The methods of satellite oceanography

2.1 OCEAN REMOTE-SENSING TECHNIQUES—A SUMMARY

This book is primarily about the oceanographic applications of remote sensing, and is written to complement the detailed descriptions and discussion of the techniques of satellite oceanography published in the companion volume MTOFS (Robinson, 2004). However, it would be unfortunate if a reader new to the whole field were left in complete ignorance of how the data presented in the rest of this book have been acquired. This chapter is therefore provided to give a very basic introduction to the methods of ocean remote sensing. It aims to summarize the essentials of the subject and to provide the minimum knowledge that a university graduate in oceanography ought to have, without going into all the detail that someone working in the fields of satellite oceanography research or applications would need. Thus although it only skims the surface, the reader who needs further information to make an explanation clear, or who is stimulated to find out more about a particular technique, can be confident of being able to do so by consulting MTOFS (Robinson, 2004) where a much fuller set of relevant references will also be found. On the other hand, the reader who is already familiar with Robinson (2004) can safely skip this chapter.

Figure 2.1 illustrates schematically what is involved in measuring properties of the ocean using a sensor that is typically hundreds or thousands of kilometers from the sea surface. An electromagnetic signal of a particular kind leaves the sea carrying information about one of the primary observable quantities which are the color, the radiant temperature, the roughness, and the height of the sea. This signal must pass through the atmosphere where it may be changed, and where noise may be added to it, before it is received by the sensor which detects particular properties of the radiation and converts each measurement into a digital signal to be coded and sent to the ground. The sensor geometry restricts each individual observation to a particular instantaneous field of view (IFOV). In order to convert the numbers
received at the ground station into scientific measurements of useful precision and quantifiable accuracy, the remote-sensing process represented in the left-hand side of Figure 2.1 must be inverted digitally using the knowledge and information identified on the right-hand side.

It is their acquisition from a unique vantage point in space which gives satellite data their special character and so in Section 2.2 our brief summary of methods takes a look at the way image datasets are acquired, identifies the different satellite orbits available for remote sensing, and considers how both of these factors affect the space-time sampling capacity of the datasets. Then Section 2.3 gives an overview of the generic data-processing tasks that are implied in the right-hand side of Figure 2.1. These consist of operations that must be performed on the raw data received from the satellite in order to turn them into estimates or measurements of ocean parameters, with quantifiable accuracy, suitable for use in the scientific analysis or operational applications described in the rest of this book.

Section 2.4 introduces the diverse techniques of satellite oceanography, which use distinct parts of the electromagnetic spectrum and different aspects of radiation to measure particular properties of the ocean. Although several pages each are devoted to the different methods, ocean color, thermal infrared temperature detection, passive-microwave radiometry, altimetry, and oblique-viewing radars, these are very abbreviated descriptions of what are today extensive subjects each worthy of a book for themselves. A summary of the more important satellites and sensors used
for ocean remote sensing is provided by Section 2.5, and Section 2.6 concludes the chapter with guidance to readers on how to browse, access, and manipulate some of the wide variety of satellite ocean data available through the Internet.

2.2 THE UNIQUE SAMPLING CAPABILITIES OF SENSORS ON SATELLITES

The use of Earth-orbiting satellites as platforms for ocean-viewing sensors offers a number of unique advantages such as the opportunity to achieve wide synoptic coverage at fine spatial detail, and repeated regular sampling to produce time series several years long. It is these capabilities that distinguish satellite remote sensing from all other oceanographic observing techniques. The capacity for synoptic imaging depends primarily on the spatial sampling characteristics of the sensor, which are ultimately limited by detector sensitivity and the data flow capacity of the telecommunications system between the satellite and ground stations. Another set of limitations follow from the unavoidable constraints imposed by the physical laws of satellite orbital dynamics. Ultimately the sampling characteristics of different satellite oceanography methods depends on the sensor–platform combination. This section provides an outline of the important issues, but a more detailed discussion of sampling by remote sensors can be found in chapters 3 and 4 of MTOFS (Robinson, 2004).

2.2.1 Creating image-like data fields from point samples

What makes satellite data so useful and interesting for many users is their unique capability for dense two-dimensional spatial sampling which enables images to be formed corresponding to the surface distribution of the measured variable. But unlike the “snapshot” pictures we obtain from cameras, remotely sensed image data fields consist of millions of individual scientific measurements built up over a short length of time from a regular sampling pattern over the ground. Typically just one sensor is used, which ensures consistency of sensitivity for all the samples making up the image dataset. Those remote-sensing instruments that use an array of detectors must ensure uniform intercalibration of all elements.

