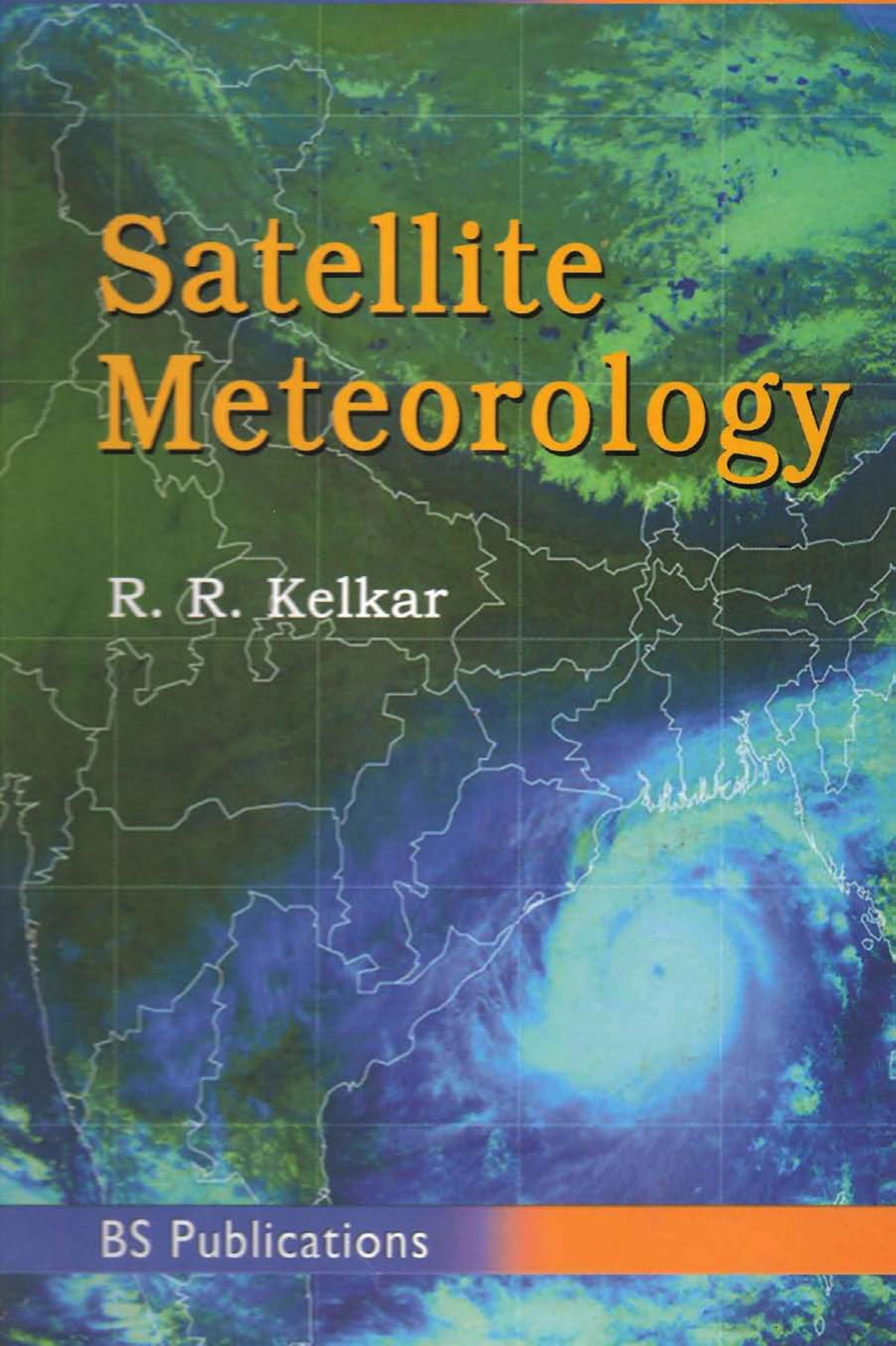

Satellite Meteorology

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BS Publications

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Chapter 1

Fundamentals

Fifty years ago, satellite meteorology did not exist except perhaps in science fiction. Today, satellite images of the world's weather, animated sequences of tropical storms, 5-day weather forecasts, are all being beamed through satellite channels into our television sets every hour or half-hour. Anyone, not just meteorologists, can access the latest images scanned by meteorological satellites around the world, on home PCs connected to the internet. The fascinating origin of satellite meteorology as an independent branch of the science of meteorology, and its phenomenal growth, have indeed had a touch of fantasy. What satellite meteorology happens to be today, is the result of an interplay of science on one hand, and the technology of satellites, computers and communications on the other. Limitations of technology have been overcome by scientific ingenuity, and the requirements of science have driven technology to the cutting edge.

