increase in certified seed use and herbicide availability.

Figure 2.4: Comparison of Difference in Vegetation Index for the year 1999 and 2000 (Data Source: SPOT4 Vegetation Index).

2.7 MICROWAVE REMOTE SENSING

2.7.1 Physical Principles of Radar Scatterometry

The Scatterometer on board of the ERS-1/2 and NASA’s QuikSCAT satellites are non-imaging systems that provide a quantitative measure of the backscattering coefficient $\sigma^0$. They are working in the microwave region of the electromagnetic spectrum that extends from about 0.3 to 300 GHz. We humans have no sensory perception of microwaves therefore it is often difficult for us to understand the information content of $\sigma^0$ measurements. It is only through empirical observations and theoretical considerations that one starts to learn what $\sigma^0$ tells us about the observed target.

In this chapter some principles of radar scatterometry are presented. It is shown that the backscattering coefficient is determined by the geometrical
structure and the dielectric properties of the observed target. Then the
dielectric properties of natural media including soil, vegetation, and snow are
discussed. This discussion shows that the dielectric properties are to a large
extent determined by the fraction of water within these media which explains
the very important role of water in microwave remote sensing.

2.7.2 Radar Equation

A radar scatterometer transmits an electromagnetic wave and
measures the energy of the wave that is scattered backwards from the
illuminated object. The received power depends on the technical
characteristics of the radar, the distance between the radar and the object,
and on the properties of the object itself. This is expressed by the (monostatic)
radar equation (Ulaby et al., 1982).

2.7.3 Normalized Radar Cross Section

Suppose a point source emits a microwave pulse of power $P_T$ uniformly
in all directions in space. The power flux $\Phi_T$ at a distance $R$ from the source
will then be

$$\Phi_T = \frac{P_T}{4\pi R^2}. \quad (2.6)$$

The antenna gain is defined as $G = 4\pi /\Omega$, where $\Omega$ is the beam width
in steradian (space angle) in which the emitted power is contained. For an
emission uniform in all directions $G = 1$, whereas for a narrow radar beam $G
\gg 1$, and the power flux becomes

$$\Phi_T = G \frac{P_T}{4\pi R^2}. \quad (2.7)$$

Figure 2.5 depicts radar backscattering.
Now suppose that microwave radiation hits a scatterer. Then the radar cross section is defined as

$$\sigma = \frac{P_s}{\Phi_r}$$  \hspace{1cm} (2.8)

Where, $P_s$ is the backscattered power by the target. The radar cross section depends on the geometry of the target with respect to the incident microwaves and the dielectric constant of the material of the scatterer. If the power detected by the receiving antenna, is identical to the source antenna and of size $A$, then

$$P_R = P_s A / (4\pi R^2) = G P_T \sigma A (4\pi R^2)^{-2}.$$  \hspace{1cm} (2.9)

For a narrow beam rectangular antenna, i.e., $L_X$ by $L_Y$, are the length and breadth and a system without losses, a relationship exists between the gain $G$ and the antenna surface area $A$ since

$$\Omega = \langle \lambda / L_X \rangle . \langle \lambda / L_Y \rangle = \lambda^2 / A.$$  \hspace{1cm} (2.10)

Therefore

$$A = \lambda^2 G / 4\pi.$$  \hspace{1cm} (2.11)

A scatterometer has a footprint, $F$, of several tens of kilometre diameters with generally a large number of scattering elements in it. For such a distributed target one may define a dimensionless microwave cross section...
per unit surface, often denoted by “Normalized Radar Cross Section” or NRCS and by convention written as $\sigma^0$. We can now write:

$$P_R = \frac{\lambda^2}{(4\pi)^3} \int \frac{P_R \sigma^0 G^2}{R^4} dF$$  \hspace{1cm} (2.12)

In order to solve this integral one usually assumes that $\sigma^0$ does not vary over the area of interest, such that:

$$\sigma^0 = \frac{(4\pi)^3 R^4 P_R}{\lambda^2 G^2 FP_I}$$  \hspace{1cm} (2.13)

However, in reality the roughness elements on the ocean surface will largely depend on the local wind condition, which in turn exhibits large variability over a scatterometer footprint, which is shown in Figure 2.6.

The SeaWind scatterometers aboard QuikSCAT and ADEOS II provide normalised radar cross section ($\sigma^0$) measurements of the Earth’s surface at unprecedented coverage and resolution. While originally designed for wind observation, scatterometers have proven in a variety of land and Ice studies. To further improve the utility of the data, resolution enhancement algorithms have been developed and applied to the data. These algorithms produce images of the surface $\sigma^0$ at enhanced resolution to a better than 5 km (Figure 2.7).
Figure 2.6: Schematic representations of microwave scattering and reflection at (a) smooth, (b) rough and (c) very rough surface. As the roughness increases, more microwave power is returned towards the direction of the microwave source (Source: Stoffelen, 1962).

Figure 2.7: SAs (South Asia section) QuikSCAT enhanced resolution land regions (courtesy MERS 05-04, Long, 2005).

As the scattering mechanism does not linearly depend on the geophysical conditions, the geophysical variability within the footprint will contribute to $\sigma^0$. This will be particularly acute at low winds. $\sigma^0$ is generally expressed in dB, i.e. the value of $10 \log (\sigma^0)$. 
2.7.4 Scattering and Absorption of Electromagnetic Waves

In the radar equation it is assumed that an object can scatter an incident wave in all directions. But what are the physical reasons for the scattering phenomenon? Let us consider an electromagnetic wave which is characterized by the electric and magnetic field vectors \( \mathbf{E} \) and \( \mathbf{B} \). According to Lorentz force law an electromagnetic field exerts a force \( \mathbf{F} \) on a charge \( Q \) moving with a velocity \( \mathbf{v} \):

\[
\mathbf{F} = Q \left( \mathbf{E} + \mathbf{v} \times \mathbf{B} \right) \quad (2.14)
\]

When the wave is incident on an object then the positive and negative charges within any material will be accelerated into different directions. The induced microscopic currents will reradiate electromagnetic waves in all directions thereby modifying the original field. The amount of energy scattered backwards into the direction of the incident wave therefore depends on the geometrical and electromagnetic properties of the illuminated object. A mathematical description of this process, which is provided by electromagnetic theory, is presented below. The treatment follows mainly Jackson (1983) and Ishimaru (1978).

2.7.5 Maxwell’s Equations

The basic equations of electromagnetic theory in macroscopic media are Maxwell’s equations:

\[
\nabla \cdot \mathbf{D} = \rho; \quad \nabla \cdot \mathbf{B} = 0; \quad \nabla \times \mathbf{E} = -\frac{\partial \mathbf{B}}{\partial t}; \quad \nabla \times \mathbf{B} = \frac{\partial \mathbf{D}}{\partial t} + \mathbf{J} \quad (2.15)
\]

Where,

\( \mathbf{E} \) = electric field intensity (voltage/distance);

\( \mathbf{H} \) = magnetic field intensity (current/distance);

\( \mathbf{D} \) = electric displacement (charge/area); 

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\( \mathbf{B} \) = magnetic induction (voltage X time)/ (area);

\( \rho \) = volume charge density (charge/volume);

\( \mathbf{J} \) = surface current density (current/area).

For linear, isotropic and homogeneous media the electric field variables \( \mathbf{E} \) and \( \mathbf{D} \) are related via:

\[
\mathbf{D} = \varepsilon \varepsilon_0 \mathbf{E}. \tag{2.16}
\]

Where \( \varepsilon_0 \) is the permittivity of vacuum \( (\varepsilon_0 = 8.854 \times 10^{-12} \text{As/Vm}) \) and \( \varepsilon \) is the relative permittivity of the medium.

The corresponding relation for the magnetic field variables \( \mathbf{B} \) and \( \mathbf{H} \) is:

\[
\mathbf{B} = \mu \mu_0 \mathbf{H}. \tag{2.17}
\]

Where \( \mu \) is the relative permeability and \( \mu_0 \) is the permeability of vacuum \( (\mu_0 = 4\pi \times 10^{-7} \text{Vs/Am}) \). \( \varepsilon \) and \( \mu \) are empirical parameters that describe the interaction of electromagnetic fields and media. All substances show magnetic effects but only for ferromagnetic and ferrimagnetic materials. \( \mu \) takes on values significantly different from one (Duffin, 1965). Water, for instance, has a relative permeability of 0.999988 (Mätzler, 1987). Therefore, in remote sensing, the small magnetic interaction can be neglected and \( \mu \) can be set equal to one.

The relative permittivity \( \varepsilon \) is also called the dielectric constant of a material media and is in general a complex quantity to account for losses in the media:

\[
\varepsilon = \varepsilon' + j\varepsilon'' \tag{2.18}
\]

Amongst natural media water plays a very important role because of its distinct dielectric properties. For example, at frequencies below 2 GHz the dielectric constant of liquid water is around 80, while for other natural media
like soil particles, dry vegetative matter or rocks it rarely exceeds 7 (Ulaby et al., 1986).

