The core of GIScience

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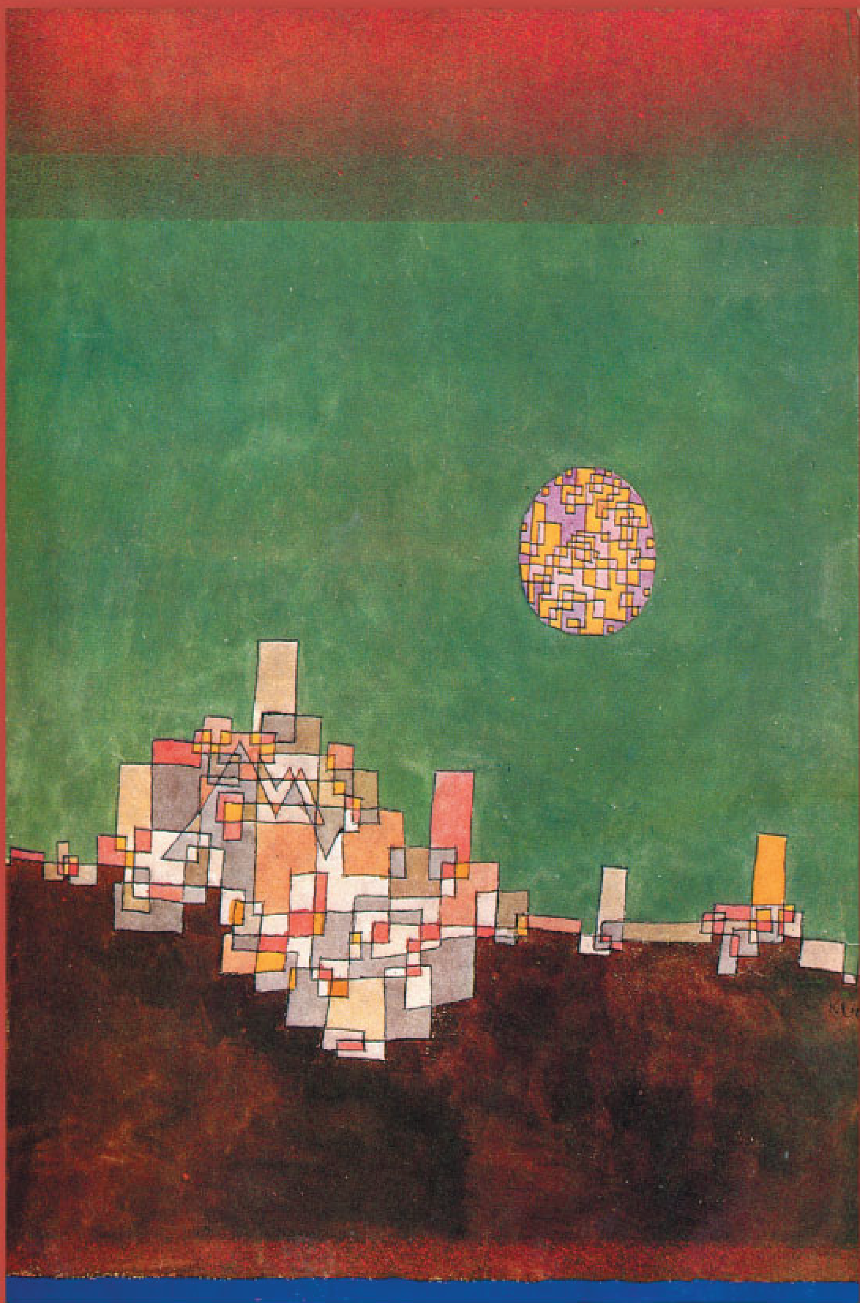

ITC Educational textbook series

UNIVERSITY OF TWENTE
FACULTY OF GEO-INFORMATION SCIENCE AND EARTH OBSERVATION

ITC

Chapter 2

Physics

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Introduction

Geospatial Data Acquisition (GDA) challenges us to make choices: on which one of the many sensors available should the agronomist rely for accurate yield predictions? If he or she chooses a sensor producing several images, such as a multispectral scanner, which image or which combination of images to use? How to properly process sensor recordings to increase the chances of a correct interpretation? When interpreting a colour image, what causes the sensation *red*? Instead of writing a thick book of recipes to answer such questions for every application, we can better review the physics of RS. Understanding the basics of electromagnetic (EM) radiation will help you in making more profound choices and enable you to deal with sensors of the future.

A standard photograph is an image of an object or scene that very closely resembles direct sensing with our eyes. The sensation of colour is caused by EM radiation. Red, green and blue relate to forms of radiation that we commonly refer to as light. *Light* is EM radiation that is visible to the human eye. As we are interested in Earth Observation, our light source is the Sun. The Sun emits light, the Earth's surface features reflect light, and the photosensitive cells (cones and rods) in our eyes detect light. When we look at a photograph, it is the light reflected from the photograph that allows us to interpret the photograph. Light is not the only form of radiation from the Sun and other bodies. The sensation *warm*, for example, is is the result of thermal emissions. Another type of emissions, ultraviolet (UV) radiation, triggers our body to generate vitamin D and also produces a suntan.

This chapter explains the basic characteristics of EM radiation, its sources and what we call the EM spectrum, the influence of the atmosphere on EM radiation, interactions of EM radiation with the Earth's surface, and the basic principles of sensing EM radiation and generic properties of sensors.

2.1 Waves and photons

EM wave

EM radiation can be modelled in two ways: by waves, or by radiant particles called photons. The first publications on the wave theory date back to the 17th century. According to the wave theory, light travels in a straight line (unless there are external influences) with its physical properties changing in a wave-like fashion. Light waves have two oscillating components: an electric field and a magnetic field. We refer, therefore, in this context to electromagnetic waves. The two components interact—an instance of a positive electric field coincides with a moment of negative magnetic field (Figure 2.1). The wave behaviour of light is common to all forms of EM radiation. All EM waves travels at the speed of light, which is approximately equal to 2.998×10^{8} m s^{-1}. This is fast, but the distances in space are literally astronomical: it takes eight minutes for the sunlight to reach the Earth, thus when we see, a sunrise, for example, the light particles actually left the Sun that much earlier. Because they travel in a straight line, we use the notion of light rays in optics.

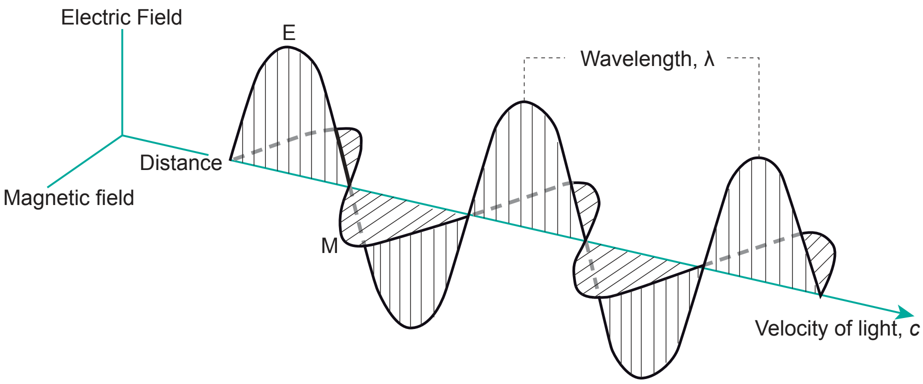

Figure 2.1
The two oscillating components of EM radiation: an electric and a magnetic field.

A sine wave can be described as:

$$e = \alpha sin\left(\frac{2\pi}{\lambda}x + \varphi\right). \tag{2.1}$$

wavelength

where α is the amplitude of the wave, φ is the phase (it depends on time) and λ is the *wavelength*. The wavelength is a differentiating property of the various types of EM radiation and is usually measured in micrometres (1 µm = 10^{-6} m). Blue light is EM radiation with a wavelength of around 0.45 µm. Red light, at the other end of the colour spectrum of a rainbow, has a wavelength of around 0.65 µm (Figure 2.2). Electromagnetic radiation outside the range 0.38–0.76 µm is not visible to the human eye.

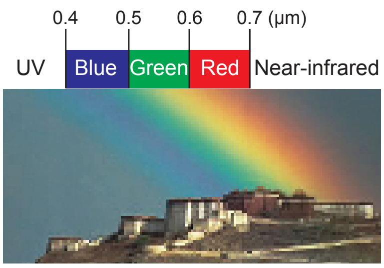

Figure 2.2
The spectrum of light.

We call the amount of time needed by an EM wave to complete one cycle the *period* of the wave. The reciprocal of the period is called the *frequency* of the wave. Thus, the frequency ν is the number of cycles of the wave that occur in one second. We usually measure frequency in hertz (1 Hz = 1 cycle s^{-1}). Since the speed of light c is constant, the relationship between wavelength and frequency is:

period and frequency

$$c = \lambda \times \nu. \tag{2.2}$$

Obviously, a short wavelength implies a high frequency, while long wavelengths are equivalent to low frequencies. Blue light has a higher frequency than red light (Figure 2.3).

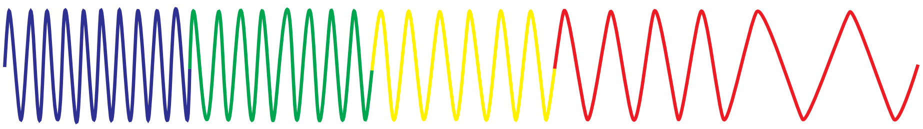

Figure 2.3
Relationship between wavelength, frequency and energy.

Although wave theory provides a good explanation for many EM radiation phenomena, for some purposes we can better rely on particle theory, which explains EM radiation in terms of photons. We take this approach when quantifying the radiation detected by a multispectral sensor (see Section 2.6). The amount of energy carried by a photon of a specific wavelength is:

energy of a photon

$$Q = h \times \nu = h \times \frac{c}{\lambda}, \tag{2.3}$$

where Q is the energy of a photon measured in joules (J) and h is Planck's constant ($h \approx 6.626 \times 10^{-34}$ J s).