The sea or ground area observed by a single detector is limited to its instantaneous field of view (IFOV) which is defined by a given directional spread relative to the pointing direction. Two-dimensional sampling to cover the sea surface is achieved by utilizing any relative motion between the platform and the ground and by pointing the sensor in a systematic sampling pattern. The instantaneously acquired measurement of an ocean property would be a single value representing the average property over the region defined by the intersection of the IFOV with the ground. However, because every sensor requires a finite time to record a measurement, during which the pointing of the sensor moves a finite distance over the ground, the effective “footprint” of each measurement must be somewhat larger.
than the ground IFOV (as illustrated in Figure 2.2). The footprint determines the spatial resolution of the sensor.

Some sensors simply make downward-looking observations at periodic intervals while the satellite moves over the ground, to give an average value of radiation or other property from an area of the Earth surface, typically centered at the nadir point (the point on the Earth immediately below the satellite). A nadir-viewing altimeter is an example of this type of sensor. The only way to produce image-like datasets from such sensors is to wait until the satellite track has covered the ground with sufficient density to enable the variable to be smoothly mapped from the available point measurements, but this may take many days to achieve.

To obtain a truly near-instantaneous image requires a sensor that explicitly scans sideways across the satellite track direction. Figure 2.3 illustrates a typical arrangement where scan lines are perpendicular to the satellite track. Normally it is arranged for the sensor to scan a complete line and return to start the next in the time it takes for the satellite subpoint on the ground to travel a distance equal to the footprint size in the along-track direction. Thus the scan line spacing matches the sensor spatial resolution and adjacent scan lines are contiguous as shown. Along the scan lines, the sensor is arranged to take one sample in the time it takes for the pointing to swing through an angle equivalent to the IFOV, thus matching the sample spacing to the sensor resolution in the scan direction also. In this way a wide swath of ground is imaged, normally centered on the satellite subtrack.

The detailed scanning mechanism varies from sensor to sensor. Note that in general the footprint size increases and changes shape slightly towards the extremity of wide swath scan lines. In some cases the scan geometry may not be rectangular but curved, as when a conically scanning mirror is used instead of a rectangular scanning mirror. In the case of a geostationary satellite (see Section 2.2.2) the whole platform rotates about a north–south axis to achieve scanning parallel to lines of latitude. At the same time the sensor’s field of view is rotated north–south to point at different latitudes for each satellite rotation, thus achieving a coverage of the whole face of the globe as visible from that location in space. For microwave devices, the scanning...
may be achieved by electronic (beam steering) or radar signal-processing methods without the need for mechanically moving reflectors. In this case it may be possible to reduce the geometric distortion inherent in mechanical scanning. A fuller discussion of scanning and imaging methods is given in section 4.1 of *MTOFS* (Robinson, 2004).

It is worth emphasizing the fact that a remote sensor integrates the incoming radiation over the IFOV and so the estimates it makes of ocean properties correspond to averages over the measurement footprint. In a well-designed scanning system in which the sea surface is covered by contiguous but not overlapping footprints, the resulting set of measurements are directly comparable with the way a two-dimensional model describes the sea, representing ocean variables as averages within each cell of a rectangular grid. For many applications of satellite data this provides a distinct advantage compared with conventional *in situ* instruments that make single-point measurements in the sea. To compare such *in situ* data with models requires measurements that are representative of the whole cell, difficult to achieve if subcell-scale variability is large unless many different measurements can be spatially averaged within the cell. Remote-sensing observation avoids this point-sampling problem, although it is encountered in a different way during the procedure of validating satellite data by comparing them with *in situ* measurements.

### 2.2.2 Satellite orbits and how they constrain remote sensing

Earth-orbiting satellites are constrained by forces due to gravitation and inertia. Based on Newtonian dynamics, the period, $T$, for a satellite to travel once round
a circular orbit at distance \( r \) above the center of the Earth is:

\[
T = 2\pi \sqrt{\frac{r^3}{GM}},
\]

where \( G \) is the constant of gravitation; \( M \) is the mass of the Earth; and

\[
GM = 3.98603 \times 10^{14} \text{ m}^3 \text{ s}^{-2}.\]

In terms of the satellite height \( h \) above the ground and the Earth radius \( R \) (about 6,378 km) \( r = R + h \) and so

\[
T = 2\pi \sqrt{\frac{(R + h)^3}{GM}}.\] (2.2)

There are just two basic types of orbit useful for ocean remote sensing, geostationary and near-polar, as illustrated in Figure 2.4. The geostationary orbit, at a height of about 35,785 km, has a period of one sidereal day (\( \approx 23.93 \) h) which is the time taken for the Earth to rotate through 360°. Placed over the Equator, the satellite flies west–east at the same rate as the Earth’s rotation, so it always remains fixed at the same place in the sky relative to objects on the ground, allowing it to view the ground at any sampling frequency. Being fixed it can see only that part of the world within its horizon and cannot usefully view much beyond about 7,000 km in any direction measured from the satellite nadir point on the Equator, at the longitude of the satellite.