After the successful launch of the first weather satellite in 1960 and the growth of satellite coverage of the earth's atmosphere and oceans within just a decade afterwards, meteorologists were in fact overwhelmed by the new satellite data that became available to them. Prior to the satellite era, meteorologists had laid the greatest emphasis on atmospheric pressure. Lows and highs showing up in synoptic isobaric analysis were of their main interest and many other weather elements including cloud cover, although observed, were not analysed on a synoptic scale in a similar manner. With the availability of satellite images, however, the accent shifted to observing and examining clouds and cloud patterns in the imagery, from which the state of the atmosphere could be directly observed or inferred. Images received from geostationary and orbiting satellites together revealed the presence of a wide spectrum of atmospheric phenomena across individual cumulus cells, thunderstorms, tropical cyclones and jet streams, just at a glance. So much so, that ground weather observation stations began to face the risk of redundancy and closure. Competition from satellite imagery forced many national meteorological services to take a re-look at their network configurations and trim them, particularly the upper air stations which are expensive to operate and maintain.

While the question whether meteorological satellites can give us all that we want for weather analysis and forecasting has been a subject of debate, the answer as of now at least, appears to be firmly in the negative. In spite of the limitations of conventional data and the clear advantages of satellites, the global ground observing network can not just be done away with. Over land, the atmospheric pressure at the surface is measured most accurately with barometers and automatic instruments. This is the data on which all synoptic weather charts are constructed. Over the sea, pressure data is very sparse. However, as of now, atmospheric pressure can not be retrieved from satellites, except that low pressure areas can be identified in a qualitative sense from certain cloud patterns in satellite imagery. It has not so far been possible to re-create a synoptic weather chart with remote sensing data alone.

The greatest help from weather satellites comes in observing weather over oceans, mountains, deserts, and unpopulated places where conventional data is either sparse or just unavailable. Here again, satellite data may not be able to replace the functions of an observing network for various reasons. For example, ship observations of sea surface temperature are made within the upper 1 m layer of the sea, while only the ocean skin temperature is retrieved from satellites, and it has errors associated with it due to the presence of clouds and atmospheric moisture. On ground, temperature measurement is made at 1.4 m height above the ground in a Stevenson Screen. It is extremely difficult to get this from satellites as the emissivity of land is highly variable and is not accurately known.

Over land, accurate measurements can be made with anemometers and automatic instruments and over sea, ships and buoys provide surface wind data. The popularly used cloud drift technique applied to satellite imagery cannot be used at the surface, but sea surface winds can be obtained from scatterometers.

An observer on the ground looks up to the sky and estimates visually how much of the sky above is occupied by clouds (expressed in 1/8ths of the hemisphere or oktas). Cloud amount has a different meaning when seen from a satellite looking down and becomes the fraction of an area covered by cloud within a prescribed area as seen from the satellite. From the ground, it is the height of the cloud base that can be estimated by an observer or determined by instrumental methods and vertical growth of cloud cannot be visualized. On the contrary, satellites can give the height of the cloud top and help in seeing the vertical growth of clouds from fall in temperature over time.

Rainfall is measured over land with raingauges and automatic instruments, is a point measurement and areal averages are difficult to obtain. Over sea, rainfall measurements are rarely available. From satellite data, large-scale rainfall can be estimated from the type of cloud, its persistence and assumed rain rate and areal averaging is easy. A space-borne precipitation radar can make direct measurements of rainfall over the sea.

Upper level winds are routinely measured with pilot balloons, radiosondes, wind profilers and radars. Balloon ascents are taken at the most 2-4 times a day due to the high cost of consumables. By using successive cloud imageries, winds can be estimated from the movement of cloud tracers. Here, the availability of suitable tracers determines whether the winds can be derived or not and different types of errors are involved in the derivation process.