2.7.6 Wave Propagation

The propagation of waves in macroscopic media in the absence of a source is described by a wave equation. The wave equation can be directly derived from Maxwell’s equations:

\[
\nabla^2 E = \frac{\varepsilon}{c^2} \frac{\partial^2 E}{\partial t^2} \quad (2.19)
\]

Where, \( c \) is the speed of light in vacuum which is equal to \( (\varepsilon_0 \mu_0)^{-1/2} \). The equation for \( B \) has exactly the same form. One solution of (2.20) is a plane wave which is given by

\[
E = E_0 e^{j(\pm kx - \omega t)}, \quad (2.20)
\]

Where \( E_0 \) is a vector constant in time, \( \omega \) is the angular frequency, \( k \) is the propagation vector, and the propagation of wave is in \( x \) direction. The magnitude of the propagation vector is called the wave number \( k \) and the value of

\[
k = 2\pi/\lambda. \quad (2.21)
\]

The vectors \( E, B, \) and \( k \) form orthogonal vectors set mutually perpendicular to each other such that the vector \( E \times B \) which is also known as Poynting vector that points along the direction of propagation vector \( K \), When (2.18) is substituted into (2.17) it can be seen that:

\[
k = \sqrt{\varepsilon \omega \over c} = n \cdot k_0, \quad (2.22)
\]
Where $k_0 = \omega/c$ is the wave number in vacuum, and $n$ is the refractive index. As already mentioned, to account for losses in the medium, the dielectric constant and the refractive index are complex quantities:

$$n = n' + jn'' = \sqrt{\varepsilon' + \varepsilon''}. \tag{2.23}$$

The amplitude of the wave decreases exponentially which can be readily seen if the direction of propagation of the plane wave is chosen, for example, to be along the $z$-axis:

$$E = E_0 e^{-n'k_0z} e^{j(n'k_0z - \omega t)} \tag{2.24}$$

The distance at which the power of the wave $|E|^2$ decreases by $e^{-1}$ is called the penetration depth $L_p$:

$$L_p = \frac{1}{2n''k_0} \tag{2.25}$$

Another solution of the wave equation (2.16) is a spherical wave which is of the form:

$$E = E_0 \frac{1}{r} e^{j(\pm k r - \omega t)} \tag{2.26}$$

Where, $r$ is the radius of the spherical coordinate system. A spherical wave travels along the radius vector. As is the case for plane waves, $E$, $B$ and the direction of propagation are mutually perpendicular.

### 2.7.7 Scattering of a Wave by a Particle

In this section the scattering of a plane wave by a single particle is described in mathematical terms. The objective is to demonstrate that scattering arises from microscopic currents which are generated by the incoming wave, and that this phenomenon is accounted for by describing the object in terms of its geometrical and dielectric properties. The treatment follows closely the introductory chapter in Ishimaru (1978). Let us consider a particle whose dielectric constant is a function of the position within the body. It occupies a volume $V$ and is surrounded by a medium with a dielectric constant $\varepsilon_0$. It is illuminated by a plane wave of frequency $\omega$. With the understanding of a time factor $\exp (-j\omega t)$, the incident wave $E_i$ is given by:
\[ E_i(x) = \hat{e}_i e^{jkix} \]  

(2.27)

Where, the unit vector in the direction of propagation is \( \hat{e}_i \). The amplitude of \( E_i \) is chosen to be one. The direction of polarisation is given by \( \hat{e}_i \). The electric field induces microscopic currents within the particle which reradiate electromagnetic fields. In the far field of the object, i.e. at distances much larger than the wavelength and the diameter of the volume \( V \), the scattered field \( E_s \) behaves like a spherical wave. The total field given by:

\[
E(x) = E_i(x) + E_s(x) = \lim_{kr \to \infty} \hat{e}_i e^{jkix} + f(\hat{i}, \hat{\delta}) e^{ikr} / r
\]  

(2.28)

Where \( x = r \hat{0} \). The parameter \( f(\hat{i}, \hat{\delta}) \) is called the scattering amplitude. The ratio of power of the scattered to the incident waves defines the differential cross section \( \sigma \) of the particle (see also Equation 2.9) which is given by

\[
\sigma(\hat{i}, \hat{\delta}) = \lim_{r \to \infty} \frac{1}{|E_i|^2} 4\pi r^2 |E_s|^2 = 4\pi |f(\hat{i}, \hat{\delta})|^2
\]  

(2.29)

The problem is solved using Maxwell’s equation in the absence of free currents and charges:

\[
\nabla \cdot E = 0; \quad \nabla \cdot H = 0;
\]

\[
\nabla \times E = j\omega \mu_0 H \quad \nabla \times H = -j\omega \varepsilon(x) \varepsilon_0 E
\]  

(2.30)

Let us rewrite Ampère’s circuital Law in the following manner:

\[
\nabla \times H = -j\omega \varepsilon(x) \varepsilon_0 E + \mathbf{J}_{eq}
\]  

(2.31)

Where,

\[
\mathbf{J}_{eq} = -j\omega \varepsilon_0 [\varepsilon(x) - 1] \mathbf{E} \quad \text{in } V
\]

\[
= 0 \quad \text{outside}
\]  

(2.32)

Ampère’s Law would take on exactly the same form if the problem of finding the radiation field of a localized oscillating current embedded in vacuum is
considered. The only difference would be that, instead of \( J_{eq} \), Equation (2.31) would contain the free current source \( J \). This shows that \( J_{eq} \) represents an equivalent current source which generates the scattered wave. The solution of equation system (2.30) is given by

\[
\begin{align*}
\mathbf{E} &= \mathbf{E}_i + \mathbf{E}_s \\
\mathbf{H} &= \mathbf{H}_i + \mathbf{H}_s
\end{align*}
\] (2.33)

Where, the field \((\mathbf{E}_i, \mathbf{H}_i)\), represents the homogenous solution. In our case \((\mathbf{E}_i, \mathbf{H}_i)\) can be identified with the incident plane wave given by (2.33). The field \((\mathbf{E}_s, \mathbf{H}_s)\) is the scattered field generated by the microscopic currents within \( V \). It is given by:

\[
\begin{align*}
\mathbf{E}_s &= \nabla \times \nabla \times \mathbf{A} \\
\mathbf{H}_s &= -j\omega \varepsilon_0 \nabla \times \mathbf{A}
\end{align*}
\] (2.34)

Where, \( \mathbf{A} \) is the Hertz vector:

\[
\mathbf{A}(\mathbf{x}) = \frac{1}{4\pi} \int [\varepsilon(\mathbf{x}) - 1] \mathbf{E}(\mathbf{x}) \frac{e^{jk|x-x'|}}{|x-x'|} d^3\xi.
\] (2.35)

In the far field we can make the approximation

\[
\frac{e^{jk|x-x'|}}{|x-x'|} = \lim_{r \rightarrow \infty} \frac{e^{jkr}}{r} e^{-jk\hat{\mathbf{x}} \cdot \mathbf{\hat{x}}}
\] (2.36)

So that \( \mathbf{E}_s \) can be written as:

\[
\mathbf{E}_s = f(\mathbf{t}, \mathbf{\partial}) \frac{e^{jkr}}{r}
\] (2.37)

With

\[
f(\mathbf{t}, \mathbf{\partial}) = \frac{k^2}{4\pi} \int [\mathbf{\hat{\partial} \cdot \hat{\mathbf{\partial}} E(\mathbf{x})} [\varepsilon(\mathbf{x}) - 1] e^{-j k \hat{\mathbf{x}} \cdot \mathbf{\hat{x}}} d^3\xi
\] (2.38)

Equation (2.38) represents only a formal solution to our problem because, in general, the electric field \( E \) within the body is not known. However, for our purpose this solution is sufficient because it shows that the scattered amplitude depends on the dielectric properties of the medium and its geometrical arrangement or structure.
2.8.0 DIELECTRIC PROPERTIES OF NATURAL MEDIA

The dielectric behaviour of media arises because positive and negative charges move in opposite directions when an E field is applied, a phenomenon that is referred to as polarisation. Let us consider a perfect insulator. Insulators are build-up from groups of atoms or ions which we shall call a “molecule” here. In an inert gas our “molecule” would be a single atom, while in a solid such as potassium chloride a single KCl group may be considered as a “molecule” even though the structure is an ionic framework in which molecules do not exist.

When an external field is applied then the positive and negative charges within our “molecules” move into opposite directions forming electric dipoles. There are three ways in which polarisation may arise, two involving distortion and one orientation (Duffin, 1965). If the “molecule” has no permanent dipole then polarisation may be due to the relative movement of nuclei and electrons (electronic polarisation) or movement of positive and negative ions in a solid (ionic polarisation).

If the “molecule” has a permanent dipole moment then the external field imposes a preferential direction. This orientational polarisation is temperature dependent since thermal agitation randomises the directions of the dipoles. In an oscillating field the molecules will never be quite in phase with the applied field, and for high frequencies will not follow it at all. At low frequencies the three polarisation mechanisms may occur together. The orientational distribution disappears in general in the microwave region, ionic polarisation in the infra-red, and electronic polarisation in the visible and infrared.