The energy carried by a single photon of light is just sufficient to excite a single molecule of a photosensitive cell of the human eye, thus contributing to vision. It follows from Equation 2.3 that long-wavelength radiation has a low level of energy while short-wavelength radiation has a high level. Blue light has more energy than red light (Figure 2.3). EM radiation beyond violet light is progressively more dangerous to our body as its frequency increases. UV radiation can already be harmful to our eyes, so we wear sunglasses to protect them. An important consequence of Formula 2.3 for RS is that it is more difficult to detect radiation of longer wavelengths than radiation of shorter wavelengths.

2.2 Sources of EM radiation

All matter with a temperature above absolute zero emits EM radiation because of molecular agitation. *Planck's law of radiation* describes the amount of emitted radiation per unit of solid angle in terms of the wavelength and the object's temperature:

Planck

$$L(\lambda, T) = \frac{2hc^2}{\lambda^5} \frac{1}{e^{\frac{hc}{\lambda kT}} - 1}, \tag{2.4}$$

where h is the Planck's constant, $k \approx 1.38 \times 10^{-23}$ J K^{-1} is the Boltzmann constant, λ is the wavelength (m), c is the speed of light and T is the absolute temperature (K). $L(\lambda, T)$ is called the spectral radiance.

radiometric units

We can use different measures to quantify radiation. The amount of radiative energy is commonly expressed in joules (J). We may, however, be interested in the radiative energy per unit of time, called the *radiant power*. We measure the power in watts (W = J s^{-1}). *Radiant emittance* is the power emitted from a surface; it is measured in watts per square metre (W m^{-2}). *Spectral radiant emittance* characterizes the radiant emittance per wavelength; it is measured in W m^{-2} μm^{-1} (this is the unit used in Figure 2.4). *Radiance* is another quantity frequently used in RS. It is the radiometric quantity that describes the amount of radiative energy being emitted or reflected in a specific direction per unit of projected area per unit of solid angle and per unit of time. Radiance is usually expressed in W sr^{-1} m^{-2} (sr is steradian, unit of solid angle). *Spectral radiance* used in Equation 2.4 is radiance per wavelength and is measured in W sr^{-1} m^{-2} μm^{-1}. *Irradiance* is the amount of incident radiation on a surface per unit of area and per unit of time. Irradiance is usually expressed in W m^{-2}.

Planck's law of radiation is only applicable to black bodies. A black body is an idealized object with assumed extreme properties that helps us when explaining EM radiation. A black body absorbs 100% of incident EM radiation; it does not reflect anything and thus appears perfectly black. Because of its perfect absorptivity, a black body emits EM radiation at every wavelength (Figure 2.4). The radiation emitted by a black body is called black-body radiation. Real objects can re-emit some 80 to 98% of the radiation received. The emitting ability of real objects is expressed as a dimensionless ratio called emissivity $\epsilon(\lambda)$ (with values between 0 and 1). The *emissivity* of a material depends on the wavelength; it specifies how well a real body made of that material emits radiation as compared to a black body.

Wien's displacement law

The Sun behaves similarly to a black body. It is a prime source of the EM radiation that plays a role in Earth Observation, but it is not the only source. The global mean temperature of the Earth's surface is 288 K and over a finite period the temperatures of objects on the Earth rarely deviate much from this mean. The surface features of the Earth therefore emit EM radiation. Solar radiation constantly replenishes the energy that the Earth radiates into space. The Sun's temperature is about 6000 K. Planck's law of radiation is illustrated in Figure 2.4 for the approximate temperature of the Sun (about 6000 K) and the ambient temperature of the Earth's surface (288 K). The figure shows that for very hot surfaces (e.g. the Sun), spectral emittance of a black body peaks at short wavelengths. For colder surfaces, such as the Earth, spectral emittance peaks at longer wavelengths. This behaviour is described by *Wien's displacement law*:

$$\lambda_{max} = \frac{b}{T}, \tag{2.5}$$

where λ_{max} is the wavelength of the radiation maximum (μm), T is the temperature (K) and b ≈2898 μm K is a physical constant.

black body

We can use Wien's law to predict the position of the peak of the black-body curve if we know the temperature of the emitting object. The temperature of the black body determines the most prominent wavelength of black-body radiation. At room temperature, black bodies emit predominantly infrared radiation. When a black body is heated beyond 4450 K (approximately 4700 °C) emission of light becomes dominant, from red, through orange, yellow, and cyan, (at 6000 K) to blue, beyond which the emitted energy includes increasing amounts of ultraviolet radiation. At 6000 K a black body emits radiation of all visible wavelengths in approximately equal amounts, creating the sensation of white to us. Higher temperatures correspond to a greater contribution of radiation of shorter wavelengths.

The following description illustrates the physics of what we see when a blacksmith heats a piece of iron or what we observe when looking at a candle. The flame appears light-blue at the outer edge of its core; there the flame is hottest, with a temperature of 1670 K. The centre, with a temperature of 1070 K, appears orange. More generally, flames may burn with different colours (depending on the material being burnt, the surrounding temperature and the amount of oxygen present) and accordingly have different temperatures (in the range of $600\,^{\circ}$C to $1400\,^{\circ}$C). Colour tells us something about temperature. We can use colour, for example, to estimate the temperature of a lava flow from a safe distance. More generally, if we can build sensors that allow us to detect and quantify EM radiation of different wavelengths (also outside the visible range), we can use RS recordings to estimate the temperature of objects. You may also notice from the black-body radiation curves (Figure 2.4) that the intensity of EM radiation increases with increasing temperature; the total radiant emittance at a certain temperature is the area under the spectral emittance curve.

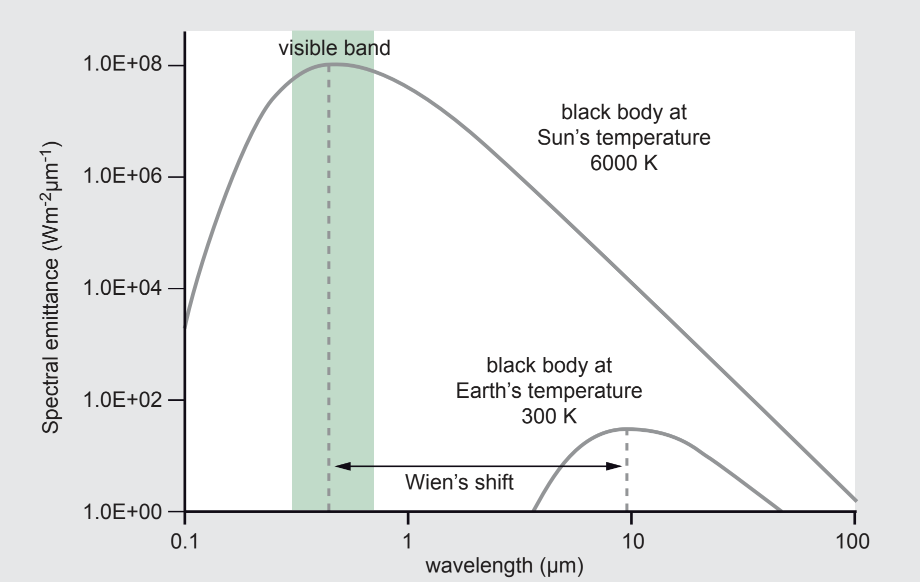

Figure 2.4
Illustration of Planck's law of radiation for the Sun (6000 K) and for the average surface temperature (300 K) the Earth. Note the logarithmic scale for both x- and y-axes. The broken lines mark the wavelength of the emission maxima for the two temperatures.

If you were interested in monitoring forest fires, which typically burn at 1000 K, you could immediately turn to wavelength bands around 2.9 µm, where the radiation maximum for those fires is to be expected. For ordinary land surface temperatures of around 300 K, wavelengths from 8 to 14 µm are most useful.

You can probably now understand why reflectance remote sensing (i.e. based on reflected sunlight) uses short wavelengths in the visible and short-wave infrared, and thermal remote sensing (based on emitted Earth radiation) uses the longer wavelengths in the range 3–14 µm. Figure 2.4 also shows that the total energy (integrated area under the curve) is considerably higher for the Sun than for the cooler Earth's surface. This relationship between surface temperature and total amount of radiation is described by the *Stefan-Boltzmann law*.

Stefan-Boltzmann law

$$M = \sigma T^4, \tag{2.6}$$

where M is the total radiant emittance (W m^{-2}), σ is the *Stefan-Boltzmann constant* ($\sigma \approx 5.6697 \times 10^{-8}$ (W m^{-2} K^{-4}), and T is the temperature in K.

The Stefan-Boltzmann law states that colder objects emit only small amounts of EM radiation. Wien's displacement law predicts that the peak of the radiation distribution will shift to longer wavelengths as the object gets colder. In Section 2.1 you will have learnt that photons at long wavelengths have less energy than those at short wavelengths. Hence, in thermal RS we are dealing with a small amount of low energy photons, which makes their detection difficult. As a consequence of that, we often have to reduce spatial or spectral resolution when acquiring thermal data, to guarantee an acceptable signal-to-noise ratio.

2.3 Electromagnetic spectrum

We call the total range of wavelengths of EM radiation the *EM spectrum*. Figure 2.2 illustrates the spectrum of visible light; Figure 2.5 illustrates the wider range of EM spectrum. We refer to the different portions of the spectrum by name: gamma rays, X-rays, UV radiation, visible radiation (light), infrared radiation, microwaves, and radio waves. Each of these named portions represents a range of wavelengths, not one specific wavelength. The EM spectrum is continuous and does not have any clear-cut class boundaries.

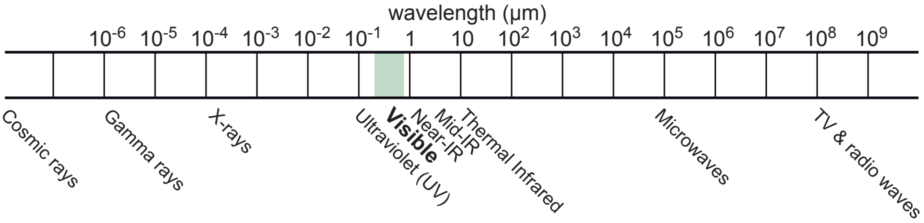

Figure 2.5
The EM spectrum.