Temperature and humidity profiles of the atmosphere are routinely obtained from the global radiosonde network. It is becoming increasingly expensive to run upper air stations, and over the sea only a few ships make radiosonde measurements. Vertical profiles of temperature and humidity are bring made globally with sounders on NOAA and GOES satellites, but the retrieval is a complex process and there are many problems particularly over hot and humid tropics.

In essence, the practices of weather analysis and forecasting currently in vogue employ a judicious combination of conventional and remotely sensed data, so as to compensate for the deficiencies of one source of data with the advantages of the other. The capability of weather satellites to observe and monitor weather systems is determined by various factors such as the number, type and resolution of spectral channels of the radiometer, the period of the satellite orbit which determines its revisit time, its height above the earth, the inclination of the orbit that delineates the geographical coverage, and so on.

1.1 Principles of Meteorological Remote Sensing

Remote sensing has been defined in various ways, but it is basically the process of observing an object in wavelengths that the human eye cannot perceive. The term has also developed a strong association with satellites, although aircrafts can be used for the same purpose. Remote sensing has applications in many diverse areas, ranging from monitoring of earth resources to medical diagnosis, but the basic principles are the same. In the forthcoming sections we shall discuss various aspects of remote sensing as applied to the area of satellite meteorology.

1.1.1 Absorption, Emission, Reflection and Scattering

It is quite a paradox that the radiative and thermal equilibrium of the earth-atmosphere system is not controlled by the two major constituents of the atmosphere, nitrogen and oxygen, but by some of the miscellaneous and numerous gases that all put together make up for 1 % of its volume. The behaviour of gases like water vapour, carbon dioxide, ozone, methane and other trace gases, and also particulate matter floating in the atmosphere, is what alters the radiation from the sun traversing the atmosphere and the radiation returned to the sun by the earth. Since satellite remote sensing is essentially the measurement of the returned radiation, it is important to know how the gases in the atmosphere influence the radiative processes in the atmosphere.

Atoms and molecules in a gas have electronic, rotational or vibrational energy, and absorption or emission of radiation takes place when there is a transition from one energy state to another. Absorption spectra of atoms, such as atomic oxygen and nitrogen, are associated with electronic transition and occur in the ultra-violet (UV) region of the electromagnetic spectrum. Tri-atomic molecules like those of water vapour, carbon dioxide and ozone, have additional rotational and vibrational transitions, which occur mainly in the infra-red (IR) region. In the visible (VIS) region, gases in the atmosphere account for very little absorption. The main absorption bands are those of three atmospheric gases: water vapour at 6.7 μ, carbon dioxide 15 μ and ozone 9.6 μ (1 μ or micron = 10^{-6} m). There are other minor absorption bands attributable to methane, nitrous oxide, and other gases.

Radiation from the sun gets reflected when it strikes a plane surface such as the ground or cloud tops and its direction gets altered. Depending upon the albedo or reflectivity of the surface, some part of the radiation will be reflected and the remaining amount will get absorbed by the medium or be transmitted through it. Snow and cumulonimbus cloud tops have high albedo values while the ocean surface reflects very little of the radiation falling upon it.

Air molecules and suspended particles or aerosols scatter radiation in the VIS wavelengths. When the size of the scattering particles is small compared to the wavelength of the incident radiation, the scattering is said to be of the Rayleigh type. In Rayleigh scattering, the intensity is inversely proportional to the 4th power of the wavelength, and the distribution of scattered radiation intensity is symmetric in both the forward and backward directions. When the sizes of the scattering particles become comparable to the wavelength of the incident radiation, the scattering processes is said to be of the Mie type, in which the angular distribution of the scattered radiation intensity is complex and does not remain symmetric. Rayleigh scattering by air molecules is what gives the blue colour to the sky, while Mie scattering by the larger sized particles and aerosols gives it a grayish appearance.