2.8.1 Water

Amongst natural media only liquid water shows orientational polarisation. The permanent dipole moment of water is due to the triangular structure of the H₂O molecule. The axis joining the two hydrogen nuclei to the oxygen form an angle of about 104° (Schanda, 1986). Above a certain
frequency, called relaxation frequency, the dipoles cannot follow the oscillations of the applied electric field $E$ any longer and thus $\varepsilon'$ decreases. At the relaxation frequency the molecules behave like harmonic oscillators at resonant frequency and absorption ($\varepsilon''$) is highest. The dielectric constant of pure water obeys the relaxation spectrum of the Debye type, which is given by a three parameter expression (Ulaby et al., 1986):

$$\varepsilon = \varepsilon' + j\varepsilon'' = \varepsilon_\infty + \frac{\varepsilon_s - \varepsilon_\infty}{1 - jf/f_0},$$

(2.39)

Where $f$ is the frequency, $f_0$ is the relaxation frequency, $\varepsilon_s$ is the static dielectric constant, and $\varepsilon_\infty$ is the dielectric constant in the high frequency limit. The parameters are temperature dependent. Table 2.3 lists the parameters for 0°C and 20°C and the resulting values of the dielectric constant at 5.3 GHz. And Figure 2.8 shows the frequency dependency of $\varepsilon$ at 20°C.

Table: 2.3 Parameters of the Debye equation at 0°C and 20°C and the obtained values of the dielectric constant of pure water at 5.3 GHz (source: Wagner, 1998).

<table>
<thead>
<tr>
<th>Temperature</th>
<th>$f_0$ (GHz)</th>
<th>$\varepsilon_s$</th>
<th>$\varepsilon_\infty$</th>
<th>$\varepsilon'$</th>
<th>$\varepsilon''$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°C</td>
<td>9</td>
<td>88.3</td>
<td>5.3</td>
<td>66.9</td>
<td>36.2</td>
</tr>
<tr>
<td>20°C</td>
<td>17</td>
<td>80</td>
<td>4.9</td>
<td>73.3</td>
<td>21.3</td>
</tr>
</tbody>
</table>
2.8.2 Soil

The term soil refers to the weathered and fragmented outer layer of the earth's terrestrial surface (Hillel, 1980). Soils consist of three phases: a solid phase made up of mineral and organic matter, a liquid phase consisting of soil water, and a gaseous phase. In the absence of water the real part of the dielectric constant $\varepsilon'$ varies over the range from two to four (Ulaby et al., 1986). The imaginary part $\varepsilon''$ is typically lower than 0.05. For explaining the dielectric behaviour of wet soil, the soil water is usually divided into two fractions according to the forces that are acting on the water molecules (Hallikainen et al., 1985; Dobson et al., 1985).

The water molecules contained in the first molecular layer surrounding the soil particles are tightly held due to matrix and osmotic forces. This fraction is referred to as bound water and exhibits a dielectric dispersion spectrum that is very different from that of free water. The dielectric properties of bound water are not known quantitatively, but it is apparent that, with an increasing fraction of bound water, $\varepsilon'$ decreases. Therefore $\varepsilon$ depends on the textural composition of the soil which determines the specific surface area of the soil particles (Koorevar et al., 1983).

Figure 2.8: Dielectric constant $\varepsilon$ of pure water versus frequency at 20°C (Source: Wagner, 1998).
The term soil texture refers to the size range of particles in the soil. The array of possible particle sizes is divided into three fractions, namely sand, silt, and clay. In the definition of the U.S. Department of Agriculture (USDA), sand are particles falling in the range from 2000 mm down to 50 mm, silt from 50 to 2 mm, and clay from 2 mm downwards.

For several frequencies, Hallikainen et al. (1985) generated empirical expressions for $\varepsilon$, as a function of the volumetric water content $W$, sand (Sa) and clay (Cl) components in percent of weight. For 6 GHz their best fit polynomial is:

$$\varepsilon' = (1.993 + 0.002Sa + 0.015Cl) + (38.086 + 0.633Cl) \cdot W$$

$$+ (10.720 + 1.256Sa + 1.522Cl) \cdot W^2$$

$$\varepsilon'' = (-0.123 + 0.002Sa + 0.003Cl) + (7.502 - 0.058Sa - 0.116Cl) \cdot W$$

$$+ (2.942 + 0.452Sa + 0.543Cl) \cdot W^2$$

(2.40)

Figure 2.9 shows the dependence of the dielectric constant on soil wetness for a loamy soil with $Sa = 40\%$ and $Cl = 20\%$. It can be seen that $\varepsilon'$ increases from about 2.3 for a dry soil to 40 for a saturated soil surface.

Figure 2.9: Dielectric constant $\varepsilon$ of a loamy soil versus the volumetric soil moisture content in % (source: Wagner, 1998).
When soil temperature drops below 0°C, the dielectric constant decreases strongly because when water freezes, polarisation by orientation is not possible any longer. However, some soil water does not freeze even at temperatures around -50°C and therefore $\varepsilon$ of frozen soil still shows some dependency on the moisture content before freezing. Hallikainen et al. (1984) made dielectric measurements of soils in the 3 GHz to 37 GHz band between -50°C and 23°C. For several soil types, two samples with a volumetric water content of 5% and 25% were prepared in the laboratory and quickly frozen.

For silt loam at -2°C Hallikainen et al. (1984) found $\varepsilon'$ to be about 3.3 and 5.5 for the dry and wet samples respectively. Despite these values are slightly larger than for a completely dry soil sample, it can be concluded that the dielectric properties of a dry and frozen soil are similar.

2.8.3 Vegetation

Direct measurements of oven-dried samples of various types of vegetation material shows that the real part of the dielectric constant $\varepsilon'$ is generally between 1.5 and 2, and $\varepsilon''$ is below 0.1 (Ulaby and El-Rayes, 1987). Water may constitute between about 80% and 90% of the fresh weight of leave like plants (Börner, 1996; Kramer, 1983). Even woody plants contain more than 50% water. Therefore, $\varepsilon$ of vegetation increases strongly with water content.

El-Rayes and Ulaby (1987) and Ulaby and El-Rayes (1987) investigated the dielectric properties of vegetation and developed an empirical formula to model $\varepsilon$ as a function of volumetric water content of vegetation matter alone. They state that the model is capable of reproducing $\varepsilon$ of various vegetation elements, leaves, stalks, branches, trunks, with a relative error of ±5%. In the model it is distinguished between bound water that is held tightly by the organic compounds and free water that can move within the plants with relative ease.
To estimate $\varepsilon$ of both bound and free water components, a dispersion model of the Debye type is employed. While the relaxation frequency of the free water component was assumed to be equal to that of pure water (17 GHz at 20°C), the relaxation frequency of the bound water component was estimated to be about 0.18 GHz by conducting dielectric measurements for sucrose-water mixtures. To account for temperature effects, a slightly modified version of the model was presented by Ulaby et al. (1990). Figure 2.10 shows how $\varepsilon$ varies with the volumetric moisture content of vegetation according to this model at 20°C and 5.3 GHz. The model suggests that for a gravimetric vegetation having moisture content around 80 %, $\varepsilon'$ takes a value of above 40.

![Dielectric constant $\varepsilon$ of vegetation versus the gravimetric soil moisture content in % (source: Wagner, 1998).]

2.8.4 Ice and Snow

Over the entire microwave frequency range $\varepsilon'$ of ice is relatively constant, $\varepsilon' = 3.17 \pm 0.03$ (Mätzler, 1987). Ice has a remarkable property in microwave range. For frequencies below 10 GHz, $\varepsilon''$ is lower than 0.001. These losses are amongst the lowest in condensed matter and consequently the penetration depth of microwaves into ice is of the order of meters.

Snow is a mixture of ice, air, and water, depending on the temperature. For dry snow $\varepsilon'$ is lower than 1.7. The magnitude of the imaginary part $\varepsilon''$ for dry snow is not much different from that of ice. When temperatures approach 0°C, liquid
water is formed and the dielectric properties are changed significantly. Figure 2.11 shows how ε of snow changes with the volume fraction of liquid water in the snow pack according to the formulas given by Mätzler (1987). The penetration depth decreases rapidly with increasing water content and consequently dry and wet snow show distinctly different backscattering behaviour.

![Dielectric constant ε of snow versus the volume fraction of water in % (source: Wagner, 1998).](image)

**Figure 2.11:** Dielectric constant ε of snow versus the volume fraction of water in % (source: Wagner, 1998).

### 2.9.0 ACTIVE AND PASSIVE REMOTE SENSING

Remote sensing is classified into two branches based on the physical technique involved in the sensing, viz. active and passive. If the generated EMR is transmitted to the earth surface and the returned response is sensed then the technique is called active RS, whereas, in passive RS the reflected part of incoming EMR of sun or emitted EMR from the surface of earth, i.e. only natural radiation is being detected.

#### 2.9.1 Reflected Solar Radiation Sensors

These sensor systems detect solar radiation that has been diffusely reflected (scattered) upward from surface features. The wavelength range that provides useful information includes ultraviolet, visible, near infrared and
middle infrared ranges. Because they depend on sunlight as a source, these systems can only provide useful images during daylight hours and hence changing atmospheric conditions and changes in illumination with time of day and season can pose interpretive problems. Reflected solar remote sensing systems are the most common type, used to monitor Earth resources.