In a near-polar orbit the satellite flies at a much lower altitude, typically between about 700 km and 1,350 km, for which the orbital period is about 100 min (Equation 2.2). It thus completes between 14 and 15 orbits a day, during which the Earth rotates once, so the satellite marks out a ground track crossing about 14 times northeast–southwest (descending tracks) and the same number of southeast–northwest ascending tracks. The tracks are distributed evenly around the globe, with successive orbits following a track about 24° of longitude to the west of the previous orbit as shown in Figure 2.5. If the satellite in orbit returned to its starting point exactly after one day then it would go on repeating the same 14 or 15 orbit

![Figure 2.4](image-url)

Figure 2.4. The two types of orbit used for Earth-observing satellites, drawn approximately to scale. The geostationary orbit is about 36,000 km above the Earth. The near-polar orbit is typically between 700 km and 1,000 km above the Earth surface.
tracks and never visit the spaces between them. Instead the orbit is normally planned so that it takes a longer time, typically between 3 days and 35 days known as the orbit repeat period, before it starts to cover its earlier track exactly. The longer the orbit repeat period the greater the number of different orbit tracks over the Earth surface that are completed within the cycle, and so the smaller the spaces between the tracks.

Most low, near-polar orbits used for Earth observation satellites are arranged to be Sun-synchronous. By choosing an inclination that is slightly greater than 90° (i.e., their path does not quite reach the poles) the orbit plane can be constrained to precess at a rate of once per year relative to the stars. This locks the overpasses to the position of the Sun and means that every orbit always crosses the Equator at the same local solar time. For most ocean-observing sensors this is very convenient, since it ensures that the longitudinal position of the Sun does not change from one sample to the next, even though the solar latitude inevitably changes with the annual cycle. However, for altimetry a Sun-synchronous orbit is to be avoided since it aliases the solar semidiurnal tidal constituent, because the solar tidal phase will be exactly the same every time the satellite revisits the same location on the sea surface.

More information about orbits can be found in section 3.2 of MTOFS. Section 11.6.3 of MTOFS explains more about tidal aliasing in altimetry.
2.2.3 The space-time sampling capabilities of satellite sensors

While a sensor on a geostationary platform is unable to see beyond its restricted horizon it does stay in the same place all the time and is therefore potentially capable of high-frequency sampling. However, it takes time to scan in spatial detail the full Earth disk field of view and its great height above the ground also makes it difficult to achieve fine spatial resolution. Geostationary sensors typically offer a revisit interval of less than 30 min and spatial resolution of 3 km to 5 km. This gives them the highest frequency time sampling of all satellite sensors, but relatively poor spatial resolution.

In contrast, a scanning sensor on a polar platform can potentially cover the whole Earth in a single day, as long as the swath is at least about 2,700 km wide, which is the distance at the Equator between the ground track of successive polar orbits (see Figure 2.6a). In this case each point on the Earth surface will be viewed at least once from a descending track and once from an ascending track. For a Sun-synchronous orbit, if the ascending track is in local daytime the descending track will be at night, or vice versa. An even wider scan permits more samples per day as swaths from successive orbits overlap at the Equator, while at higher latitudes overlapping occurs for much narrower swaths. Nonetheless, except in polar regions, the regular sampling interval for a single polar orbiter is never less than several hours.

For much narrower swaths as illustrated in Figure 2.6b, normally associated with fine-resolution imaging sensors, the time between successive views of the same location depends on the orbit repeat period. If this is just a few days then the sensor revisit interval will be the same as the repeat period. However, for too short a repeat period the spacing between the complete set of ground tracks will still be wider than

![Figure 2.6](image_url)

**Figure 2.6.** A single day’s coverage over Europe by (a) a wide (>2,000 km) swath sensor and (b) a narrow (<200 km) swath sensor. In each case the two tracks represent the typical spacing between successive orbits of a polar orbiter at an altitude of about 1,000 km.
the narrow swath and so the sensor will miss many parts of the Earth surface altogether. Global coverage by a sensor whose swath is about 200 km would take about 15 days to accomplish, and so an orbit repeat cycle of at least 15 days is required for this to be achievable. Sensors with intermediate width swaths (e.g., 500–1,000 km) will still take several days to obtain total coverage of the equatorial latitudes, but the revisit interval is likely to be shorter than the orbit repeat period, the more so at higher latitudes.