1.1.2 Black Body and Radiation Laws

Many fundamental laws governing the absorption and emission of electromagnetic radiation are commonly based on the concept of what is termed as a black body. This is largely a theoretical concept as ideal or perfect black bodies can be said to be almost non-existent. A black body is defined as an object that absorbs all radiation incident upon it, does not reflect any of it, and emits all energy at full efficiency for all wavelengths as per the following equation

$$B(\lambda,T) = 2hc^2 \lambda^{-5} / (e^{hc/\lambda kT} - 1)$$

where B is the energy in w m^{-2} μ^{-1},
T is the temperature of the black body in °K,
λ is the wavelength in μ,
h is Planck's constant 6.625×10^{-27} erg sec,
k is Boltzmann constant 1.38×10^{-16} erg °K^{-1},
c is the velocity of light 3×10^{10} cm sec^{-1}

This relationship of the black body emission to its temperature is known as Planck's Law. For any given temperature, Planck's Law gives a distribution of emitted energy or a characteristic spectrum of electromagnetic radiation that peaks at a certain wavelength. This peak shifts to shorter wavelengths for higher temperatures and the area under the curve grows rapidly with increasing temperature.

Wien's Displacement Law describes the temperature dependence of the black body radiation curves derived from Planck's Law and it states that the wavelength λ_{max} at which the black body radiation is the maximum is inversely proportional to the temperature T of the black body (in °K), or

$$\lambda_{max} = c / T$$

where c is a constant whose value is 2898 μ °K.

Stefan-Boltzmann's Law gives the total energy E emitted at all wavelengths by the black body at temperature T,

$$E = \sigma T^4$$

where σ is the Stefan-Boltzmann constant which has a value of 5.6705×10^{-5} erg cm^{-2} sec^{-1} $°K^{-4}$.

E is in fact the area under the Planck's Law curve. Thus, as the temperature of a black body increases, the shift of the peak emission to shorter wavelengths is governed by Wien's Law while the increase in the height of the curve is explained by Stefan-Boltzmann's Law. This increase in E with temperature is not linear, since it varies with the fourth power of the temperature.

There is another important law called Kirchhoff's Law which states that the ratio ε of emitted radiation to absorbed radiation is the same for all black bodies at the same temperature. This law forms the basis for the definition of emissivity. The emissivity of a perfect black body is 1 and that of a perfect reflector is 0.

Strictly speaking, the laws mentioned above are applicable only to a black body, but they are important in the real world as they can be applied as a close approximation to other bodies which have a very weak interaction with the surrounding environment and can be considered to be in a state of equilibrium. In the earth-atmosphere system, the earth's sea surface and thick clouds can be regarded as acting very similar to a black body.

1.1.3 Solar and Terrestrial Radiation

The radiation laws described in the previous section help us to understand the nature of an object and identify some of its thermal properties by interpreting the pattern of radiation emitted by it in different wavelengths. The spectrum of the solar radiation received at the top of the earth's atmosphere matches very well the spectrum of a black body having a surface temperature of about 5700 °K. This is called solar radiation and it has a peak at about 0.5 μ (Figure 1.1.3.1). The radiation emitted by a black body at the same temperature as the average temperature of the earth's surface which is 283 °K, peaks at about 10 μ (Figure 1.1.3.2). We therefore have to deal with two different radiation regimes, the radiation received by the earth from the sun, called solar or shortwave radiation and the radiation returned to space by the earth-atmosphere system, called terrestrial or longwave radiation. The intensity of terrestrial radiation is far less than that of solar radiation.

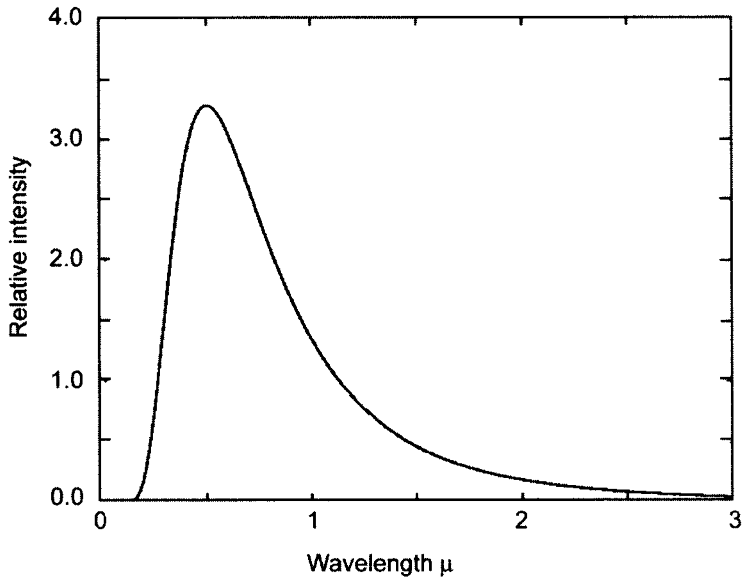

Figure 1.1.3.1 Variation of the relative intensity of black body radiation with wavelength (μ) at temperature 5700 °K, peaking at a wavelength of 0.5 μ in the visible region (Solar radiation)