2.9.2 Thermal Sensors

Sensors that can detect the thermal radiation emitted by surface features reveal information about the thermal properties of the surface features. Like reflected solar sensors, these are passive systems that rely on natural radiation as the ultimate energy source. Because the temperature of surface features changes during the day, thermal sensing systems are sensitive to time of day at which the images are acquired. Thermal sensors utilize the thermal infrared or microwave portion of the EM spectrum and therefore they are known as thermal infrared sensors or passive microwave sensors.

2.9.3 Imaging Radar Sensors

Active remote sensors create their own Electromagnetic energy that 1) is transmitted from the sensor towards the terrain (and is largely unaffected by the atmosphere), 2) interacts with the terrain producing backscatter of energy, and 3) is recorded by the remote sensor’s receiver.

The returning energy provides information about the surface roughness and water content of surface materials and the shape of the land surface. Long-wavelength microwaves suffer little scattering in the atmosphere, even penetrating thick cloud cover. Imaging radar is therefore particularly useful in cloud-prone tropical regions. Depending on the technology used in the sensor system, they are known as RADAR, Synthetic Aperture RADAR (SAR) or scatterometer.
2.10 MICROWAVE EMISSIVITY OF SOIL

Microwave emissivity from a smooth bare soil surface depends on the dielectric properties of the soil. Analytically it can be expressed in terms of the Fresnel's reflection coefficients of the surface:

\[ e(\lambda) = 1 - R(\lambda) = 1 - |(\sqrt{\varepsilon} - 1)/(\sqrt{\varepsilon} + 1)|^2 \quad (2.41) \]

Where, \( \lambda \) = wavelength; \( R(\lambda) \) = soil reflectivity at \( \lambda \) and \( \varepsilon \) = dielectric constant at \( \lambda \). Dielectric constant of soil varies with the proportion of free and bound water molecules available in the soil layer. It was found that for the estimation/measurement of soil emissivity, expressing soil moisture in volumetric units is more advantageous to minimise the textural differences computed to that of expressing soil moisture in terms of gravimetric or percentage of field capacity (Pampaloni, 1989). Hallikainen et al. (1985) have given following analytical solution to compute the dielectric constant of soil considering the volumetric moisture (\( m_v \)), proportion of sand (S) and clay (C)

\[ \varepsilon = A_1 + A_2 * m_v + A_3 * m_v^2; \quad (2.42) \]

Where \( A_i = a_i + b_i * S + c_i * C \) and \( a_i, b_i \) and \( c_i \) are regression coefficients.

Soil emissivity in microwave region varies with the frequency of observation. Emissivity is higher for higher frequency (lower wavelength) and lower for lower frequency (higher wavelength) at specified amount of fractional water. However for low amount of water, the variation gets saturated.

Figure: 2.12 Relationship of emissivity over spectrum of microwave frequencies for different fraction \( f \) of surface water content (Source: Basist et al., 1998)
As seen from the figure 2.12, change in soil emissivity between two different microwave frequencies has linear relation with the available fraction of water. This forms the basis of the detection of soil moisture using passive microwave remote sensing.

The frequencies selected for observations from air-borne or space-borne remote sensing are from the region of microwave spectrum for which atmospheric transmissivity is high. Figure 2.13 shows the transmissivity as a function of frequency. Frequencies widely used by space borne-sensors are also shown. It is seen that atmospheric transmissivity is high for the lower frequencies. It is remarked that the 21 GHz is sensitive to atmospheric water vapour, and it is not directly used to obtain surface properties but as a corrective supplement.

![Graph showing transmissivity due to oxygen and water vapour absorption.](image)

Figure 2.13: Transmissivity due to oxygen and water vapour absorption.  
(Source: compiled by other)

### 2.10.1 Sensitivity of Features to the Selected Wavelength

Results of the sensitivity analysis carried (Berger et al., 2003) out to investigate the interaction of different features to selected
wavelength/frequency are shown in Figure 2.14. The scale is a relative change in brightness temperature for a unit normalized change in the parameter of interest which we wish to observe (or infer).

![Figure 2.14: Sensitivity of 'brightness temperature' to different features as a function of frequency. (Source: Berger et al., 2003)](image)

It is seen that the sensitivity to the soil moisture variation is very high for the frequencies less than 5 GHz. Vegetation biomass induced variations are significant for the frequency ranging from 7 GHz to 20 GHz. For the frequency greater than 30 GHz, sensitivity is higher for the variations in surface roughness. Frequencies around 22 GHz are highly sensitive to the integrated water vapour content present in the atmosphere. This sensitivity analysis provides very good insight into the selection of frequencies for different applications.

### 2.10.2 Penetration Depth

The microwave emission from soil is the result of the integration of the upwelling radiation from all depths. The importance of each level decreases with depth depending on the wavelength of the sensor and the dielectric profile of the soil. A sensor with a longer wavelength provides information on thicker soil layer than is obtained with a shorter wavelength. The depth of penetration (Elachi, 1987) is expressed as
\[ Lp = \frac{\lambda}{2\pi \sqrt{\varepsilon \tan \delta}} \]  

(2.43)

Where medium loss tangent \( \tan \delta = (\varepsilon'' / \varepsilon') \). It is less than 10\(^{-2}\) for pure ice, dry soil, etc., of the order of 10\(^{-1}\) for wet soil, sea ice and vegetation. Typically \( \varepsilon' \approx 3 \) for dry soil and hence \( Lp = 9.2\lambda \), which indicates that for dry low-loss soil, microwave radiation can penetrate up to nine times the wavelength. However for soil at field capacity \( \varepsilon' \approx 40 \) and which gives the \( Lp = 0.25\lambda \). This is in agreement with the experimental depth of 5 cm found for the L band wavelength of 22 cm.

2.11 EFFECT OF SURFACE ROUGHNESS

Analytical problem for \( \varepsilon \) is much more difficult for rough surface because of the increase in emissivity over the equivalent smooth surface. The surface area which interferes with the air increases due to increase in surface roughness, which increases the upwelling emission and reduces the sensitivity to soil moisture variations. This problem is more important for wet soil. A simple model for the computation of microwave emissivity from rough soil surface is given below (Choudhury et al., 1979; Choudhury and Tucker, 1987).

\[ e(\lambda) = 1 - R(\lambda) \exp(-hc \cos^2 \mu) \]  

(2.44)

Here \( h \) = Empirical roughness parameter, which is relative and depends on the EM wavelength and \( \mu \) = incidence angle of observation.

2.11.1 Rough Surface Scattering Models

Natural surfaces can be considered as rough, and the roughness is the dominant factor for the scattering behaviour of an EM wave. The roughness of any scattering surface is not an intrinsic property of that surface but depends on the properties of a wave being transmitted. Both, the frequency and the local angle of incidence of the transmitted wave, determine how rough or
smooth any surface appears to be. The relation of the EM wave in terms of its wavelength $\lambda$ to the statistical roughness parameter $s$ is given by $k_s$.

Thus with increasing wavelength, the roughness term is decreasing, consequently, the indication of relative roughness for any surfaces is depending on the wavelength as $k = \frac{2\pi}{\lambda}$. Also the local incidence angle plays an important role for defining the roughness condition of a surface. In the near field of the propagating EM wave, the surface appears rougher than in the far field, which can be compared with the reflection of the sunset over the sea (Hejnsek and Papathanassiou, 2005).

![Figure 2.15 Fresnel reflection scheme (source: Hejnsek and Papathanassiou, 2005)](source: Hejnsek and Papathanassiou, 2005)

Considering a constant wavelength and fixed local incidence angle, the interaction of a transmitted EM wave with surface of different roughness conditions can be in general treated as: the more rough the surface, the more diffuse the scattering is, or the smoother the surface, more directional is the scattering. The Fresnel reflectivity, as shown in Figure 2.15 above, considers an ideal smooth surface boundary. In the natural environment the surface condition varies from medium to rough.

The back scattered EM wave on a surface consists of two components, a reflected or coherent and a scattered or incoherent one. The coherent component reacts as a specular reflection on a smooth surface and thus in the case of a monostatic radar there is no scatter return. The incoherent component is a diffuse scatterer and distributes the scattering power in all
directions. As the surface becomes rougher, the coherent component becomes negligible and the incoherent component consists of only diffuse scattering.

Figure 2.16: Rates of roughness components demonstrated on a (a) smooth, (b) rough and (c) very rough surface (source: Hejnsek and Papathanassiou, 2005).

Defining a surface from an electromagnetic point of view as smooth or rough, as mentioned before, is obviously somewhat arbitrary (Figure 2.16). Nevertheless, two main criteria can be found to define a smooth surface, the Rayleigh and the Fraunhofer criterion, respectively. Considering a plane monochromatic wave, transmitted at some angle $\theta$, onto a rough surface (see Figure 2.17), it is a simple matter to calculate the phase difference $\Delta \phi$ between two rays scattered from separate points on the surface:

Figure 2.17: Diagram for determining the phase difference between two parallel waves scattered from different points on a rough surface (Hejnsek and Papathanassiou, 2005).
The Rayleigh criterion states that if the phase difference $\Delta \phi$ between two reflected waves is less than $\pi / 2$ radians, then the surface may be considered as smooth, and is defined by

$$h < \frac{\lambda}{8 \cos \theta}$$

(2.45)

The usage of a more stringent criterion, which is adapted to the EM wave region, is proposed in ULABY et al. (1982) and is called Fraunhofer criterion. This criterion considers a surface as smooth, if the phase difference is

$$\Delta \phi < \frac{\pi}{8},$$

$$h < \frac{\lambda}{32 \cos \theta}$$

(2.46)

2.12.0 BACKSCATTER FROM NATURAL SURFACES

Over the land surface, backscatter is related to surface roughness and dielectric properties as well as volume scattering from vegetation and snow cover. Due to such physical mechanism, it is possible to address some of the large-scale phenomena over the land surface biophysical variables retrieval. In addition, it is feasible to address problems related to snow hydrology and soil thawing. This thesis will address some of the important features of the land like vegetation (forests, agriculture), desert, urban, water body and soil, etc.