For nonscanning instruments such as altimeters that sample only along the ground track, the longer the orbit repeat period is, the greater the spatial coverage and the finer the sampling grid that can be achieved. However, it takes an extended period of time to build up the resulting map of data. Whereas for scanning sensors the spatial-sampling grid depends on the scanner design, and is usually matched to the sensor spatial resolution, for a nonscanning sensor it is the orbit pattern that dictates the spatial-sampling grid.

For scanning and nonscanning sensors alike, there is evidently a well-defined tradeoff between spatial and temporal-sampling capability, which is discussed in more detail by chapter 4 of Robinson (2004). It is important to appreciate these fundamental constraints when designing an ocean-observing system for operational purposes. For example, the only way to ensure that even a wide-swath sensor can sample every 6 hours from a near-polar orbit is to fly sensors on two satellites. Ideally a combination of spatial and temporal resolution should be selected in order that important phenomena can be adequately sampled. For example, if mesoscale eddies are to be monitored then the spacing between orbit tracks should not be wider than their variability lengthscale, nor should the repeat cycle be longer than the characteristic lifetime of an eddy. Otherwise some eddies may be missed altogether.

Every individual remote-sensing instrument has its particular space-time sampling capabilities, depending on both the sensor itself and the platform on which it is placed. These can be expressed in a graphical form, as in Figure 2.7 which represents the sampling characteristics as a defined rectangular area in a two-dimensional space-time field, for the generic sensor types discussed in the foregoing paragraphs. The vertical axis represents a logarithmic lengthscale and the horizontal axis a logarithmic timescale. The lower boundary of the region allocated to a particular sensor represents the smallest spatial scale that can be detected by the sensor (i.e., its spatial resolution). Similarly the left-hand boundary represents the shortest time interval over which variations in the ocean can be detected (i.e., the temporal-sampling resolution). The bottom left-hand corner therefore represents the best spatiotemporal resolution that is possible using that sensor. Note that a clear field of view with no obstructions is assumed. In the case of a sensor type that can view only under clear-sky conditions, cloud cover would degrade the temporal resolution and have the effect of shifting the left-hand boundary farther to the right.

The top boundary for each sensor’s sampling capability box corresponds to the largest extent of spatial coverage that can be obtained for a near-instantaneous view. This is the size of individual images collected at a particular time, and is generally defined by the swath width, although it could be larger in the along-track direction.
Of course this does not represent the full coverage of the sensor which, for sensors on polar platforms, is global. The right-hand boundary represents the timespan of available data, and depends on the lifetime over which useful data have been collected. In the examples illustrated a continuous series of at least 10 years is assumed.

The region enclosed within the boundaries represents the space-time sampling space for that sensor. The height of the box indicates the range of lengthscales that can be resolved, and the width the range of timescales. Diagrams like this are useful not only for comparing different sensors, but also for relating the sensor sampling capability to the requirements of a particular application; specifically in matching it to the space-time variability scales of the ocean phenomenon to be observed, as discussed further in section 4.5 of *MTOFS*.

### 2.3 GENERIC DATA-PROCESSING TASKS

It is important that those who make use of satellite-derived data products should be aware of the calibrations, corrections, analyses, and resampling that may have been applied to the products before they received them, since these processes have impacts that are relevant for their oceanographic interpretation and application. This section therefore provides a short overview of these tasks, which correspond to the informa-
tion “unpacking” that is necessary (see Figure 2.1) if the raw data acquired from a satellite are to be turned into useful quantitative information about an ocean variable, property, or parameter. Figure 2.8 summarizes the sequence of tasks and indicates the different “levels” of data products that correspond to each stage of data processing. The data product levels are defined in Table 2.1. Some tasks specific to particular satellite oceanography methods will also be described in Section 2.4. A more complete discussion of generic processing tasks can be found in MTOFS (sections 3.4.2 and 5.2).