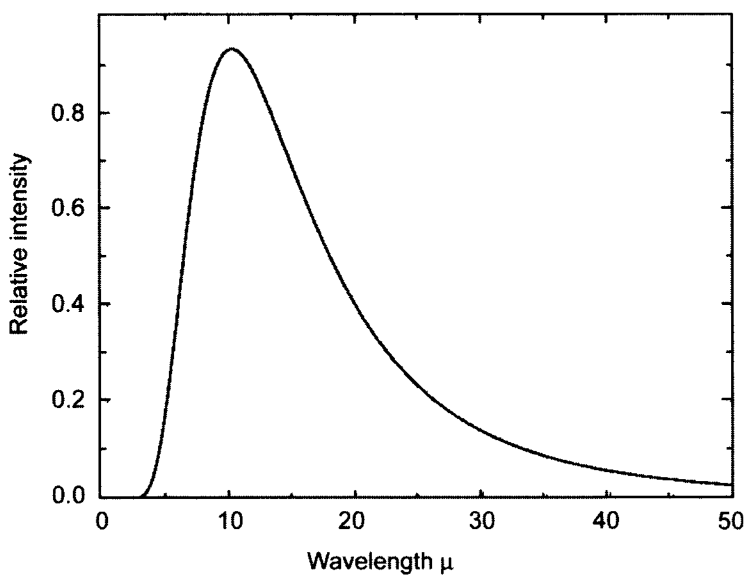

Figure 1.1.3.2 Variation of the relative intensity of black body radiation with wavelength (μ) at temperature 283 °K, peaking at a wavelength of 10 μ in the infra-red region (Terrestrial radiation)

1.1.4 Sun-Earth-Atmosphere Radiation Budget

The solar constant is defined as the annual average solar radiation received outside the earth's atmosphere on a plane normal to the incident radiation at the mean sun-earth distance and has a value close to 2 cal cm^{-2} min^{-1} or 1370 W m^{-2}. The actual solar irradiance varies by 3-4% of this value during the year due to the eccentricity of earth's orbit about the sun.

If we consider the incoming solar radiation at the top of the atmosphere as made of 100 units, 30 units are reflected back to space (6 by the atmosphere, 20 by clouds and 4 by the earth's surface). 19 units are absorbed by the atmosphere (16 by gases and 3 by clouds). The remaining 51 units are absorbed by the earth's surface.

Out of these 51 units, 6 are lost to space directly and 45 are returned upwards and absorbed by the atmosphere and clouds (7 by convection and conduction, 23 by evaporation as latent heat and 15 by longwave radiation). The atmosphere and clouds have already absorbed 19 units from the solar radiation, making a total of 64 units which are returned to space as longwave radiation. The budget is thus balanced at the top of the atmosphere.

At the earth's surface and at any level in the atmosphere, the net radiation is the balance of four radiative fluxes, downward solar radiation, downward longwave radiation, upward solar radiation and upward longwave radiation. These can be measured with special instruments installed on the ground or flown on balloons as radiometersondes. The prime factors involved in the radiation budget of the earth-atmosphere system are the albedo or reflectance properties of land, ocean and cloud tops, scattering properties of aerosols, dust and particulate matter in the atmosphere, and vertical profiles of temperature and concentration of gases which absorb longwave radiation (water vapour, CO_2, ozone). Data on these variables if available can be used to compute the radiation budget components indirectly.

1.1.5 Electromagnetic Spectrum

By the term spectrum, we commonly mean the seven colours of visible light, like in the rainbow. In scientific parlance, it refers to the entire range of wavelength or frequency of electromagnetic radiation, visible light being just a small portion of it (Table 1.1.5.1). The characteristic spectrum of a given object is the pattern of electromagnetic radiation that it absorbs, transmits and emits. The product of the wavelength λ and frequency ν of electromagnetic waves is equal to c, the velocity of light. The associated energy $E = h.\nu = h.c / \lambda$ where h is Planck's constant. This means that as the wavelength of electromagnetic radiation increases, its frequency decreases and the associated energy gets reduced.