2.12.1 Backscatter from Bare Soil

When an electromagnetic wave impinges on a soil surface, a part of the energy is scattered at the boundary surface in all directions and the rest is transmitted forward into the soil. The penetration depth of the transmitted wave is limited by absorption and scattering losses in the inhomogeneous soil medium. Considering only absorption losses the penetration depth of C-band microwaves into the soil varies from about 10 cm, when dry, to less than 1 cm, when wet. In most circumstances the layer accessible to C-band microwaves is about 0.5 cm to 2 cm thick (Schmugge, 1983; Wilheit, 1978; Newton et al.,
Therefore scatter from a bare soil surface is determined mainly by the dielectric properties of the upper few centimetres of the soil (and hence the soil moisture content in this layer) and by the geometrical structure or roughness of the soil surface. Although some scatter may arise due to in-homogeneities within the soil medium (Schanda, 1987), these may be ignored in favour of the surface scattering contribution (Ulaby et al., 1982).

To explain the influence of surface roughness on scattering let us consider Figure 2.20 from Schanda (1986). If a wave is incident on a plane surface then it is specular reflection in the forward direction. In this case no energy is scattered backwards to the sensor, except for normal incidence. A concise mathematical solution to this problem exists and the formulae for calculating the reflected and transmitted waves are known as Fresnel’s formulae (Jackson, 1983).

When the surface is slightly rough the incident wave will partly be reflected into the specular direction and partly be scattered in all directions. The first component is called the coherent component because the phase front of the coherent wave is conserved. The diffusely scattered component is called the non-coherent scattering component because the phase coherence is deteriorated or even destroyed. As the surface becomes rough, as shown in the Figure 2.18, more and more energy is scattered diffusely, while the coherent component becomes negligible. This means that for incidence angles away from nadir (greater than about 10°), the energy scattered in the backward direction increases with increasing surface roughness (Ulaby and Batlivala, 1976; Ulaby et al., 1982).

If the dielectric constant of the soil increases then the ability of the soil surface to reradiate electromagnetic waves also increases, which is why, because $\sigma_0$ and the soil moisture content are positively correlated. In numerous field experiments it was found empirically that $\sigma_0$ expressed in dB is in a first approximation linearly related to the volumetric soil moisture content $W$ (e.g. Champion, 1996):
\[ \sigma^0 \text{ (dB)} = A + B \times W \]  \hspace{1cm} (2.47)

Where by definition \( A \) is the backscattering coefficient of a completely dry Soil surface and \( B \) is the sensitivity of \( \sigma^0 \) to changes in the surface soil moisture content. The regression coefficients \( A \) and \( B \) are dependent on soil surface roughness, incidence angle and soil texture (Dobson and Ulaby, 1986).

Figure 2.18: Specular and diffuse components of the radiation scattered at, a perfect plane, b slightly rough, c very rough surfaces. \( \theta_o \) and \( \theta_s \) are the incidence and scattering angles (Source: Schanda, 1986).

\( A \) is primarily controlled by surface roughness and the incidence angle. For the incidence angle range covered by the ERS Scatterometer, increases with increasing roughness and decreases with increasing incidence angles. The sensitivity \( B \) is dependent on soil texture, because for constant volumetric soil moisture levels \( \varepsilon \) varies with soil texture. In addition, \( B \) may be dependent on incidence angle but the literature is contradictory about this question (Ulaby and Batalivala, 1976; Bertuzzi et al., 1992; Autret et al, 1989; Champion and Faivre, 1997).

Theoretical research on scattering of electromagnetic waves by rough surfaces has been extensive (Tsang et al., 1985; Fung, 1994). These studies show that backscatter is dependent on the r.m.s height of the surface and the autocorrelation function of the surface height variations. In situations such as in controlled laboratory experiments, theoretical models like the Integral Equation Method Model (IEM) shows good agreement with experimental results. As
shown in Fung (1994) the IEM is very sensitive to the choice of the autocorrelation function. In fact, the IEM stands for a multitude of models with each prototype representing different surface roughness types.

This makes the use of these models in practice quite difficult, because the statistical properties of the surface are in general not known. For this reason empirical models that explain the variation of $\sigma^0$ with the soil dielectric properties, the r.m.s. height, frequency, and incidence angle were developed based on field experiments (Oh et al., 1992; Dubois et al., 1995; Champion, 1996). However, like the theoretic models, the empirical models often fail to provide results in good agreement with field observations. One reason might be that, the parameters that have been used to describe surface roughness (r.m.s. height, correlation length) do not well represent the statistical characteristics of area extensive targets (Davidson et al., 1998).

Lehrrsch et al. (1988a) studied the spatial variability of various surface roughness parameters, derived from 1 m long surface profiles and found that these parameters are most commonly spatially independent. Research to define a scale independent description of surface roughness is on-going (Manninen et al., 1998).

2.12.2 Backscatter from Vegetated Surfaces

Vegetation canopies are inhomogeneous media, comprising of scattering elements of many different sizes, shapes, orientations, and permittivities. Usually, the vegetation constitutes 1 % or less of the canopy volume (Attema and Ulaby, 1978) and the penetration depth are of the order of meters (Ulaby et al., 1982). Scattering is caused mainly by the dielectric discontinuities within the canopy volume (volume scattering) and for vegetation canopies of low height, by the underlying soil surface (surface scattering).

Scattering from vegetation is a complex phenomenon and elaborated models have been developed to model $\sigma^0$ in terms of vegetation and soil surface parameters (Ulaby et al., 1990; Karam et al., 1992; Saatchi et al., 1994). These models have been used to stimulate $\sigma^0$ of various vegetation
canopies with some success (e.g. Touré et al., 1994) but, unfortunately, their input data requirements are very demanding. For example, for model $\sigma_0$ of an aspen canopy, the Michigan Microwave Canopy Scattering Model (MIMICS) requires 19 parameters like leaf diameter, branch length, or trunk moisture and three probability functions representing the orientational distribution of leaves, branches, and trunks.

For the discussion of backscatter from vegetation, a simple model based on radiative transfer theory is useful. Radiative transfer theory formulates the problem of absorption, scattering, and creation of radiation within a volume filled with particles (Chandrasekhar, 1960). The first-order radiative transfer solution of the problem of scattering of vertically polarized radiation by a vegetation canopy consists of three terms (Fung, 1994):

$$
\sigma_0^{\text{can}} = \sigma_0^{\text{vol}} + \sigma_0^{\text{sur}} + \sigma_0^{\text{int}}
$$

$$
= \frac{\omega \cos \theta}{2} \left(1 - e^{\frac{2\tau}{\cos \theta}}\right) + \sigma_s^{\text{vol}}(\theta) e^{\frac{2\tau}{\cos \theta}} + 2\Gamma_v(\theta) \omega e^{\frac{2\tau}{\cos \theta}},
$$

(2.48)

Where $\sigma_0^{\text{can}}$ = backscattering coefficient of the vegetation canopy;

$\sigma_0^{\text{vol}}$ = volume scattering term;

$\sigma_0^{\text{sur}}$ = surface scattering term;

$\sigma_0^{\text{int}}$ = surface-volume interaction term;

$\sigma_s^{\text{vol}}$ = backscattering coefficient of soil surface;

$\omega$= single-scattering albedo of the canopy;

$\tau$ = optical depth or thickness of the canopy;

$\Gamma_v$ = Fresnel power reflectivity for vertically polarized radiation.

In this formulation it is assumed that, the vegetation elements are isotropic scatterer and that only single scattering is important. Further it is
assumed that the reflection at the surface for the interaction term can be calculated using the Fresnel power reflection coefficient.

The volume scattering term $\sigma^0_{vol}$ is that contribution to total backscatter, which is due to direct backscatter of the incoming wave by the vegetation elements. It is proportional to the single scattering albedo $\omega$, which is a measure of the scattering efficiency of the vegetation elements. The surface scattering term $\sigma^0_{sur}$ represent direct backscattering from the soil surface attenuated by the vegetation layer. It can be seen in Equation (2.32) that the term $\gamma^2 = \frac{2r}{e^{cos^2}}$ controls the relative contributions of the vegetation layer and the soil surface. $\gamma^2$ is called the two-way transmissivity of the vegetation layer as, it describes the attenuation that a wave experiences, when it travels two times through the canopy. With increasing incidence angle, the path length increases and consequently, $\gamma^2$ decreases. As a result, the vegetation contribution becomes more important at large incidence angles.

Figure 2.19: ERS Scatterometer data acquired during June 1993 over tropical forest in Congo (2°N, 17°E) is the fitted volume scattering term from the equation (source: Wegner, 1998).