Different portions of the spectrum have differing relevance for Earth Observation, both in the type of information that we can gather and the volume of geospatial data acquisition (GDA). The majority of GDA is accomplished by sensing in the visible and infrared range. The UV portion covers the shortest wavelengths that are of practical use for Earth Observation. UV radiation can reveal some properties of minerals and the atmosphere. Microwaves are at the other end of the useful range for Earth Observation; they can, among other things, provide information about surface roughness and the moisture content of soils.

light and colour

The "visible portion" of the spectrum, with wavelengths producing colour, is only a very small fraction of the entire EM wavelength range. We call objects "green" when they reflect predominately EM radiation of wavelengths around 0.54 μm. The intensity of solar radiation has its maximum around this wavelength (see Figure 2.8) and the sensitivity of our eyes is peaked at green-yellow. We know that colour effects our emotions and we usually experience green sceneries as pleasant. We use colour to distinguish between objects and we can use it to estimate temperature. We also use colour to visualize EM radiation we cannot see directly. Section 5.1 elaborates how we can "produce colour" by adequately "mixing" the three primary colours red, green and blue.

near-infrared, short-wave infrared

Radiation beyond red light, with larger wavelengths in the spectrum, is referred to as infrared (IR). We can distinguish vegetation types and the stress state of plants by analysing *near-infrared* (and *mid-infrared*) radiation—this works much better than trying to do so by colour. For example, deciduous trees reflect more near-infrared (NIR) radiation than conifers do, so they show up brighter on photographic film that is sensitive to infrared. Dense green vegetation has a high reflectance in the NIR range, which decreases with increasing damage caused by plant disease (see also Section 2.5.1). Mid-IR is also referred to as short-wave infrared (SWIR). SWIR sensors are used to monitor surface features at night.

Infrared radiation with a wavelength longer than 3 µm is termed thermal infrared (TIR) because it produces the sensation of "heat". Near-IR and mid-IR do not produce a sensation of something being hot. Thermal emissions of the Earth's surface (288 K) have a peak wavelength of 10 µm (see Figure 2.4). A human body also emits "heat" radiation, with a maximum at $\lambda \approx 10$ µm. Thermal detectors for humans are, therefore, designed such that they are sensitive to radiation in the wavelength range 7–14 µm. NOAA's thermal scanner, with its interest in heat issuing from the Earth's surface, detects thermal IR radiation in the range 3.5– 12.5 µm. Object temperature is a kind of quantity often needed for studying a variety of environmental problems, as well as being useful for analysing the mineral composition of rocks and the evapotranspiration of vegetation.

thermal infrared

2.4 Interaction of atmosphere and EM radiation

Before the Sun's radiation reaches the Earth's surface, three RS-relevant interactions in the atmosphere have occurred: absorption, transmission, and scattering. The transmitted radiation is then either absorbed by the surface material or reflected. Before reaching a remote sensor, the reflected radiation is also subject to scattering and absorption in the atmosphere (Figure 2.6).

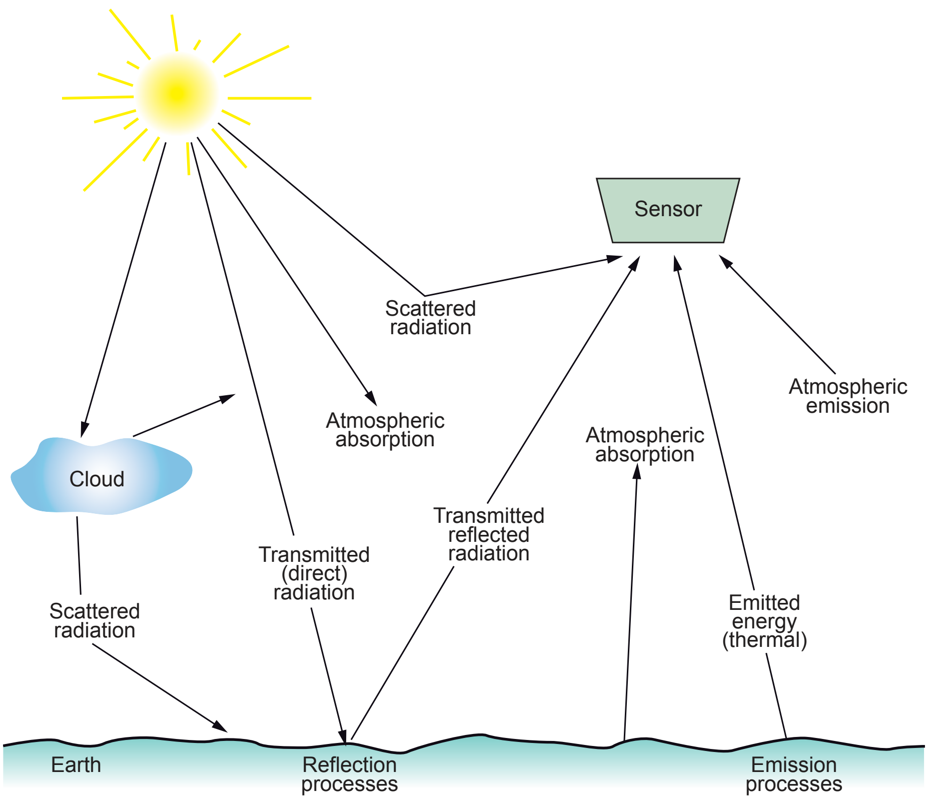

Figure 2.6
Interactions of EM radiation with the atmosphere and the Earth's surface.

2.4.1 Absorption and transmission

As it moves through the atmosphere, EM radiation is partly absorbed by various molecules. The most efficient absorbers of solar radiation in the atmosphere are ozone (O_3), water vapour (H_2O) and carbon dioxide (CO_2).

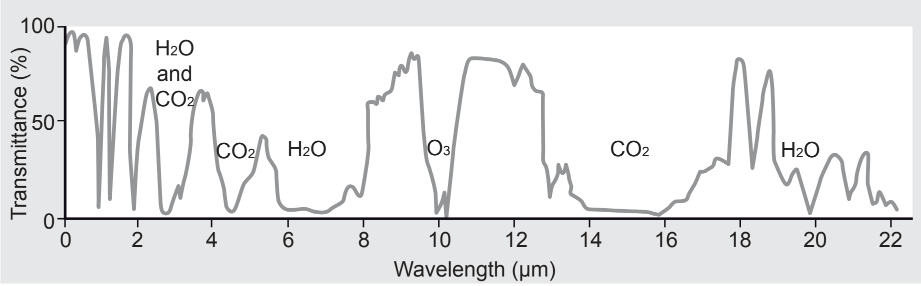

Figure 2.7
Atmospheric transmittance.

Figure 2.7 shows a schematic representation of atmospheric transmission in the wavelength range 0–22 µm. From this figure it can be seen that many of the wavelengths are not useful for remote sensing of the Earth's surface, simply because the corresponding radiation cannot penetrate the atmosphere. Only those wavelengths outside the main absorption ranges of atmospheric gases can be used for remote sensing. The useful ranges are referred to as *atmospheric transmission windows* and include:

atmospheric transmission

- the window from 0.4 to 2 µm. The radiation in this range (visible, NIR, SWIR) is mainly reflected radiation. Because this type of radiation follows the laws of optics, remote sensors operating in this range are often referred to as optical sensors.

- three windows in the TIR range, namely two narrow windows around 3 and 5 µm, and a third, relatively broad window extending from approximately 8 µm to 14 µm.

Because of the presence of atmospheric moisture, strong absorption occurs at longer wavelengths. There is hardly any transmission of radiation in the range from 22 µm to 1 mm. The more or less "transparent" range beyond 1 mm is the microwave range.

Solar radiation observed both with and without the influence of the Earth's atmosphere is shown in Figure 2.8. Solar radiation measured outside the atmosphere resembles black-body radiation at 6000 K. Measuring solar radiation at the Earth's surface shows that there the spectral distribution of the solar radiation is very ragged. The relative dips in this curve indicate the absorption by different gases in the atmosphere. We also see from Figure 2.8 that the total intensity in this range (i.e. the area under the curve) has decreased by the time the solar energy reaches the Earth's surface, after having passed through the atmosphere.

2.4.2 Atmospheric scattering

Atmospheric scattering occurs when particles or gaseous molecules present in the atmosphere cause EM radiation to be redirected from its original path. The amount of scattering depends on several factors, including the wavelength of the radiation in relation to the size of particles and gas molecules, the amount of particles and gases, and the distance the radiation travels through the atmosphere. On a clear day the colours are bright and crisp, and approximately 95% of the sunlight detected by our eyes, or a comparable remote sensor, is radiation reflected from objects; 5% is light scattered in the atmosphere. On a cloudy or hazy day, colours are faint and most of the radiation received by our eyes is scattered light. We may distinguish three types of scattering according to the size of particles in the atmosphere causing it. Each has a different relevance to RS.

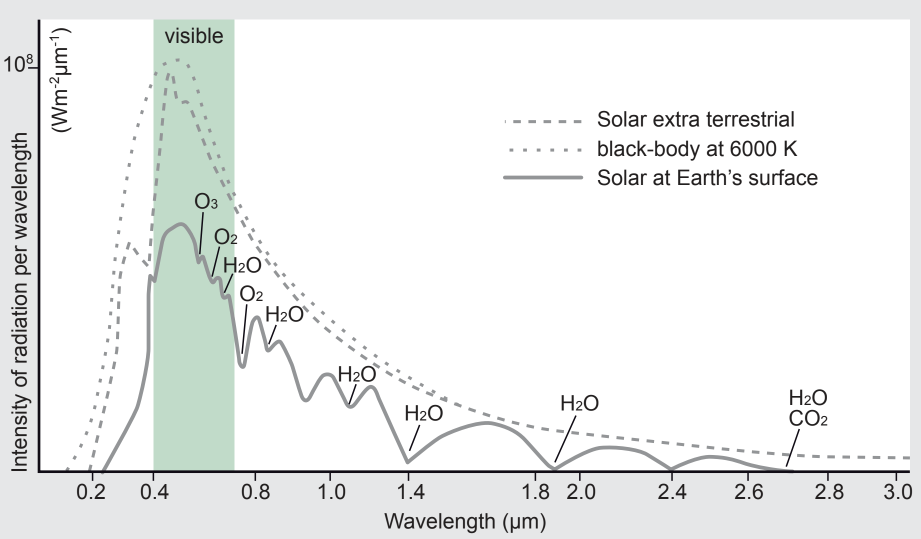

Figure 2.8
Radiation curves of the Sun and a black body at the Sun's temperature.