2.3.1 Sensor calibration

The sensor calibration stage of data processing should convert the raw data received from the sensor into an estimate of the electromagnetic property which the sensor is intended to detect, such as the radiance entering the sensor in a given waveband. It
must account for all factors affecting information flow from the environmental signal entering the sensor to the receipt of the digital data signal at the ground. The first requirement is for a sensor calibration model that will determine the radiance, phase, or other property, of radiation entering the satellite sensor, given the value transmitted by the satellite and received at the ground station. This model is needed to invert two processes: the conversion of received radiation into an electrical response (typically a voltage amplitude or frequency) and the conversion of that response into a digital value. The accuracy of the calibration model must be defined in order to characterize the errors introduced in the transducer and the digitizer. It can be assumed that the transmission of the digital signal is error-free, if the signal is received at all.

For good sensor calibration to be achieved, all the sensor components must be carefully calibrated and characterized before launch. However, for many sensor types, this by itself is not enough to ensure confident in-flight calibration. Radiometers, for example, normally require the use of calibration radiant sources to allow calibration to be regularly updated. Since the sources as well as the sensors may degrade, elaborate checking systems use other calibration targets, such as the Moon for visible wavelength light. The slow degradation of a sensor calibration over its lifetime may only be fully defined a considerable time after the data were first acquired. Consequently the historic archive of data from a particular sensor may be reprocessed and the derived products reissued, up to several years after they were first acquired. Users of satellite data need to be aware of this possibility.

<table>
<thead>
<tr>
<th>Level</th>
<th>Description of product</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>Raw data received from the satellite, in standard binary form.</td>
</tr>
<tr>
<td>1</td>
<td>Image data in sensor co-ordinates. Individual calibrated channels, of measurements made at the satellite.</td>
</tr>
<tr>
<td>1.5 (or 1a)</td>
<td>In special cases, level 1 data with atmospheric correction applied.</td>
</tr>
<tr>
<td>2</td>
<td>Atmospherically corrected and calibrated image dataset of water-leaving radiance or derived oceanic variable. Geolocated, but normally presented in image co-ordinates.</td>
</tr>
<tr>
<td>3</td>
<td>Composite images of derived ocean variable resampled onto standard map base. Averaged in space and time from several overpasses of level 2 data. Derived from a single sensor. May contain gaps.</td>
</tr>
<tr>
<td>4</td>
<td>Image representing an ocean variable averaged within each cell of a space-time grid, for which gaps at level 3 have been filled by data analysis including interpolation. The analysis may merge several level 2 and/or level 3 datasets from a number of sensors and may also use <em>in situ</em> observations or model output.</td>
</tr>
</tbody>
</table>
If the data are made publicly available after sensor calibration has been applied then they are referred to as a level 1 dataset. Not all agencies make data of this type available, since it does not serve a very practical purpose for data users, although it is useful for scientific analysis into the underlying remote-sensing processes.

### 2.3.2 Atmospheric correction

A serious disadvantage for oceanographers of remote sensing from Earth observation satellites is that we must look through another medium—the atmosphere—to see the ocean. The atmosphere is opaque to electromagnetic radiation at many wavelengths, and there are only certain wavelength windows through which radiation may be fully or partially transmitted. Atmospheric gas molecules themselves may absorb or scatter radiation, and in addition water vapor, aerosols, and suspended particles of dust will do the same. If water droplets are present in the form of clouds, they may completely change the transmission properties of the atmosphere.

A major task for the analysis of satellite data is to take into account the effect of the atmosphere. At its worst this requires the detection of phenomena which render the data unusable (such as cloud for visible and infrared sensors or heavy rain for microwave radiometers). Otherwise it involves estimation of the effect that the atmosphere has had on measured electromagnetic radiation, between leaving the sea and reaching the sensor.

#### Cloud detection

Sensors using the visible and infrared parts of the spectrum cannot view the ocean through cloud. When handling data from such sensors, the next processing task after applying sensor calibration is to detect which pixels are obscured by cloud and which are clear. The purpose of cloud detection operations is to flag pixels as cloudy or as cloud-free. In some cases the flagging can be more sophisticated, indicating probabilities of cloud corruption, or indicating which of several different cloud detection tests has been positive. A comparable process is also adopted for analyzing microwave radiometer data. Although clouds are transparent to microwaves and do not present a problem, cells of very heavy precipitation in the atmosphere are opaque to microwaves and need to be detected.