Dense forest canopies are not transparent for C-band microwaves and are representative of pure volume scatterer (Figure 2.19). It is noted that ERS
Scatterometer measurements of tropical rainforest are used for sensor calibration to show exactly the incidence angle behaviour of the volume scattering term.

The interaction term $\sigma^0_{\text{int}}$ accounts for multiple scattering by the vegetation and the soil surface. For vertical polarisation the interaction term is in general much smaller than the volume and surface scattering terms (Ulaby et al., 1986). However, it may be important for wet soil surface conditions and relatively small values of the optical thickness. Under these conditions the interaction term partly, compensates for the attenuation of the soil contribution by the vegetation layer, especially at high incidence angles. This implies that the interaction term tends to reduce the loss in sensitivity to soil moisture as the optical depth increases (Fung and Eom, 1985).

The Cloud Model developed by Attema and Ulaby (1978) is of the same form as Equation (2.48), but without the interaction term. This model was employed in numerous studies to simulate $\sigma^0$ of various vegetation cover types: alfalfa (Attema and Ulaby, 1978), barley (Bouman, 1991), beet (Bouman, 1991; Clevers and Leeuwen, 1996, Leeuwen et al., 1996), boreal forest (Pulliainen et al., 1994 and 1996), corn (Attema and Ulaby, 1978), grass (Mo et al., 1984), milo (Attema and Ulaby, 1978), potato (Bouman, 1991), savanna vegetation (Magagi and Kerr, 1997), and wheat (Attema and Ulaby, 1978; Bouman, 1991; Champion and Guyot, 1991; Prévot et al., 1993a and 1993b; Taconet et al., 1994 and 1996). Extensions to the Cloud Model can (for example) be found in Ulaby et al. (1984) and Paris (1986).

To stimulate vegetation growth the optical depth was related to vegetation parameters like plant water content, plant height or leaf area index (LAI). In the majority of the studies, the single-scattering albedo $\omega$ was kept constant. Despite its theoretical background the Cloud Model may be considered to be an empirical model whose parameters, i.e. the single scattering albedo and optical depth, may be tuned to fit observations. For example, Wigneron et al. (1996) introduced an incidence angle dependency for $\omega$ and $\mathcal{T}$ to account for vegetation structure in radiometric measurements.
However, the relatively successful application of radiative transfer theory to a wide variety of vegetation types and radar systems shows that the main phenomena involved in the problem of backscattering from vegetation can be well described by the interplay of surface and volume scattering effects.

To illustrate principle backscattering trends in ERS Scatterometer data over vegetated land surfaces, a simple model based on radiative transfer theory is used (Table 2.4). The model is discussed in detail in Wagner (1998). It was developed by postulating its general form and by identifying the possible ranges of model parameters based on a literature review and on a comparison of model stimulations with ERS Scatterometer data acquired under known conditions in the Canadian Prairies and over tropical rain forest. A final set of vegetation and surface parameters, were chosen to achieve an agreement of model stimulations and ERS Scatterometer measurements over the Iberian Peninsula. The major characteristics of the model are:

- The ERS Scatterometer measured backscattering coefficient is modelled as a mixture of non-transparent (forests, bushes, shrubs) and translucent (grassland, agricultural land) vegetation. This formulation was adopted because the literature review showed that the single scattering albedo of translucent vegetation types like grasses or agricultural crops is, in general, smaller than the single scattering albedo of non-transparent vegetation like forests. It can be shown that, the percentage area of non-transparent respective translucent vegetation within one ERS Scatterometer pixel is important to explain the backscatter behaviour. The percentage area of one ERS Scatterometer pixel covered by non-transparent vegetation is denoted by Ant.
- Because currently available empirical models describing backscatter from bare soil surfaces (Oh et al., 1992; Dubois et al., 1995; Champion, 1996) fail to reproduce the observed incidence angle behaviour of ERS Scatterometer measurements, a simple model assuming a linear relationship between $\sigma^0$ in decibels and $\theta$ is used.
Model stimulations show that the slope of $\sigma^0(\theta)$ in the logarithmic range varies strongly with changing soil dielectric properties if the interaction term given in Equation (2.48) is used. Since this is not in line with empirical observations, a new solution for the surface-volume interaction term was proposed. This solution is based on the assumption that the soil surface is perfectly rough instead of assuming a perfectly flat soil surface which leads to the interaction term given in (2.49). Unfortunately, the proposed formulation of the interaction term contains a multiplier that depends on the exact form of the ground scattering phase and the phase function of the vegetation canopy, which are unknown. Therefore, this multiplier could only be determined empirically, which is why, because it cannot be said, if it is physically meaningful or not. However, the new term is used to improve the agreement of model stimulations with observations. Only for incidence angles greater than $40^\circ$ and wet soil conditions, it contributes more than 1 dB to total backscatter.

$$\sigma^0 = (1 - A_{nt}) \cdot \sigma^0_{nt} + A_{nt} \cdot \sigma^0_{nt}$$

$$\sigma^0_{nt} = \frac{\omega_{nt} \cos \theta}{2}$$

$$\sigma^0_{nt} = \frac{\omega_{nt} \cos \theta}{2} \left( 1 - e^{2\tau_s} \right) + \sigma_s^0(\theta) e^{2\tau_s} + 2 z \Gamma_{nt}^n \omega_{nt} \tau_s e^{2\tau_s},$$

$$\sigma_s^0 = \sigma_{s,dry}^0(40) + \sigma_s'(\theta - 40) + s_m^m m_s,$$ in dB \hspace{1cm} (2.49)

The following trends are evident in the model stimulations:

- With increasing percentage area of non-transparent vegetation $Ant$, the backscattering coefficient, in general, increases, thus the slope of the curve $\sigma^0(\theta)$ becomes less steep, and the difference of $\sigma^0$ for wet and dry soil conditions decreases.

- Except for complete coverage by non-transparent vegetation types, $\sigma^0$ increases with increasing soil moisture.
The growth of translucent vegetation types, such as grasses or agricultural crops, results in a less steep decline of $\sigma^0$ with the incidence angle, and may increase or decrease $\sigma^0$, depending on the incidence angle and soil wetness.

Table: 2.4 Backscattering model for ERS Scatterometer data over vegetated land surfaces from Wagner (1998). The forth column shows the estimated range of possible values based on a literature review and a comparison with ERS Scatterometer measurements over the Canadian Prairies, the Iberian Peninsula and tropical rain forest. The last column shows the values used in stimulation.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Name</th>
<th>Unit</th>
<th>Range of Values</th>
<th>Value used in Simulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\sigma^0$</td>
<td>ERS Scatterometer backscattering coefficient</td>
<td>m^2/m^2</td>
<td>0 – 1</td>
<td></td>
</tr>
<tr>
<td>$\sigma^0_{nt}$</td>
<td>$\sigma^0$ of non-transparent vegetation</td>
<td>m^2/m^2</td>
<td>(0 – 100 %)</td>
<td>(0 – 100 %)</td>
</tr>
<tr>
<td>$\sigma^0_{tr}$</td>
<td>$\sigma^0$ of translucent vegetation</td>
<td>m^2/m^2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$A_{nt}$</td>
<td>Percentage area of non-transparent vegetation</td>
<td>deg</td>
<td>18 – 59</td>
<td>18 – 59</td>
</tr>
<tr>
<td>$\theta$</td>
<td>Incidence angle</td>
<td>deg</td>
<td>0.38 – 0.48</td>
<td>0.45</td>
</tr>
<tr>
<td>$\omega_{nt}$</td>
<td>Single scattering albedo of non-transparent vegetation</td>
<td>deg</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\omega_{tr}$</td>
<td>Single scattering albedo of translucent vegetation</td>
<td>deg</td>
<td>0.05 – 0.1</td>
<td>0.06</td>
</tr>
<tr>
<td>$\omega_{tr}$</td>
<td>Single scattering albedo of translucent vegetation</td>
<td>deg</td>
<td>0.05 – 0.3</td>
<td>0.1</td>
</tr>
<tr>
<td>$r_{tr}$</td>
<td>Optical depth of translucent vegetation:</td>
<td>Np</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\chi$</td>
<td>Empirical multiplier</td>
<td></td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>$\Gamma_0$</td>
<td>Fresnel power reflectivity at nadir</td>
<td>m^2/m^2</td>
<td>0.05 – 0.5</td>
<td>0.05 – 0.5</td>
</tr>
<tr>
<td>$\sigma^0_s$</td>
<td>$\sigma^0$ of soil surface</td>
<td>dB</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\sigma^0_{s, dry}(40)$</td>
<td>$\sigma^0$ of a dry soil surface at $\theta = 40^\circ$</td>
<td>dB</td>
<td>0.18 – 0.16</td>
<td>0.16</td>
</tr>
<tr>
<td>$\sigma^0_{s, wet}(40)$</td>
<td>Slope of $\sigma^0(\theta)$ at $\theta = 40^\circ$</td>
<td>dB/deg</td>
<td>0.03 – 0.4</td>
<td>0.36</td>
</tr>
<tr>
<td>$S_c$</td>
<td>Sensitivity of $\sigma^0$ to changes in the surface soil moisture content</td>
<td>dB</td>
<td>7 – 8</td>
<td>7</td>
</tr>
<tr>
<td>$n_s$</td>
<td>Degree of saturation of soil surface layer</td>
<td></td>
<td>0 – 1</td>
<td></td>
</tr>
<tr>
<td>$n_s$</td>
<td>Degree of saturation of soil surface layer:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Dry conditions</td>
<td></td>
<td>0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Wet conditions</td>
<td></td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

(Source: Wagner, 1998)

Figure 2.20: shows how $\sigma^0(\theta)$ varies according to the model from grassland to densely forested areas ($A_{nt}$), from dry to wet soil conditions.
conditions (increase of soil moisture), and from winter to summer (decrease of the optical depth of translucent vegetation types).