Rayleigh scattering

Rayleigh scattering dominates where electromagnetic radiation interacts with particles that are smaller than the wavelengths of light. Examples of such particles are tiny specks of dust and molecules of nitrogen (NO_2) and oxygen (O_2). Light of shorter wavelengths (e.g. blue) is scattered more than light of longer wavelengths (e.g. red); see Figure 2.9.

Figure 2.9
Rayleigh scattering is caused by particles smaller than the wavelengths of light and is greater for small wavelengths.

In the absence of particles and scattering, the sky would appear black. During the day, solar radiation travels the shortest distance through the atmosphere; Rayleigh scattering causes a clear sky to be observed as blue. At sunrise and sunset, the sunlight travels a longer distance through the Earth's atmosphere before reaching the surface. All the radiation of shorter wavelengths is scattered after some distance and only the longer wavelengths reach the Earth's surface. As a result we do not see a blue but an orange or red sky (Figure 2.10).

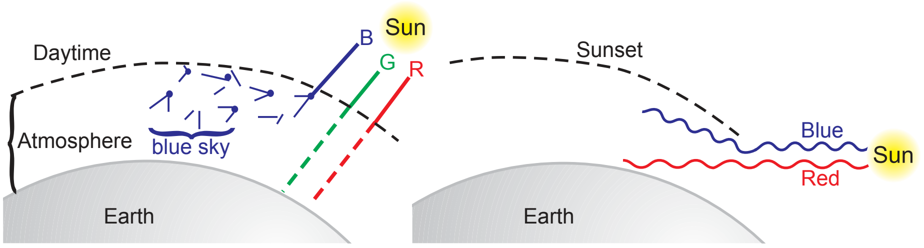

Figure 2.10
Rayleigh scattering causes us to see a blue sky during the day and a red sky at sunset.

Rayleigh scattering disturbs RS in the visible spectral range from high altitudes. It causes a distortion of the spectral characteristics of the reflected light as compared to measurements taken on the ground: due to Rayleigh scattering, the shorter wavelengths are overestimated. This accounts for the blueness of colour photos taken from high altitudes. In general, Rayleigh scattering diminishes the "crispness" of photos and thus reduces their interpretability. Similarly, Rayleigh scattering has a negative effect on digital classification using data from multispectral sensors.

Mie scattering

Mie scattering occurs when the wavelength of EM radiation is similar in size to particles in the atmosphere. The most important cause of Mie scattering is the presence of aerosols: a mixture of gases, water vapour and dust. Mie scattering is generally restricted to the lower atmosphere, where larger particles are more abundant, and it dominates under overcast, cloudy conditions. Mie scattering influences the spectral range from the near-UV up to mid-IR range and has a greater effect on radiation of longer wavelengths than Rayleigh scattering.

non-selective scattering

Non-selective scattering occurs when particle sizes are much larger than the radiation wavelength. Typical particles responsible for this effect are water droplets and larger dust particles. Non-selective scattering is independent of the wavelength within the optical range. The most prominent example of non-selective scattering is that we see clouds as white bodies. A cloud consists of water droplets; since they scatter light of every wavelength equally, a cloud appears white. A remote sensor like our eye cannot "see through" clouds. Moreover, clouds have a further limiting effect on optical RS: clouds cast shadows (Figure 2.11).

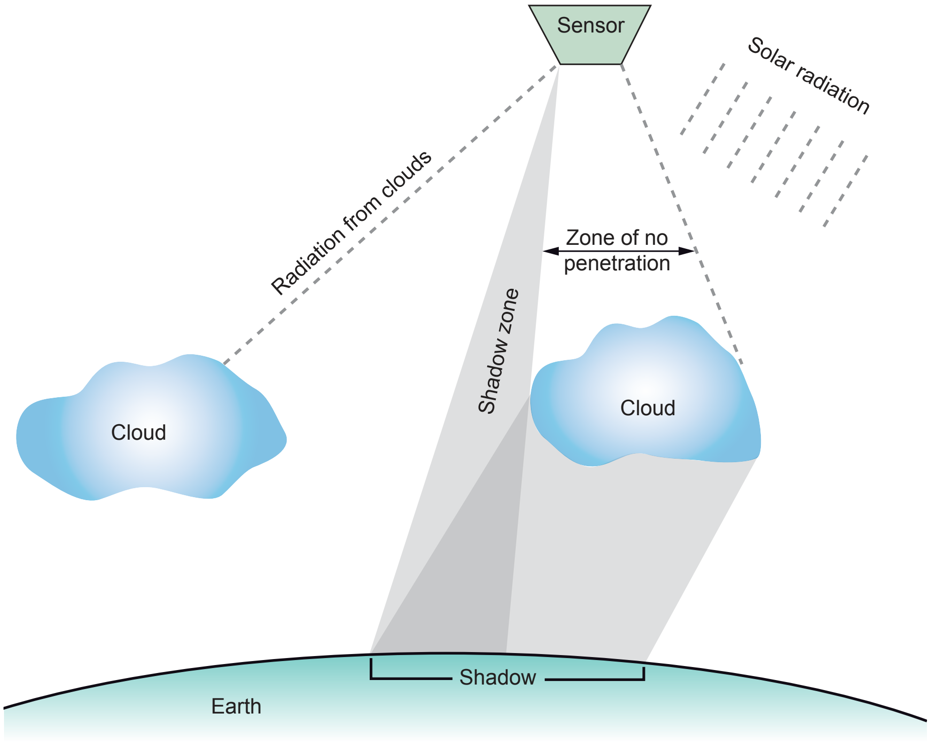

Figure 2.11
Direct and indirect effects of clouds for optical remote sensing.

2.5 Interactions of EM radiation with the Earth's surface

The EM radiation that reaches an object interacts with it. As a result of this interaction, EM radiation is absorbed, transmitted or reflected by the object. The energy conservation law, applied to interaction of EM radiation with the object, states that *all* incident EM radiation (I) is absorbed (A), reflected (R), or transmitted (T):

$$A(\lambda) + R(\lambda) + T(\lambda) = I(\lambda) \tag{2.7}$$

It is important to note that Equation 2.7 applies for each wavelength. Dividing both sides of Equation 2.7 by I we get.

$$\frac{A(\lambda)}{I(\lambda)} + \frac{R(\lambda)}{I(\lambda)} + \frac{T(\lambda)}{I(\lambda)} = \alpha(\lambda) + \rho(\lambda) + \tau(\lambda) = 1 \tag{2.8}$$

where $\alpha(\lambda)$ is absorptance, $\rho(\lambda)$ is reflectance and $\tau(\lambda)$ is transmittance of the object, all depend on wavelength λ and range from 0 to 1. For opaque objects $\tau(\lambda) = 0$ and Equation 2.8 reduces to

$$\alpha(\lambda) + \rho(\lambda) = 1 \tag{2.9}$$

Absorption of EM radiation leads to an increase in the object's temperature, while emission of EM radiation leads to a decrease in the object's temperature. The amount of emitted EM radiation is determined by the object's temperature (see Planck's law) and emissivity $\epsilon(\lambda)$. In equilibrium the total amounts of absorbed and emitted radiation at all wavelength are equal and the object's temperature is constant.

equilibrium

Kirchhoff's law of thermal radiation states that in equilibrium absorptance and emissivity at each wavelength are equal:

Kirchhoff's law

$$\alpha(\lambda) = \epsilon(\lambda) \tag{2.10}$$

The reflectance, transmittance and absorptance will vary with wavelength and type of target material. Here and further in the book we define a *target* as an object on the Earth surface that is being detected or sensed. Also the surface of target influences interaction of EM radiation and the target. Two types of reflection that represent the two extremes of the way in which radiation is reflected by a target are "specular reflection" and "diffuse reflection" (Figure 2.12). In the real world, usually a combination of both types is found.

reflection

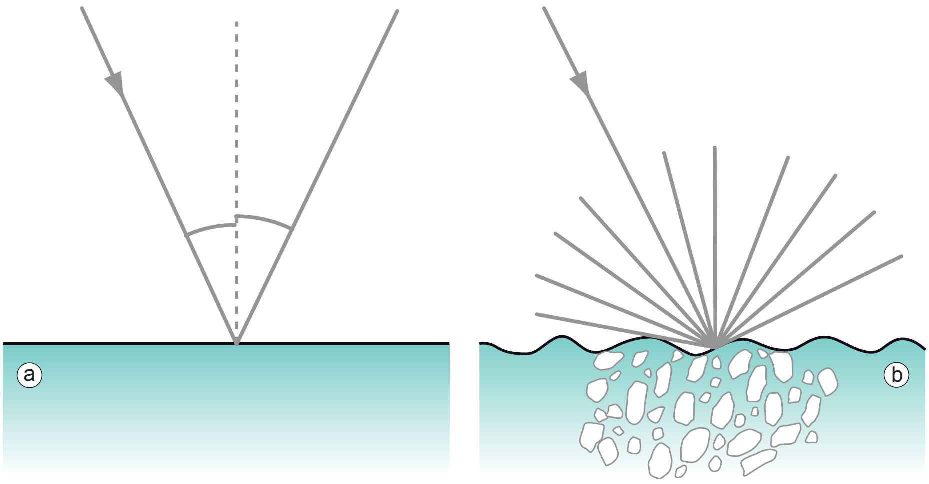

Figure 2.12
Schematic diagrams showing (a) specular and (b) diffuse reflection.

- *Specular reflection*, or mirror-like reflection, typically occurs when a surface is smooth and (almost) all of the radiation is directed away from the surface in a single direction. Specular reflection can occur, for example, for a water surface or a glasshouse roof. It results in a very bright spot (also called "hot spot") in the sensed image.