The region of the electromagnetic spectrum with which we are most concerned in real life is the region of visible light, to which the human eye is very sensitive and in which the sun and stars emit the strongest radiation. In recent times, we are getting familiar with other wavelength regions as FM radio stations, mobile phones, satellite television or microwave ovens become more and more a part of our daily life.

Table 1.1.5.1 Electromagnetic Spectrum

Wavelength		**Wavelength**	
10^{-6} nm	Gamma Rays (MeV)	1 mm	Millimetre Waves (mm)
10^{-5} nm		1 cm	Microwaves (cm, GHz)
10^{-4} nm		10 cm	
10^{-3} nm		1 m	
10^{-2} nm		10 m	Radio Waves (MHz, kHz)
10^{-1} nm	X-Rays (Å)	100 m	
1 nm		1 km	
10 nm		10 km	
100 nm	Ultra-violet (nm),	100 km	
1 μ	Visible, Near Infra-red (μ)	10^3 km	
10 μ	Thermal Infra-red (μ)	10^4 km	
100 μ	Far Infra-red (μ)	10^5 km	

Table 1.1.5.2 Wavelength Range of Visible Colours

Colour	**Wavelength**	
	(nm)	**(μ)**
Violet	380-430	0.38-0.43
Indigo	430-500	0.43-0.50
Blue	500-520	0.50-0.52
Green	520-565	0.52-0.565
Yellow	565-590	0.565-0.59
Orange	590-625	0.59-0.625
Red	625-740	0.625-0.740

The visible spectrum consists of the seven colours familiar to us which are identified by their wavelength in nm (1 nanometre = 10^{-9} m) or in μ (1 micrometre or micron = 10^{-6} m) as given in Table 1.1.5.2. The visible wavelength region is, however, an extremely small part of the whole spectrum. Radiation of wavelengths shorter than violet is called ultra-violet (UV) radiation. This has very high energy that can break chemical bonds, ionize molecules, damage skin cells or cause cancer. However, most of the UV radiation coming from the sun is absorbed by the layer of atmospheric ozone which resides in the stratosphere, and shields life on earth from its harmful effects.

X-rays have wavelengths that are even shorter than UV, which are expressed in Å (Angstrom Units or 10^{-10} m). Gamma rays have wavelengths that could be as short as 10^{-15} m and it is more convenient to express their magnitude in terms of their energy levels which are of the order of keV (kilo electron volts) or MeV (Million electron Volts). X-rays and gamma rays have great penetration power and have applications in astronomy, radioactivity and other fields.

Towards the other end of the visible spectrum, radiation which has wavelength higher than red is called infra-red (IR). The IR region of the spectrum can be further sub-divided into near (NIR), short-wave (SWIR), middle (MIR), and thermal (TIR) with increasing wavelength.

Radiation with still longer wavelengths are called millimetre waves, followed by microwaves and radio waves. These again are further classified with respect to their frequency as given in Tables 1.1.5.3 and 1.1.5.4.

1.2 Satellite Orbits

The design of an optimum orbit around the earth for a meteorological satellite is a complex process. There are two main classes of orbits, polar orbiting and geostationary, and they are complementary to each other. However, many new types of orbits have now come into use or are being considered. The following sections describe the fundamental principles behind orbit design, technical considerations and the operational implications.

Table 1.1.5.3 Nomenclature of Microwave and Radio Wave Frequencies

Abbreviation	Full Form	Frequency	Wavelength
EHF	Extremely high frequency (Microwaves)	30-300 GHz	1 mm-1 cm
SHF	Super high frequency (Microwaves)	30-3 GHz	1 cm-10 cm
UHF	Ultra-high frequency	3 GHz-300 MHz	10 cm-1 m
VHF	Very high frequency	300-30 MHz	1 m-10 m
HF	High frequency	30-3 MHz	10 m-100 m
MF	Medium frequency	3 MHz-300 kHz	100 m-1 km
LF	Low frequency	300-30 kHz	1-10 km
VLF	Very low frequency	30-3 kHz	10-100 km
VF	Voice frequency	3 kHz-300 Hz	100-10^3 km
ELF	Extremely low frequency	300-30 Hz	10^3-10^4 km