Figure 2.20 Stimulated backscattering coefficient for increasing percentage area of non-transparent vegetation Ant within one ERS Scatterometer pixel, and for dry/wet soil conditions and winter/summer conditions. The model and parameters used for the stimulation are given in Table 2.4.

With soil moisture studies, looking at retrieving vegetation, parameters have tended to be either empirical or model based. (Frison and Mougin 1996a, 1996b; Frison et al., 1998a, 1998b, 2000; Kerr and Magagi 1993; Long and Hardin 1994; Meseh and Quegan 1997; Mougin et al., 1995, 1995a; Pulliainen et al., 1996b; Schmullius 1997; Stephen et al., 2000; Wagner 1998a, 1998c; Watt et al., 1999, Wismann et al., 1996; Woodhouse et al., 1999),
Phenological observations on tree species in tropical moist forest of Uttara Kannada district during the years 1983-1985 revealed that there exists a strong seasonality for leaf flush, leaf drop and reproduction. Young leaves were produced in the pre-monsoon dry period with a peak in February, followed by the expansion of leaves which was completed in March. Abscission of leaves occurred in the post-monsoon winter period with a peak in December.

There were two peaks for flowering (December and March), while fruit ripening had a single peak in May-June, preceding the monsoon rainfall. The duration of maturation of leaves was the shortest, while that of full ripening of fruits was the longest. Mature flowers of evergreen species lasted longer than those of deciduous species; in contrast the phenol-phase of ripe fruits of deciduous species was longer than that of evergreen species (BHAT, 1992).

Empirical studies have implied that vegetation is not retrievable over a single season because it is not distinguishable from the varying soil moisture signal. Dual peak nature of temporal backscatter curve around the heading stage of rice crop was observed in Ku-band. The decrease of backscatter after first peak was associated with the threshold value of 60% crop canopy cover. The symmetric (Gaussian) and asymmetric (lognormal) curve fits were attempted to derive the date of initiation of the heading phase (Oza, et. al. 2008). A method is proposed to estimate both green leaf area index (GLAI) and soil moisture (hv), based on radar measurements at the Ku-band (14.85 GHz) and C-band (5.35 GHz) frequencies.

The Ku-band backscatter at large incidence angles was found to be independent of soil moisture conditions and could be used alone to estimate GLAI. Then, the Ku-band estimate of GLAI could be used with the measurement of C-band backscatter in a canopy radiative transfer model to isolate the value of hv (Moran et al. 1998).

This is certainly a limitation, but other studies have indicated that under some circumstances, it is possible to differentiate the surface scattering from
the vegetation component using a model-based approach (Woodhouse and Hoekman 2000a). Such methods are likely to be most effective if modelling approaches agree that, long term trends and large-scale spatial variations is sufficiently coupled with a suitable yield or growth model.

As discussed by Woodhouse and Hoekman (2000a) both the empirical and vegetation cover can be determined, although a complete interpretation of the observed variation is not yet clear. Results over Spain for instance have demonstrated that there is an underlying seasonal effect at high incidence angles related to changes in vegetation cover, yet the seasonal pattern does not correlate directly with the NDVI or known seasonal patterns (Woodhouse and Hoekman 2000b; Wagner et al., 1999b).

C- and Ku-band scatterometer backscatter time series are analysed and compared to meteorological data for two biomes, the African Steppe and the Scandinavian Boreal Forest (Scipal et al. 2002). The desert region looks brighter in optical bands whereas, higher brightness temperature and low backscattering coefficient are characteristics of desert region, respectively in passive and active microwave remote sensing responses (Mishra et al., 2002).

Estimates of the aerodynamic roughness lengths $z_0$ in arid and semi-arid regions are for the first time provided for the whole globe, using satellite ERS scatterometer observations. A statistical relationship is derived between the ERS scatterometer backscattering coefficients and quality in situ and geomorphological $z_0$ estimates (Prigent et al. 2005). A method to control a simple Sahelian land surface model, coupled to a radiative transfer model (RTM), on the basis of ERS wind scatterometer (WSC) observations.

The backscattering coefficient measured by space borne wind scatterometers over Sahel, shows a marked seasonality linked to the drastic changes of both soil and vegetation dielectric properties associated to the alternating dry and wet seasons. For lack of a direct observation, METEOSAT rainfall estimates are used to calculate temporal series of soil moisture with the
help of a water balance model. This information is used as input of the radiative transfer model that stimulates the interaction between the radar wave and the surface components (soil and vegetation) (Jarlan et al., 2003).

The quality of delivered products in terms of radiometry, geometry and additional processing for directional and atmospheric effects stands VEGETATION as an excellent tool for the monitoring of surface hydrology, crops, forest and land cover. In a first step, a sensitivity study is implemented to identify those parameters of the land surface model that can be estimated through the assimilation of WSC data. The assimilation scheme relies on evolution strategies (ES) algorithm that aims at solving the parameter evaluation problem. These algorithms are particularly well suited for complex (nonlinear) inverse problems (Jarlan et al., 2005). The phenology scheme developed for the Canadian Terrestrial Ecosystem Model (CTEM), designed for inclusion in the Canadian Centre for Climate Modelling and Analysis coupled general circulation model, is described (Arora et al. 2005).

In a comparison between the X-band and the C-band data it was found that the X-band data are more sensitive to the smaller scale undulations on the compound dunes and better revealed the full height of the dunes (Blumberg et al., 2006). A study has been carried out to analyse the high temporal Ku-band scatterometer data from QuikSCAT with 4.45 km resolution for regional assessment of rice crop phenology. Analysis shows the dual peak backscatter profile of rice crop (at tillering stage and another at maturity). Minima of the backscatter profile were found to coincide with the heading (Oza et al. 2007).

The forests of northeast India are subjected to severe fire episodes during the period of January to May every year mainly due to slash-and-burn agricultural practices. Lidar remote sensing is a breakthrough technology for deriving forest canopy structural characteristics (Dubayah et al. 2000).

Indian remote sensing satellite (IRS) data have been extensively used to map mangroves and other coastal vegetation for the entire country of the coastline (Nayak and bahuguna 2001). The WiFS vegetation cover map was
compared to the estimates of forest cover area derived from IRS LISS III images (Joshi et al. 2006).

Both optical and radar classifications allow to assess wetland characteristics that potentially influence plant and animal Meta community structure. Envisat imagery, however, was less suitable than Land sat imagery for the extraction of detailed ecological information, as only large wetlands can be detected.

2.12.3 Backscatter from Dry and Wet Snow

Dry snow is a horizontally stratified medium, which can be thought of, as consisting of ice particles of various shapes embedded in air. Although the grain clusters in coarse spring snow can exceed 5 mm in radius, snow grain radii normally varies between 50 μm and 1000 μm, representing a range from new snow to spring snow (Dozier, 1989). The scattering efficiency of snow grains is proportional to the third power of the ice grain radii (Mätzler, 1987). Therefore the largest grains are responsible for most of the cross section. When two snow layers are considered, the layer with the largest particle determines $\sigma^0$ (West et al., 1993).

At C-band, dry snow is highly transparent (Rott, 1993; Mätzler and Schanda 1984) because of the small scattering efficiency and negligible absorption capacity of the snow grains at this frequency. A study of ERS Scatterometer data over the Canadian Prairies has shown that, a dry shallow snow pack overlying a bare soil is almost identical to $\sigma^0$ of the snow free situation (Wagner, 1995; Wagner et al., 1995).

If liquid water is present in the snow layer then absorption increases by many orders in magnitude and dominates all other snow-microwave interactions. According to Weise (1996) a water layer with a column height as small as 0.001 mm is detectable. The radar return of a wet snow layer is thus dominated by scattering of the air-snow boundary. If the snow surface is smooth, then $\sigma^0$ of a wet snow surface may be lower than $\sigma^0$ for bare soil (Rott and Nagler 1993 and 1994; Kosokowsky et al., 1993).
If, however, the surface is very rough then \( \sigma^0 \) of wet snow is comparable to \( \sigma^0 \) of bare soil, then it is no more possible to detect the presence of wet snow. For example, it was not possible to identify wet snow in the Canadian Prairies with the ERS Scatterometer (Wagner, 1995; Wagner et al., 1995). The reason for this is thought to be: generally shallow snow peaks exhibits a rough surface because it follows the underlying ground and dormant vegetation.