It must be appreciated that clouds can never be “removed” from image datasets to leave a cloud-free field. To do so would imply that we knew what the ocean conditions were like beneath the cloud, but that is precisely what the cloud prevents us from seeing! To be able to present cloud-free datasets derived from visible and infrared sensors requires either (a) that data from several overpasses have been combined in a composite, or (b) that gaps in the data where clouds obscured the sea have been filled by some type of analysis such as optimal interpolation based on cloud-free pixels, or using the most recent previous cloud-free data, or substituting climatological values. Sometimes a combination of (a) and (b) is used. Whether such gap-filling is acceptable depends on the application. The user should always adopt a...
critical approach to such data to ensure that erroneous conclusions are not drawn from aspects of the data which are themselves purely artifacts, and not real data at all.

One of the most challenging aspects of visible and infrared remote sensing is detecting those clouds which only partly cover a pixel and may be very hard to distinguish from clear-sky conditions. If incorrectly accepted as clear, they will yield anomalous values of the ocean variable being measured, and can corrupt a dataset. If there is a danger of such false values introducing a bias (e.g., to a climate dataset), then it is best to err on the side of caution. Although it may be disappointing to throw away some good data because they are suspected of being cloud-contaminated, this may be worth it in order to maintain confidence in the quality of the overall dataset. On the other hand, for operational applications long-term biases may not matter so much and a more relaxed approach to cloud contamination can be adopted.

**Atmospheric correction strategies**

Once any pixels obscured by cloud have been detected and flagged, attention must be turned to the remaining "clear-sky" pixels. In general these still contain contributions from the atmosphere, which corrupt the water-leaving signal that we must estimate from the top-of-atmosphere signal, which is what the satellite sensor actually measures. The atmosphere scatters light, especially in the visible and near-infrared wavebands. Some of the radiant energy from the sea to the sensor, which represents the true "signal", is deflected away from the sensor by atmospheric scattering. In addition solar radiation is scattered into the field of view of the sensor to add "noise" to what is observed. In the thermal infrared and the microwave parts of the spectrum, used to measure the temperature of surfaces emitting the radiation, absorption and emission by the atmosphere is the major problem.

The detailed efforts made to eliminate atmospheric effects are specific to individual sensor types. In general it must be noted that for most contexts it is not possible to base an atmospheric correction on *a priori* knowledge of the atmosphere itself. Significant atmospheric interference with the remote-sensing signal comes from atmospheric constituents like water vapor or aerosols whose distribution is not known to sufficient accuracy. Thus we must rely on the remote-sensing measurement itself to contain enough information about atmospheric variability to allow a correction to be made. This is where the use of multiple spectral channels provides an answer. By sampling in several different wavebands, chosen because of their different responses to both the ocean-leaving signal and the effect of the atmosphere, strategies have been developed which allow atmospheric variability to be accommodated under most conditions.

Datasets that have had atmospheric correction applied used to be referred to sometimes as level 1.5 or 1A data products. However, in the case of recent Earth-observing missions, water-leaving radiance datasets in individual visible wavebands, atmospherically corrected but without any further geophysical algorithm being applied, are referred to as level 2 products.
2.3.3 Positional registration

By positional registration we mean the identification in geographical co-ordinates of the place to which a remotely sensed measurement refers (i.e., the location on the ground or sea surface to which the sensor was pointing when it recorded the measurement). Sometimes this is referred to as image navigation, or geolocation. The adjustment of an image to bring it into conformation with a map base is sometimes called geometric correction or rectification of the image.

Fundamentally, the problem is one of knowing where the satellite was when a measurement was made, and in which direction the sensor was pointing. The accuracy with which this information is required depends on the type of sensor and the application. For single measurements with a downward-pointing sensor providing an area-integrated measurement over a $50\text{ km} \times 50\text{ km}$ square, precision in locating the nadir point need not be better than a few kilometers. In such cases it is sufficient to know the expected satellite orbit and nominal pointing direction.

For other nonimaging sensors, a more precise positional registration is required, involving ground station tracking of the satellite and accurate onboard monitoring and transmission of the satellite vehicle’s pitch, roll, and yaw. For some remote-sensing systems this information from diverse sources is brought together and published with the sensor measurement. The precise track of the satellite is known as its ephemeris. Given the tremendous improvement in satellite navigation in recent years using the U.S. global positioning system (GPS) or its Russian counterpart (GLONASS), it is now possible in principle to derive positional registration based on navigational models for all image data. A greater uncertainty than the satellite location is knowing the direction in which the sensor is pointing.