- *Diffuse reflection* occurs in situations where the surface is rough and the radiation is reflected almost uniformly in all directions.

Whether a particular target reflects specularly, diffusely, or both, depends on the surface roughness relative to the wavelength of the incident radiation.

2.5.1 Spectral reflectance curves

We can establish for each type of material of interest a *reflectance curve*. Such a curve shows the portion of the incident radiation ρ that is reflected as a function of wavelength λ (expressed as percentage; see Figure 2.13). Remote sensors are sensitive to ranges, albeit narrow, of wavelengths, not just to one particular λ, for example the "spectral band" from $\lambda = 0.4$ µm to $\lambda = 0.5$ µm. The spectral reflectance curve can be used to estimate the overall reflectance in such bands by calculating the mean of reflectance measurements in the respective ranges. Reflectance measurements can be carried out in a laboratory or in the field, in the latter case using a field spectrometer. Reflectance curves are typically collected for the optical part of the electromagnetic spectrum and large efforts are made to store collections of typical curves in "spectral libraries". The reflectance characteristics of some common land cover types are discussed in the following subsections.

reflectance measurements

Vegetation

The reflectance characteristics of vegetation depend on the properties of the leaves, including the orientation and structure of the leaf canopy. The amount of radiation reflected for a particular wavelength depends on leaf pigmentation, thickness and composition (cell structure), and on the amount of water in the leaf tissue. Figure 2.13 shows an ideal reflectance curve of healthy vegetation. In the visible portion of the spectrum, the reflection of the blue and red components of incident light is comparatively low, because these portions are absorbed by the plant (mainly by chlorophyll) for photosynthesis; the vegetation reflects relatively more green light. The reflectance in the NIR range is highest, but the amount depends on leaf development and cell structure. In the SWIR range, reflectance is mainly determined by the free water in the leaf tissue; more free water results in less reflectance. Wavelengths around 1.45 µm and 1.95 µm are, therefore, called water absorption bands. The plant may change colour when its leaves dry out, for instance at harvest time for a crop (e.g. to yellow). At this stage there is no photosynthesis, which causes reflectance in the red portion of the spectrum to become higher. Also, the leaves will dry out, resulting in a higher reflectance of SWIR radiation, whereas reflectance in the NIR range may decrease. As a result, optical remote sensing can provide information about the type of plant and also about its health.

Bare soil

Reflectance from bare soil depends on so many factors that it is difficult to give one typical soil reflectance curve. The main factors influencing reflectance are soil colour, moisture content, the presence of carbonates, and iron oxide content. Figure 2.14 gives the reflectance curves for the five main types of soil occurring in the U.S.A. Note the typical shapes of most of the curves, which are convex shape in the range 0.5–1.3 µm and dip at 1.45 µm and 1.95 µm. These dips correspond to water absorption bands and are caused by the presence of soil moisture. Iron-dominated soil (e) has quite a different reflectance curve since iron absorption dominates at longer wavelengths.

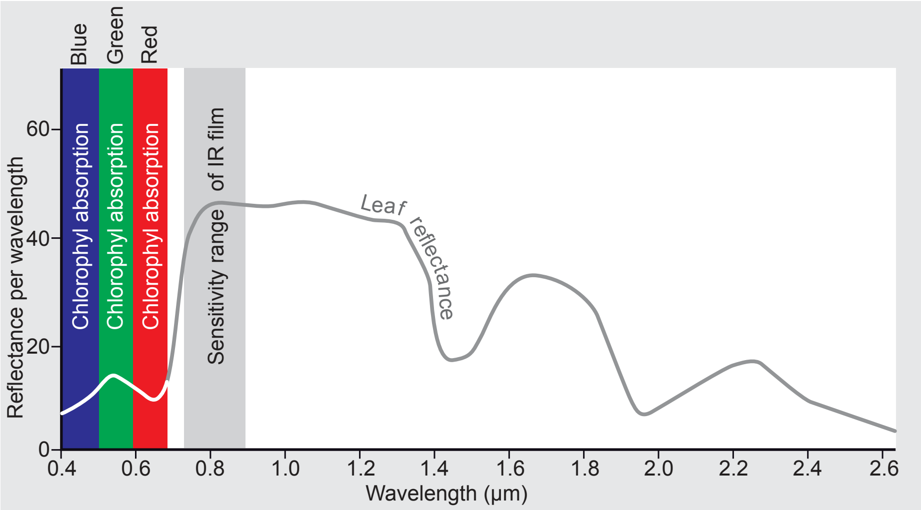

Figure 2.13
An idealized spectral reflectance curve of healthy vegetation.

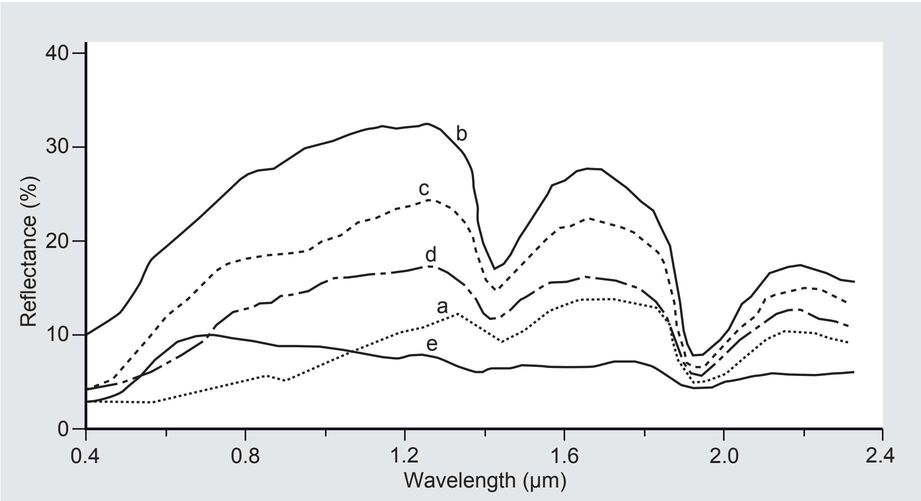

Figure 2.14
Spectral reflectance of five mineral soils: (a) organic dominated, (b) minimally altered, (c) iron altered, (d) organic affected and (e) iron dominated (from [71]).

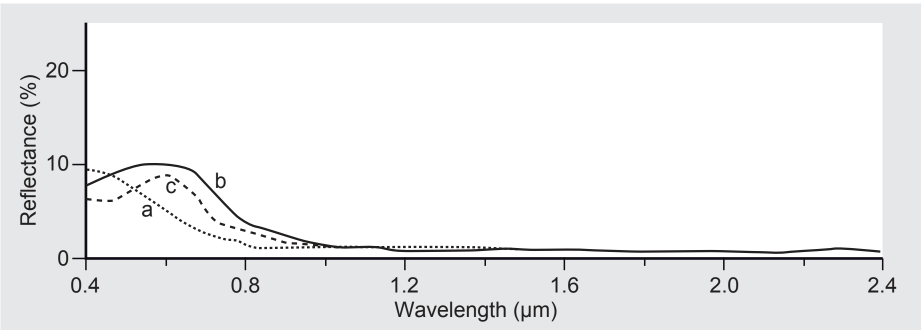

Figure 2.15
Typical effects of chlorophyll and sediments on water reflectance: (a) ocean water, (b) turbid water, (c) water with chlorophyll (from [71]).

Water

Compared to vegetation and soils, water has a lower reflectance. Vegetation may reflect up to 50% and soils up to 30–40%, while water reflects at most 10% of the incident radiation. Water reflects EM radiation in the visible range and a little in the NIR range. Beyond 1.2 μm, all radiation is absorbed. Spectral reflection curves for water of different compositions are given in Figure 2.15. Turbid (silt loaded) water has the highest reflectance. Water containing plants or algae has a pronounced reflectance peak for green light because of the chlorophyll present.

2.6 Sensing of EM radiation

The review of properties of EM radiation shows that different forms of radiation can provide us with different information about terrain-surface features and that different applications of Earth Observation are likely to benefit from sensing in different ranges of the EM spectrum. A geoinformatics engineer who wants to discriminate objects for topographic mapping will prefer to use an optical sensor operating in the visible range. An environmentalist who needs to monitor heat losses of a nuclear power plant will use a sensor that detects thermal emission. A geologist interested in surface roughness, because it indicates to him rock type, will rely on microwave sensing. Different demands combined with different technical solutions have resulted in a multitude of sensors. In this section we will classify various remote sensors and discuss their common features. Peculiarities will then be treated later in appropriate sections.

2.6.1 Sensing properties

A *remote sensor* is a device that detects EM radiation, quantifies it and, usually, records it in an analogue or digital form. A remote sensor may also transmit recorded data (to a receiving station on the ground). Many sensors used in Earth Observation detect reflected solar radiation. Others detect the radiation emitted by the Earth itself. There are, however, some obstacles to be overcome. The Sun does not always shine brightly and there are regions on the globe almost permanently under cloud cover. There are also regions that have seasons with very low Sun elevation, so that objects cast long shadows over long periods. Furthermore, at night there are only emissions—and perhaps moonlight. Sensors detecting reflected solar radiation are useless at night and face problems when dealing with unfavourable seasonal and weather conditions. Sensors detecting emitted terrestrial radiation do not directly depend on the Sun as a source of illumination; they can be operated any time. The Earth's emissions, we have learned, occurs only at longer wavelengths because of the relatively low surface temperature and because long EM waves do not hold much energy, which makes them more difficult to sense.

obstacles to sensing

Luckily we do not have to rely only on solar and terrestrial radiation. We can build instruments that emit EM radiation and then detect the radiation returning from the target object or surface. Such instruments are called *active sensors*, as opposed to passive ones, which measure reflected solar or terrestrial radiation (Figure 2.16). An example of an active sensor is a laser rangefinder, a device that can be bought for a few euros in any DIY store. Another very common active sensor is a camera with a flash unit (which will operate below certain levels of light). The same camera without the flash unit is a passive sensor. The main advantages of active sensors are that they can be operated day and night and have a controlled illuminating signal. They are often designed to work in an EM spectrum range that is less affected by the atmosphere and weather conditions. Laser and radar instruments are the most prominent active sensors for GDA.