Table 1.1.5.4 Microwave Bands

Band	Wavelength	Frequency
mm-Band	1-7.5 mm	40-300 GHz
Ku-K-Ka- Band	0.75-2.5 cm	12-40 GHz
X-Band	2.5-4 cm	8-12 GHz
C-Band	4-8 cm	4-8 GHz
S-Band	8-15 cm	2-4 GHz
L-Band	15-30 cm	1-2 GHz

1.2.1 Gravitational and Astronomical Laws

There are certain classical laws that were originally formulated to explain the motion of planets in the solar system and their orbits around the sun. They are, however, very fundamental and general in nature and we now know that they are equally applicable to the orbits of artificial satellites placed around the earth and other planets or moons in the solar system.

Kepler's laws of motion state that: (i) A planet moves around the sun in an elliptical orbit, with the sun at one focus, (ii) The vector joining the sun's centre to the planet sweeps out equal areas in equal time, and (iii) The square of the period of revolution of the planet is proportional to the cube of its semi-major axis. As mentioned above, these laws also hold good for artificial satellites orbiting the earth.

Kepler's laws have to be considered in conjunction with another fundamental physical law. As per Newton's law of gravitation, the gravitational force between two bodies of mass m_1 and m_2 is proportional to the product of m_1 and m_2 and inversely proportional to the square of the distance r between them, or

$$F = G\, m_1\, m_2 / r^2$$

where G is the universal gravitational constant having a value of 6.67×10^{-8} $cm^3\, g^{-1}\, sec^{-2}$.

The forces acting on a satellite around the earth are the gravitational force and the centripetal force, which should balance for the satellite to attain a stable orbit. So we must have

$$m_s\, v^2 / R = G M_e m_s / R^2$$

where M_e and m_s are the masses of the earth and satellite respectively, R is the mean distance between them and v is the velocity of the satellite. The term m_s appears on both sides of the equation and can be cancelled out. We then have

$$v^2 = G M_e / R$$

For a circular orbit, the velocity v can be expressed as $2\pi R / T$ where T is the time period of revolution of the satellite, which is inversely related to its mean distance from the earth and is given by the equation

$$T^2 = 4\pi^2 R^3 / G M_e$$

This means that the time period, speed and acceleration of an artificial satellite orbiting the earth are not dependent upon its mass. So theoretically speaking we can put into orbit as big a satellite as we wish, the only practical constraint being that of lifting it into space with the rockets that we have.

We can evaluate T for a given R as

$$T = (4\pi^2 R^3 / G M_e)^{1/2}$$

R can be expressed as $R_e + h$, where R_e is the radius of the earth and h is the height of the satellite above the earth. For the limiting case of $h = 0$, and using $M_e = 5.98 \times 10^{24}$ kg, and $R_e = 6370$ km, T works out to be 84 min. In reality of course it is not possible to have a satellite grazing the earth. For h = 200 km, T will work out to be 88 min. For h = 1000 km, T will be 105 min, and this is a popular choice for meteorological satellites.

If T is set at 24 hr, the height of the satellite will work out to be about 35,840 km above the earth's surface. Such an orbit is called geosynchronous as the satellite matches the angular velocity of the earth at this height.

1.2.2 Orbital Elements

As depicted in Figure 1.2.2.1, the satellite's orbit is specified by the following parameters (Kelkar et al 1980):

(a) a - the semi-major axis of the orbit (defines the size of the orbit)
(b) e - the eccentricity of the orbit (defines the shape of the orbit)
(c) I - the inclination of the orbit (with respect to the earth's equatorial plane)
(d) Ω - the right ascension of the ascending node (longitude of the north-bound equatorial crossing)
(e) ω - the argument of the perigee
(f) f - true anomaly
(g) T_e - epoch time

For a given orbit, all the above fixed parameters are fixed, except f which specifies the position of a satellite at a given time.

The semi-major axis a of the ellipse is the maximum height of the satellite above the earth's surface. As per Kepler's third law, satellites move faster when they are close to the earth and slowly when they are further away. The semi-major axis is thus an indirect measure of the average speed or mean motion of the satellite.

The satellite orbit is generally an ellipse. The eccentricity e is an indicator of the shape of the ellipse. A circular orbit is a special case of an ellipse when e is 0, and as e increases towards the limiting value of 1, the ellipse becomes more and more eccentric or elongated.