It is worth noting here that, although positional registration is often essential information for both atmospheric correction and geophysical calibration of the data, it is never advisable to use geolocation information to resample a dataset until after these other operations have been performed. Correction and calibration algorithms often use multispectral information in which the ratios or differences between different data channels is crucial. Resampling an image onto a new geographical base requires resampling of the pixels, during which band ratio information may be distorted. If resampling is unavoidable, the use of “nearest neighbor” substitution should be chosen rather than a more elaborate process which averages values from different pixels and exacerbates the problem.

2.3.4 Geophysical product derivation

When the raw data have been calibrated and atmospheric correction has been applied, the next stage in the processing chain is to manipulate the data in order to derive estimates of a particular ocean variable. Because the resulting product is fundamentally different from the measurement originally made by the sensor, it should be considered as a measure or estimate of an ocean variable rather than a measure of electromagnetic radiation. This process is often termed geophysical calibration. In practical terms it is normally performed simply by applying a set of
Developing calibration algorithms

The geophysical calibration of remote sensors generally requires a combination of empirical and theoretical methods. It is customary to start the construction of a calibration model by basing it on a theoretical analysis of the physical processes by which an oceanographic parameter influences the electromagnetic radiation observed by the sensor. In the case of emitted thermal infrared radiation, for example, the theoretical model of radiation emission from black bodies goes a long way towards providing a complete calibration algorithm. Visible wavelength remote sensing of suspended particles in water is less amenable to theoretical analysis. While quite a lot is known about the optical processes in the ocean and atmosphere they are so complex as to make it impossible to produce from first principles a calibration of radiance data in terms of suspended sediment load. In the case of active radar sensors the physical processes which control radar backscatter from rough surfaces cannot be modeled by readily invertible analytical functions and so a more empirical approach is taken.

While it is desirable, if possible, to build calibration algorithms on a sound, physically based, theoretical model, eventually nearly all geophysical algorithms tend to require a set of calibration data in which observations from the satellite are matched to in situ measurements of the ocean variable which is to be derived, obtained at the same time and location as the satellite overpass. If the geophysical algorithm is to be widely applicable, it needs to be based on in situ matchup data that are representative of a wide range of locations, and conditions. Otherwise the algorithm will be biased towards conditions represented by the matchup data, where it performs well, while its performance may be much poorer in other conditions.

Once a geophysical calibration has been performed on satellite data from a single overpass, it is typically distributed as a level 2 dataset in its original “native” grid consisting of the original pixels arranged in rows and columns matching the sensor scan lines and the along-track direction of the satellite.

Ocean product validation

Whether the geophysical algorithm is based on purely theoretical foundations or on empirical matchup datasets, its performance cannot be critically assessed unless there is also in place a robust system of validation. This should generate a set of error statistics indicating the bias and standard deviation of the data products produced from the satellite sensor when compared against a set of reliable independent measurements of the same ocean property.

Validation, at its simplest, consists of comparing the value of an ocean variable determined from the satellite data with an in situ measurement of the same variable, coincident in space and time. In order to be meaningful this must be performed a large number of times. There is a need to test the validity of the algorithm using data spanning the whole range of variable values for which it is specified for use. This
means that data from a single cruise in a limited region, however many coincident
points are obtained, will not be adequate to validate a global algorithm. Hence the
advantage of an array of drifting buoys, or the use of long-transect ships of oppor-
tunity. There is also a need to test the algorithm under a variety of those
environmental conditions which have an influence on it or the earlier parts of the
data-processing chain. For example, if the geophysical products are sensitive to the
atmospheric correction, validation must be performed in a variety of representative
atmospheric conditions before full confidence in its performance can be justified.
Validation must also be continued over the lifetime of the sensor.

Finally it should be obvious that the validation dataset must not be the same as
those used to develop an empirical calibration, since in that case it would not be truly
independent. Those who use data provided by others in developing new scientific or
operational applications of satellite remote sensing must ensure for themselves that
the data products have been validated, at least to the degree of reliability required by
the application.

2.3.5 Image resampling onto map projections

Another stage in the chain of formal processing required to produce useful maps of
ocean variables derived from satellite data is to present them in a geographically
intelligible form. Of all the processes considered so far, this is perhaps the least
significant because it does not essentially affect the accuracy or applicability of the
data. At the image navigation stage mentioned in Section 2.3.3 a model should have
been established for the data which enables the location of each pixel in the original
satellite image to be defined in terms of latitude and longitude. This information by
itself is adequate for many applications of the data, including matching to in situ
observations.