active versus passive RS

Most remote sensors measure either the intensity or the phase of EM radiation. Some—like a simple laser rangefinder—only measure the elapsed time between sending a radiation signal and receiving it back. Radar sensors may measure both intensity and

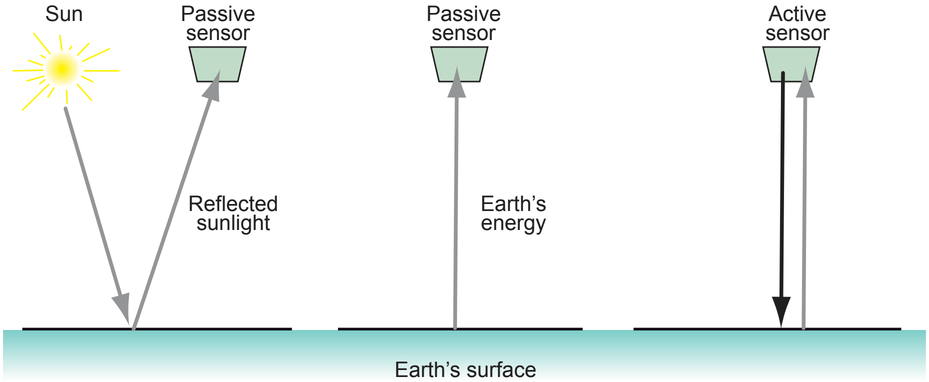

Figure 2.16
A remote sensor measures reflected or emitted radiation. An active sensor has its own source of radiation.

phase. Phase measuring sensors are used for precise ranging (distance measurement), e.g. by GPS "phase receivers" or continuous-wave laser scanners. The intensity of radiation can be measured from the photon energy striking the sensor's radiation-sensitive surface.

intensity or phase

By considering the following equation, you can relate the intensity measure of reflected radiation to Figure 2.6 and link the Figures 2.13 to 2.17. When sensing reflected light, radiance at the sensor is equal to the radiance at the Earth's surface attenuated by atmospheric absorption, plus the radiance of scattered light:

measuring radiance

$$L = \frac{\rho E \tau}{\pi} + \text{sky radiance} \tag{2.11}$$

where L is the total radiance at the sensor, E is the irradiance (the intensity of the incident solar radiation, attenuated by the atmosphere) at the Earth's surface, ρ is the terrain reflectance, and τ is the atmospheric transmittance. The radiance at the Earth's surface depends on the irradiance and the terrain surface reflectance. The irradiance, in turn, stems from direct sunlight and diffuse light, the latter caused by atmospheric scattering, particularly on a hazy days (see Figure 2.17). This indicates why you should study radiometric correction (Subsection 5.1.3 and Subsection 5.2.2), to enable you to make better inferences about surface features.

The radiance is observed for a *spectral band*, not for a single wavelength. A *spectral band* or wavelength band is an interval of the EM spectrum in which the average radiance is measured. Sensors such as a panchromatic camera, a radar sensor and a laser scanner only measure in one specific band, while a multispectral scanner or a digital camera measures in several spectral bands at the same time. Multispectral sensors have several *channels*, one for each spectral band. Figure 2.18 shows spectral reflectance curves, together with the spectral bands, of some popular satellite-based sensors. Sensing in several spectral bands simultaneously allows us to relate properties that show up well in specific spectral bands. For example, reflection characteristics in the spectral band 2 to 2.4 μm (as recorded by Landsat-5 TM channel 7) tell us something about the mineral composition of soil. The combined reflection characteristics in the red and NIR bands (from Landsat-5 TM channels 3 and 4) can tell us something about biomass and plant health.

spectral band

Landsat MSS (MultiSpectral Scanner), the first civil space-borne Earth Observation sensor, had sensing elements (detectors) for three rather broad spectral bands in the visible range of the spectrum, each with a width of 100 nm, and one broader band in the NIR range. A hyperspectral scanner uses detectors for many more, but narrower, bands, which may be as narrow as 20 nm, or even less. We say a hyperspectral sensor

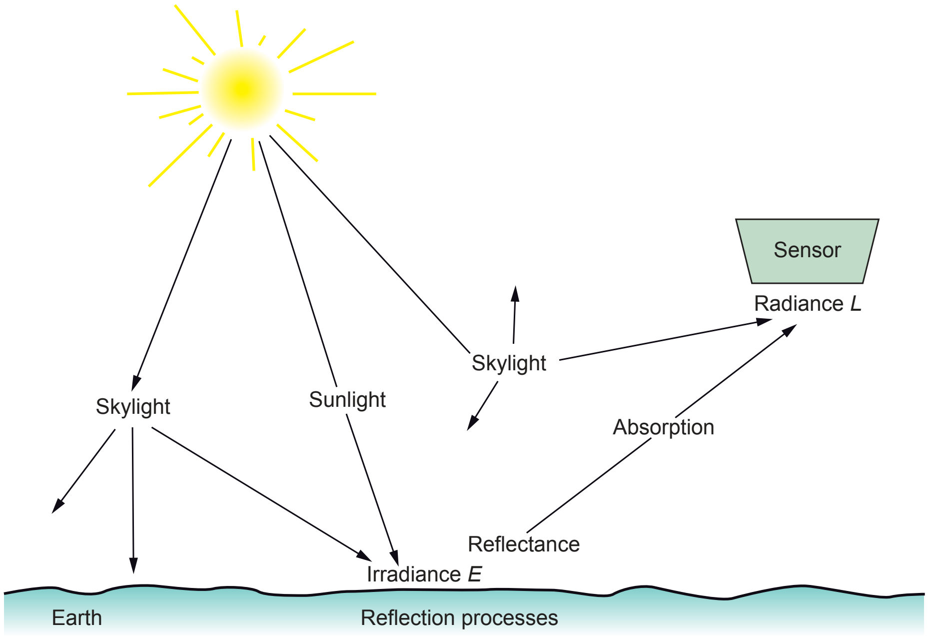

Figure 2.17
Radiance at the sensor, adapted from [67].

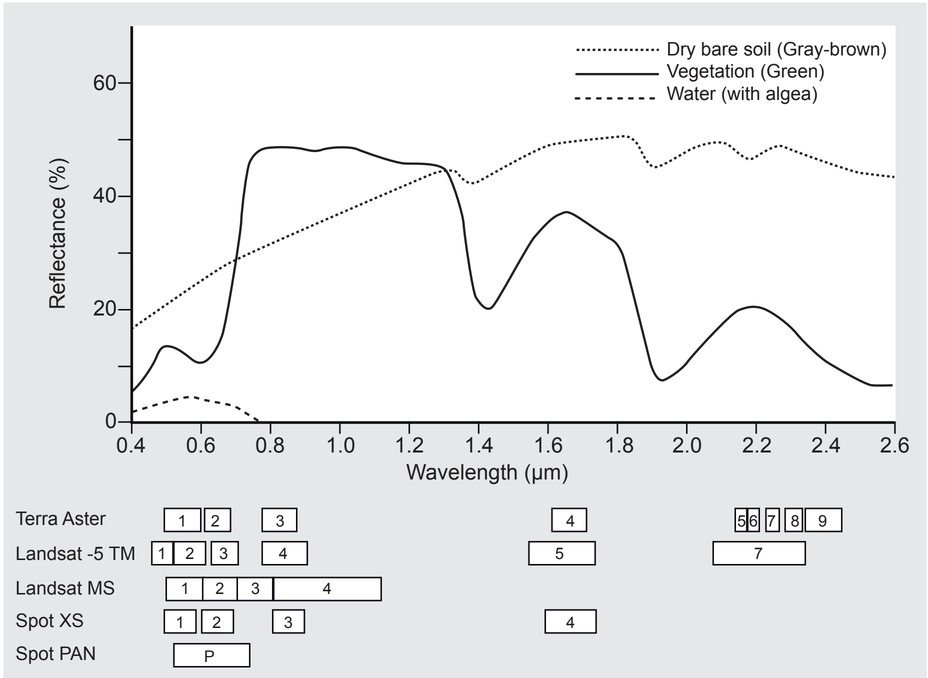

Figure 2.18
Spectral reflectance curves and spectral bands of some multispectral sensors.

spectral resolution

has a higher 'spectral resolution' than a multispectral one. A laser instrument can emit (and detect) almost monochrome radiation, with a wavelength band no wider than 10 nm. A camera loaded with panchromatic film or a space-borne electronic sensor with a panchromatic channel (such as SPOT PAN or WorldView-1) records the intensity of radiation of a broad spectral band covering the entire visible range of the EM spectrum. Panchromatic—which stands for "across all colours"—recording is comparable with the function of the 120 million rods of a human eye. They are brightness sensors and cannot sense colour.

photographic detector

In a camera loaded with panchromatic film (*black & white film*), the silver halide crystals of the light-sensitive emulsion detect radiation. The silver halide grains turn to silver metal when exposed to light, the more so the higher the intensity of the incident light. Each light ray from an object/scene triggers a chemical reaction of some particular grain. This way, variations in radiance within a scene are detected and an image of the scene is created at the time of exposure. The record obtained is only a latent image; the film has to be developed to turn it into a photograph.