2.12.4 Backscatter from Water Surfaces

The penetration depth of C-band microwaves into water is less than about 2 mm and therefore, as is the case for bare soil and wet snow, \( \sigma^0 \) of water is dependent on the roughness of the surface. When the water surface is calm then specular reflection occurs and \( \sigma^0 \) at off-nadir angles is very low. Wind generates water waves that increase scattering into the backward direction. The main contributions do not come from large ocean waves, even if they are many meters in height. Rather, scattering is dominated by short waves that ride on the top of the larger waves (Ulaby et al., 1982).

The radar return is highest, when the radar looks into the upwind or downwind direction and is smallest, when it looks normal to the wind vector. Therefore \( \sigma^0 \) of water surfaces depends on the azimuthal look direction. These dependencies allow estimating the direction and speed of wind from the backscatter triplet measured with the ERS Scatterometer.

2.12.5 Backscatter from Desert

Ridley et al. (1996) used Satellite radar altimeter measurements and studied spatial and temporal variations in the backscattered signal from desert surfaces. In order to understand these variations they had carried out extensive in situ measurements for the validation of a compilation of existing backscatter models. Over the Sahara desert, Marticorena et al. (2004) had prepared a map of roughness length \( Z_0 \) using the POLDER instrument and a corresponding map derived from a geomorphologic classification. This has been tested by comparing the predicted dust event frequencies obtained, to dust indices (IDDI) derived from Meteosat IR observations.
Singhvi and Kar (2004) observed that the wind and the sand dynamics cease with the arrival of monsoon rains (end of June along the eastern margin of the desert, and mid-July in the western part. His analysis shows that higher wind strength and lower rainfall favour erosivity of the wind. Prigent et al. (2005) provided for the first time the aerodynamic roughness lengths $z_0$ in arid and semi-arid regions for the whole globe, using satellite ERS scatterometer observations. Blumberg, (2006) used DEM data from the C-band 90 m data set and the X-band 30 m data set, to map and characterize the height and spacing of the large sand seas (Ergs) on Earth.

2.12.6 Backscatter from Soil Moisture

The agronomists are interested in soil moisture because of either too little or too much water, mostly the former limit plant growth. According to Kozlowski (1968) tremendous losses in plant growth occur annually because of recurrent or sustained internal water deficits in plants. He thinks that these losses are not realized because, for many areas, data is not available to indicate how much more growth would occur if plants had favourable water supplies throughout the growing season.

In meteorology and climate change studies, soil moisture is important because it directly affects the partitioning of energy at the surface between latent and sensible heating. Therefore the hydrology of the Earth’s surface needs to be incorporated into General Circulation Models (Manabe, 1969). Modelling studies show that strong perturbations in soil moisture on global and regional scales can affect atmospheric circulation, and may persist for several months (Dirnmeyer and Shukla 1993). For example, over the North American plains, there is a tendency for dry springs to be followed by hot summers, and wet springs to be followed by cool summers. Also in hydrology soil water plays an important role, because excess soil moisture can lead to large runoffs and stream flows while soil water deficits can aggravate a hydrological drought (Lawford, 1992). Traditional methods for measuring soil moisture are essentially point measurements.
The thermo gravimetric method consists of removing a soil sample and by determining its weight before and after it has been dried in an oven at 105°C for 24 hours (Hillel, 1980). It is the standard method for soil wetness determination on which all other methods are ultimately calibrated (Schulin et al., 1992). In the Time- Domain Reflectory (TDR) method the velocity of propagation of a high frequency voltage pulse in the soil is measured and related to the soil dielectric properties. Since the dielectric constant of a soil increases with the water fraction, the soil moisture content can be estimated (Topp, 1992).

Other measurement devices are capacitance probes, resistance probes, neutron probes, and tension meters. One problem of these traditional measurement techniques is the great spatial variability of soil moisture that scales from 1 millimetre to hundreds of meters. And therefore a question arises, how reliable can we derive the value of area soil wetness (Schulin et al., 1992). An even bigger problem may be that they are labour intensive and costly (Hollinger and Isard, 1994; Rombach and Mauser, 1997). Consequently, only relatively few programs have accumulated substantial soil moisture data (Georgakakos and Baumer, 1996). The lack of data is strongly felt by hydrologists and scientists from related subjects. Engman (1986) wrote that ‘improved performance of hydrological processes and models is pretty much at a standstill because of lack of proper types and amounts of data’. Dirnmeyer (1995) states that, ‘the lack of widespread observations of soil moisture continues to hamper efforts to verify and improve hydrological models’. Blöschl and Sivapalan (1995) think that ‘reliable measurements of soil moisture patterns would be ideal for model evaluation and could be the key to progress in hydrological Modelling’. In this thesis the term “scale” refers in general, to a characteristic time or length.

Determining surface soil moisture has been the focus of many scatterometer studies. (Magagi and Kerr 1997; Pulliainen et al., 1998; Wagner 1998b, 1998d; Wagner et al., 1999a-1999b, 1999c, 2000a, 2000b; Woodhouse and Hoekman 2000b). Such studies tend to take either a physical modelling approach (e.g. Woodhouse and Hoekman 2000a; Pulliainen et al., 1998) or a
straightforward empirical approach (e.g. Wagner et al., 1999; Wagner and Scipal, 2000). Studies using the latter have clearly demonstrated the sensitivity of scatterometer measurements to variations in surface soil moisture but, often, require calibration for a given geographical area.

The modelling approach on the other hand has provided insight into the measurement process and helped demonstrate a methodology that can provide quantitative estimates and associated uncertainties. It is, however, difficult to effectively implement and may require substantial development to provide an operational methodology. The method developed by Wagner et al. (1999a) has been tested in two pre-operational application demonstrations in the Ukraine (Wagner et al., 2000) within the Data User Programme of ESA. Medium to low resolution (1–50 km) active microwave sensors such as space borne scatterometers and wide-swath mode synthetic aperture radars have great potential, as tools for long term monitoring over land and ice (Manuela and Woodhouse, 2002).

Jackson and Hsu (2002) Studies have shown the advantages of low-frequency (<5 GHz) microwave sensors for soil moisture estimation. Although higher frequencies have limited soil moisture retrieval capabilities, there is a vast quantity of systematic global high-frequency microwave data that have been collected for 15 yrs. by the Special Sensor Microwave Imager (SSM/I). Despite its importance, availability of information on soil moisture is limited. Only recently it could be demonstrated that low resolution radar data in combination with a change detection method can resolve this constraint.

Experience gained in a number of successful pilot projects, lead to an initiative, setting up a global soil moisture archive (Scipal, 2002). The WSC backscattered signal may be represented as a combination of the vegetation and bare soil contributions weighted by their respective fractional covers. Over our temperate regions and time periods of interest, the vegetation signal is assumed to be principally due to forests backscattered signal. Then, forest contribution may be quantified from the analysis of the SAR image, and then
removed from the total WSC signal in order to estimate the soil contribution (Zribi et al., 2003).

Currently orbiting SAR sensors combined with available SVAT models could provide distributed profile soil moisture information with known accuracy at the watershed scale (Moran et al. 2004). The multi-scale behaviour of soil-drying rates is described using the richness–area curves, and characteristic curves are determined to four soil formations typical to a climatic gradient between Mediterranean and semi-arid environments in Israel (Svoray and Shoshany, 2004).

The first Earth-orbiting satellite to carry four complementary microwave experiments—the Radar Altimeter (ALT) to measure ocean surface topography by measuring spacecraft altitude above the ocean surface; the Seasat-A Satellite Scatterometer (SASS), to measure wind speed and direction over the ocean; the Scanning Multichannel Microwave Radiometer (SMMR) to measure surface wind speed, ocean surface temperature, atmospheric water vapour content, rain rate, and ice coverage; and the Synthetic Aperture Radar (SAR), to image the ocean surface, polar ice caps, and coastal regions (Evansa, 2005).

Pellarin et al (2006) investigates the ERS Scatterometer soil moisture products precision over a half-degree region in South-western France. Based on a high resolution soil moisture stimulation (1 km$^2$) validated at the local scale, the ERS-scat product is assessed at its own resolution (50 km$^2$). The study points out the suitable quality of the surface soil moisture product (r.m.s. error equal to 0.06 m$^3$m$^{-3}$ for the 4-year period), and assesses the retrieved root-zone soil moisture accuracy, provided by a semi-empirical methodology, exclusively based on surface soil moisture products.

Using model parameters derived from eight years of ERS scatterometer data, first global soil moisture maps have been produced from ASCAT data (Bartalis, 2007). A temporal analysis was carried out between backscatter on a local (1 km) and a regional (25 km) scale. The resulting scaling layer consist of
the coefficient of determination between the two scales, which can be interpreted as the amount of soil moisture variation on the local scale explained by the variation of soil moisture on the regional scale (Sabel et al. 2007). This study compares ERS scatterometer top soil moisture observations with stimulations of a dual layer conceptual hydrological model. The comparison is performed for 148 Austrian catchments in the period 1991–2000. The results indicate, that the agreement between the two top soil moisture estimates, changes with the season and the weight given to the scatterometer in hydrologic model calibration. The differences tend to be smaller for lower altitudes and the winter season. The average correlation between the two estimates is more than 0.5 in the period from July to October, and about 0.2 in the winter months, depending on the period and calibration setting. Using both, ERS scatterometer based soil moisture and runoff for model calibration, provides more robust model parameters than using either of these two sources of information (Parajka, 2009).