AD conversion

Digital cameras and multispectral scanners are examples of sensors that use electronic detectors instead of photographic ones. An electronic detector (CCD, CMOS, photodiode, solid state detector, etc.) is made of semiconductor material. The detector accumulates a charge by converting the photons incident upon its surface to electrons. (It was Einstein who won the Nobel prize for discovering and explaining that there is an emission of electrons when a negatively charged plate of light-sensitive (semiconductor) material is subject to a stream of photons.) The electrons can then be made to flow as a current from the plate. So the charge can be converted to a voltage (electrical signal). The charge collected is proportional to the radiance at the detector (the amount of radiation "deposited" in the detector). In a process called A/D conversion, the electrical signal is *sampled* and quantified. The output is a digital number (DN), which is recorded. A DN is an integer within a fixed range. Older remote sensors used 8 bits for recording, which allows a differentiation of radiance into $2^8 = 256$ levels (i.e. DNs in the range 0 to 255). The recently launched (in 2007) WorldView-1 sensor records with a *radiometric resolution* of 11 bits $(2^{11} = 2048)$. ASTER records the visible spectral band using 8 bits and the thermal infrared band using 12 bits. A higher radiometric resolution requires more storage capacity but has the advantage of offering data with greater information content (see Figure 2.19).

radiometric resolution

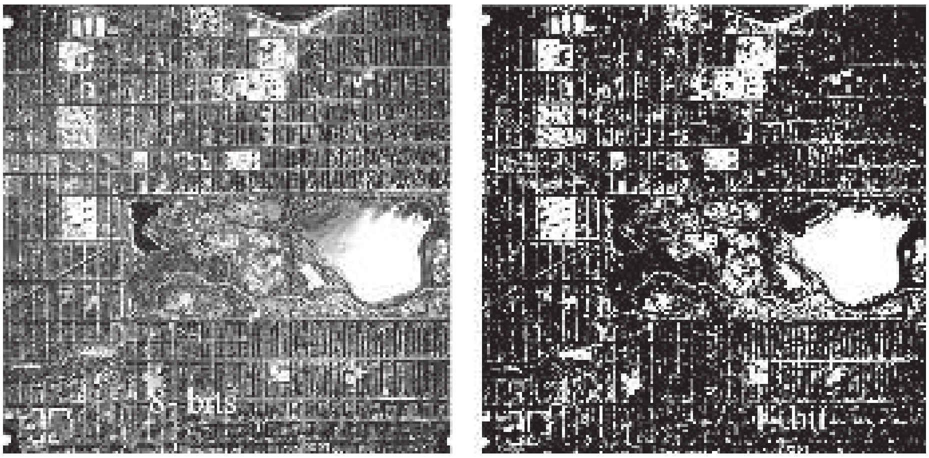

Figure 2.19
8-bit versus 11-bit radiometric resolution.

A digital panchromatic camera has an array of detectors instead of silver halide crystals suspended in gelatine on a polyester base of photographic film. Each detector (e.g. a CCD, which stands for charge-coupled device) is very small, in the order of 9 μm × 9 μm. Space-borne cameras use larger detectors than aerial cameras to ensure that enough photons are collected despite the great distances at which they operate from the Earth. At the moment of exposure, each detector yields one DN, so in total we obtain a data set that represents an image similar to the one created by "exciting" photographic material in a film camera.

imaging, spatial resolution

When arranging the DNs in a two-dimensional array, we can readily visualize them as grey values. We refer to the obtained "image" as a *digital image* and to a sensor producing digital images as an *imaging sensor*. The array of DNs represents an image in terms of discrete picture elements, called *pixels*. The value of a pixel—its DN—corresponds to the radiance of the light reflected from the small ground area viewed by the relevant detector. The smaller the detector, the smaller will be the area on the ground that corresponds to one pixel. The size of the "ground resolution cell" is often referred to as "pixel size on the ground". Early digital cameras for consumers had 2×10^6 CCDs per spectral band (named 2 megapixel cameras); today we can get for the same price a 10 megapixel camera. The latter has much smaller CCDs so that they can fit on the same board, with the consequence that an image can reveal much more detail; we would say the *spatial resolution* of the image is higher.

A digital camera for the consumer market does not record intensity values for a single (panchromatic) spectral band, but for three bands simultaneously, namely for red, green, and blue light, in order to obtain colour images. This is comparable with our eyes: we have three types of cones, one for each primary colour. The data set obtained for one shot taken with the camera (the *image file*) therefore contains three separate digital images (Figure 2.20). Multispectral sensors record in as many as 14 bands simultaneously (e.g. ASTER). For convenience, a single digital image is then often referred to as "band" and the total image file as a *multi-band image*.

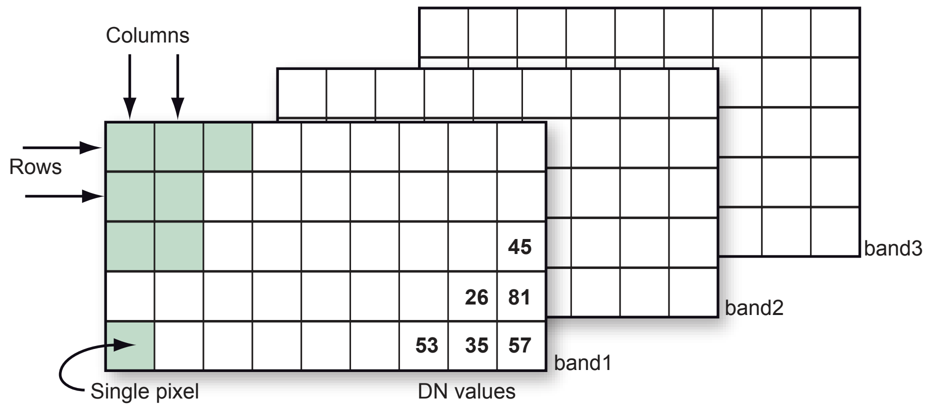

Figure 2.20
An image file comprises a digital image for each of the spectral bands of the sensor. The DN values for each band are stored in a row-column arrangement.

storage media

Various storage media are used for recording the huge amount of data produced by electronic detectors: solid state media (such as memory cards as used in consumer cameras), magnetic media (disk or tape) and optical discs (some video cameras); satellites usually have several recorders on board.

Light sensor systems often transmit data to ground receiving stations at night. Data can also be transmitted directly to a receiving station using satellite communication technology. Airborne sensors often use the hard disk of a laptop computer as a recording device. The huge amounts of data collected demand efficient data management systems. This issue will be examined in Section 8.4.

2.6.2 Classification of sensors

Remote sensors can be classified and labelled in different ways. According to whatever our prime interest in Earth Observation may be—geometric properties, spectral differences, or an intensity distribution of an object or scene—we can distinguish three salient types of sensors: altimeters, spectrometers, and radiometers.

Laser and radar altimeters are non-imaging sensors that provide information about the elevation of water and land surfaces.

altimeter

Thermal sensors, such as the channels 3 to 5 of NOAA's AVHRR or the channels 10 to 14 of Terra's ASTER, are called (imaging) radiometers. Radiometers measure radiance and typically sense in one broad spectral band or in only a few bands, but with high radiometric resolution. Panchromatic cameras and passive microwave sensors are other examples of radiometers. The spatial resolution depends on the wavelength band of the sensor. Panchromatic radiometers can have a very high spatial resolution, whereas microwave radiometers have a low spatial resolution because of the low levels of energy inherent in this spectral range. Scatterometers are non-imaging radiometers. Radiometers are used for a wide range of applications: for example, detecting forest/bush/coal fires; determining soil moisture and plant response; monitoring ecosystem dynamics; and analysing energy balance across land and sea surfaces.

radiometer

Spectrometers measure radiance in many (usually about 100 or 200) narrow, contiguous spectral bands and therefore have a high spectral resolution. Their spatial resolution is moderate to low. The prime use of imaging spectrometers is to identify surface materials—from the mineral composition of soils, to concentrations of suspended matter in surface water and chlorophyll content. There are also androgynous sensors: spectro-radiometers, imaging laser scanners, and Fourier spectrometers, for example.

spectrometers

We can also group the multitude of remote sensors used for GDA according to the spectral domains in which they operate (Figure 2.21). The following list gives a short description of each group and refers to the section in which they are treated in more detail.

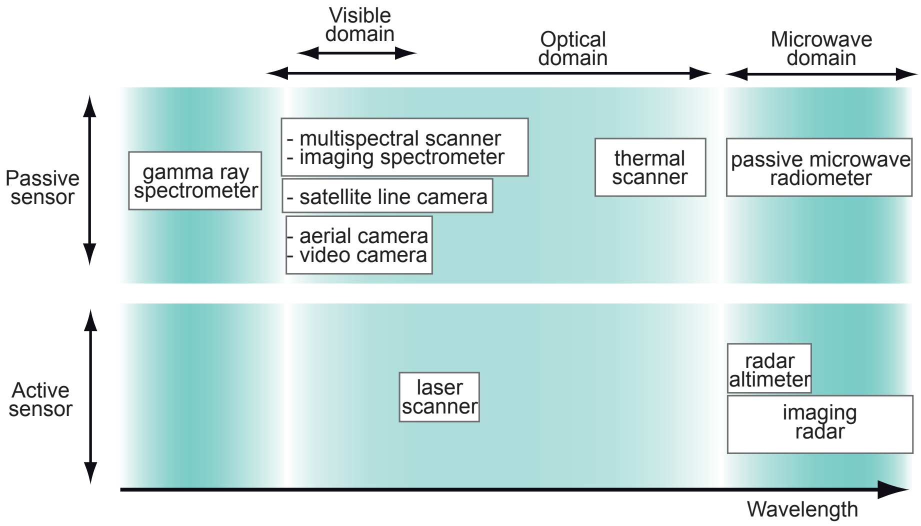

Figure 2.21
Overview of the sensors that are described in this book.

- gamma ray spectrometers are mainly used in mineral exploration.

gamma ray sensors

- aerial film cameras have been the remote sensing workhorse for decades. Today, they are used primarily for large-scale topographic mapping, cadastral mapping, and orthophoto production for urban planning, to mention a few examples; they are discussed in Section 4.6.

film cameras

- digital aerial cameras are not conquering the market as quickly as digital cameras did on the consumer market. These cameras use CCD arrays instead of film; they are treated together with optical scanners in Section 4.1. Line cameras operated from satellites have very similar properties.

digital cameras

video cameras

- digital video cameras are not only used to record movies. They are also used in aerial Earth Observation to provide low cost (and low resolution) images for mainly qualitative purposes, for instance to provide visual information about an area covered by "blind" airborne laser scanner data. Handling images from video cameras is similar to dealing with images from digital "still" cameras; this is not explicitly discussed any further in this book.

multispectral scanners

- multispectral scanners are mostly operated from satellites and other space vehicles. The essential difference between multispectral scanners and satellite line cameras is the imaging/optical system employed: multispectral scanners use a moving mirror to "scan" a line (i.e. a narrow strip on the ground) and a single detector instead of recording intensity values of an entire line at one instant by an array of detectors as for line cameras. Multispectral scanners are treated in Section 4.1. Figure 2.22 shows an image obtained by combining the images of Landsat TM channels 4, 5 and 7, which are displayed in red, green and blue, respectively. Section 5.1 explains how to produce such a "false colour" image.

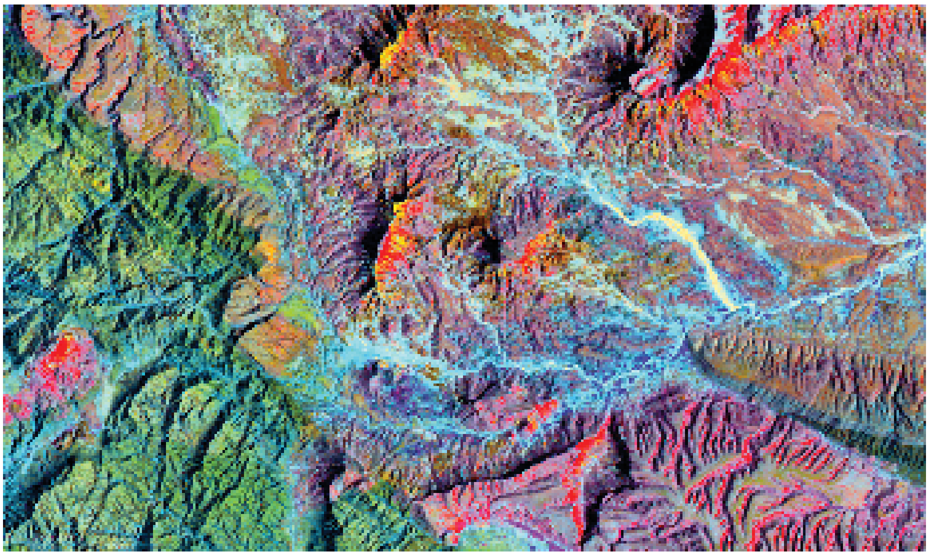

Figure 2.22
Landsat-5 TM false colour composite of an area of 30 km × 17 km.

imaging spectrometers

- hyperspectral scanners are imaging spectrometers with a scanning mirror; they are treated in detail in Section 4.3.

thermal scanners

- thermal scanners are placed here in the optical domain purely for the sake of convenience. They exist as special instruments and as a component of multispectral radiometers; they are included in Section 4.2. Thermal scanners provide us with data that can be directly related to object temperature. Figure 2.23 is an example of a thermal image acquired by an airborne thermal scanner at night.

microwave radiometers

- passive microwave radiometers detect emitted radiation of the Earth's surface in the 10 to 1000 mm wavelength range. These radiometers are mainly used in mineral exploration, for monitoring soil-moisture changes, and for snow and ice detection. Microwave radiometers are not discussed further in this book.

- laser scanners are the scanning variant of laser rangefinders and altimeters (as on ICESat). Laser scanners measure the distance from the laser instrument to many points of the target in "no time" (e.g. 150,000 points in one second). Laser

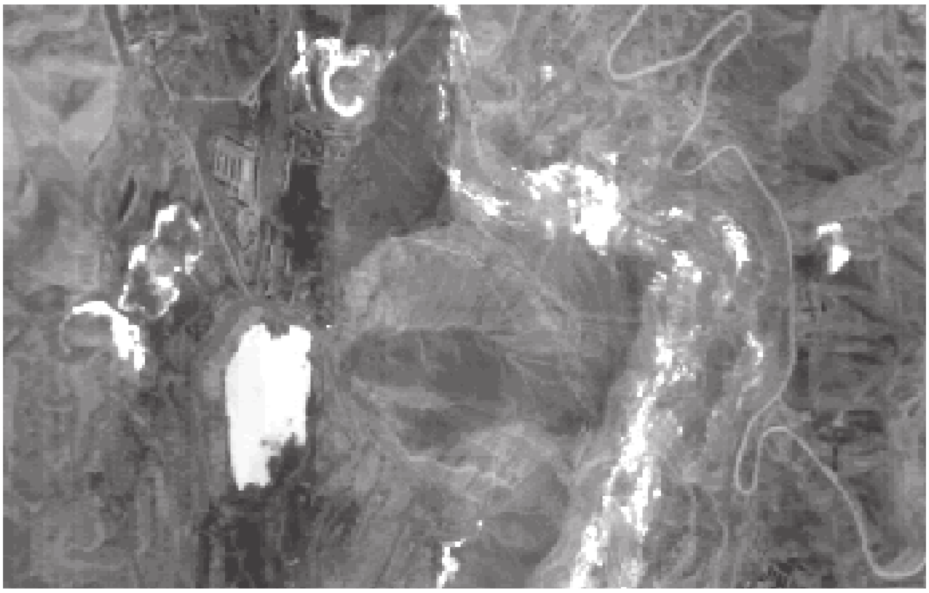

Figure 2.23
'Thermal image' at night of a coal mining area affected by underground coal fires. Darker tones represent colder surfaces, while lighter tones represent warmer areas. Most of the warm spots are due to coal fires, except for the large white patch, which is a lake; at that time of the night, apparently the temperature of the water was higher than the temperature of the land. On the ground, the area depicted is approximately 4 km across.

ranging is often referred to as LIDAR (LIght Detection And Ranging). The prime application of airborne laser scanning (ALS) is for creating high resolution digital surface models and digital terrain models (see Section 5.3). We can also create a digital terrain model (DTM) from photographs and similar panchromatic images. However, because of the properties of laser radiation, ALS has important advantages in areas of dense vegetation and for sandy deserts and coastal areas. Surface modelling is of interest for many applications, such as, for example, biomass estimation of forests, volume calculations for open-pit mining (see Figure 2.24), flood plain mapping, and 3D modelling of cities. Laser scanning is dealt with in more detail in Section 4.5.

Figure 2.24
Pictorial representation of a digital surface model of the Sint Pietersberg open-pit mine in the Netherlands. The size of the pit is roughly 2 km $\times$ 1 km. The terraced rim of the pit is clearly visible. The black strip near the bottom of the image is the River Meuse. Courtesy Survey Department, Rijkswaterstaat.

- imaging radar (RAdio Detection And Ranging) operates in the spectral range

10–1000 mm. Radar instruments are active sensors and because of the range of wavelengths used they can provide data day and night, under all weather conditions. Radar waves can penetrate clouds; only heavy rainfall affects imaging to some degree. One of its applications is, therefore, the mapping of areas that are subject to permanent cloud cover. Figure 2.25 shows an example of a SAR (Synthetic Aperture Radar) image from the ERS-1 satellite. Radar data from the air or space can also be used to create surface models. Radar imaging has a peculiar geometry and processing raw radar data is not simple. Radar is treated in Section 4.4.

imaging radar

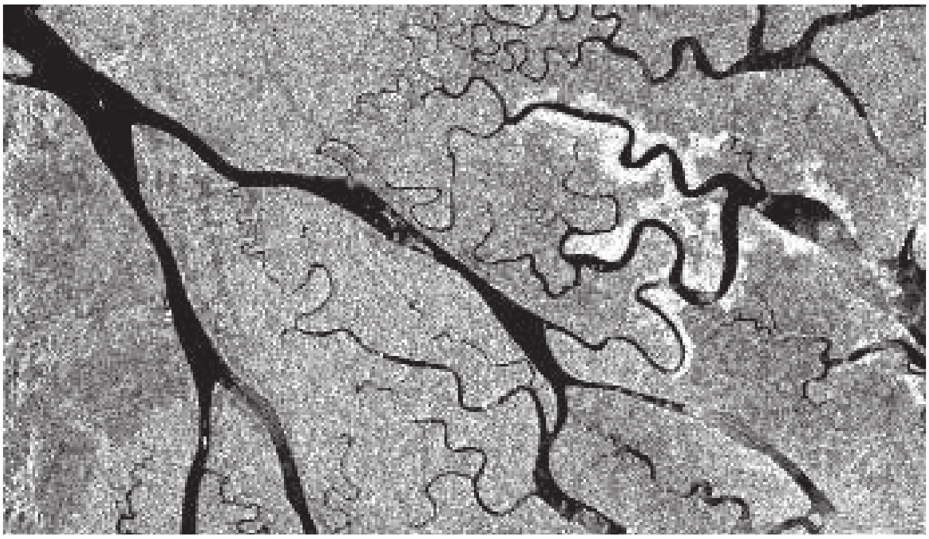

Figure 2.25
An ERS-1 SAR image of the Mahakam Delta, Kalimantan. The image shows different types of land cover. The river is black. The darker patch of land on the the left is inland tropical rainforest. The rest is a mixed forest of Nipa palm and mangrove on the delta. The right half of the image shows light patches, where the forest has been partly cleared. The image covers an area on the ground of 30 km x 15 km.

radar altimeters

- radar altimeters are used to measure elevation profiles of the Earth's surface that is parallel to the receiving satellite's orbit. Radar altimeters operate in the 10–60 mm range and allow us to calculate elevation with an accuracy of 20–50 mm. Radar altimeters are useful for measuring relatively smooth surfaces.

sonar

- for the sake of completeness, sonar, another active sensor, is included here. Sonar, which stands for SOund NAvigation Ranging, is used, for example, for mapping river beds and sea floors, and for detecting obstacles underwater. Sonar works by emitting a small burst of sound from a ship. The sound is reflected off the bottom of the body of water. The time taken for the reflected pulse to be received corresponds to the depth of the water. More advanced systems also record the intensity of the return signal, thus giving information about the material on the sea floor. In its simplest form, sonar "looks" vertically and is operated very much like a radar altimeter. The body of water will be traversed in paths resembling a grid; not every point below the water surface will be monitored. The distance between data points depends on the ship's speed, the frequency of the measurements, and the distance between the adjacent paths.

  One of the most accurate systems for imaging large areas of the ocean floor is side-scan sonar. It is an imaging system that works in a way that is somewhat similar to side-looking airborne radar (see Section 4.4). The images produced by side-scan sonar systems are highly accurate and can be used to delineate even very small (< 1 cm) objects. From sonar data, we can produce contour maps of sea floors and other water bodies, which can be used, for example, for navigation and water-discharge studies.