However, one of the great strengths of remotely sensed data is the capacity to
provide detailed spatial views, and to communicate information by allowing a user
to view an image of the ocean. Images are painted on a screen or drawn by a printer
in rectangular form corresponding to the rows and columns of the matrix in which
the parameter values are notionally stored in digital form by the computer. The
orientation of the rows and columns is defined by the viewing geometry of the
sensor, and does not in general bear any relation to normal geographical conventions
for drawing maps. Moreover if the image is drawn in “satellite co-ordinates” there
may well be distortion of shapes compared with their form on the ground. For these
reasons resampling onto a new grid conforming to a standard map projection is
often performed.

The choice of map projection can itself be an issue of some importance for
producers and users of satellite data. The purpose of a map projection is to
present data, whose positions are defined on a spherical surface, as if they were
distributed on a flat plane such as a computer screen or printed page. For a local
region spanning no more than a few hundred kilometers, geometric distortion is
minor and there are few problems. For larger dimensions of thousands of kilometers,
approaching the radius of the Earth, the distortions are much greater. Resampling
data onto the map projection may (a) require data from several sampled points to be located in just one new pixel on the map projection, while (b) other pixels on the new map may contain no satellite data samples at all. In this case careful thought must be given to the rules used for transforming satellite data, whose geolocated position in latitude and longitude places it on a spherical surface, to the map on a plane surface. It is reasonable to suppose that we would preserve the most complete information from the satellite data by applying weighted averages in case (a) and using interpolation from neighboring pixels in case (b). However, as noted in Section 2.3.3, if the differences or ratios of the values in different data channels (e.g., different wavelengths of a multispectral dataset) are to be obtained and exploited from the resampled version of the data, it is essential that the much simpler “nearest neighbor substitution” is used to create the map because only that approach will preserve the original channel ratios as measured by the sensor.

A further difficulty of interpretation arises with the presentation of global datasets on a single flat map. Put at its simplest the problem is this: a spherical surface has no edges but a flat map is always bounded by an edge. For ocean data at low to mid-latitudes the solution is to use cylindrical projection (see Figure 2.9a) in which lines of latitude are horizontal and lines of longitude are always vertical. The main issue then is deciding at which longitude to cut what would otherwise be a continuous strip of data. In the illustration, longitude 180° has been chosen but this splits the Pacific Ocean into two parts. When interpreting such maps it is important not to forget that in reality the data are continuous across the left and right boundaries. However, at high latitudes such maps are inappropriate because east–west distances are greatly exaggerated. A Mollweide type of projection (Figure 2.9b), in which lines of longitude converge to the poles, preserves relative surface areas at different latitudes but at the expense of gross shape distortion. However, like cylindrical projection, it completely obscures any sense of the neighbor relations between high-latitude locations at different longitudes. For this a polar projection (Figure 2.9c) is needed, in order to do justice to the capacity of wide-swath sensors in near-polar orbits to map the polar regions coherently.

For global maps of many ocean properties, the polar regions are of little interest and so cylindrical maps are satisfactory. However, where properties over polar regions are also important (e.g., where sea ice and sea surface temperature are being mapped together), it is necessary to present the data on three separate maps: a global cylindrical map projection and two polar projections, one for each pole.

In future it is likely that the problems of producing datasets in specific map projections, leading to distortion and problems of interpretation, will be overcome by present trends in information technology. There is increasing availability of software that allows any set of data for which the pixel locations are defined in latitude and longitude to be presented in a wide variety of projections. Interactive software offers a three-dimensional view of the Earth and allows users to navigate their vantage point and viewing direction so as to explore the data fully. As this becomes widely available, there will be less need to prepare data in special map projections. In the long term, scientific ocean datasets retrieved from satellites can
Figure 2.9. Examples of map projection types. (a) Cylindrical projection, satisfactory at low to mid-latitudes, amplifies areas at high latitudes. (b) Mollweide projection, preserves equal area at the expense of shape distortion. (c) Polar projection, avoids the loss of a coherent polar view inherent in (a) and (b).
be archived with a spatial density that matches the sampling capacity of the sensors from which they are derived, while viewing software will automatically perform the resampling necessary to display the data to the user at their chosen vantage point. Such an approach will also be able to offer the user the capacity to animate time sequences of mapped data in order to observe the evolution of dynamical processes.

### 2.3.6 Composite image maps

An important process which may be performed to produce a satellite ocean data product is that of combining multiple sets of data from the same sensor but different overpass times and different locations into a single map of data. The result is a composite dataset, generally referred to as a level 3 data product which differs in a number of ways from a level 2 dataset derived from a single overpass. Its typical characteristics are

- It is almost always on a regular geographical grid, resampled onto pixels which are normally oriented with north upwards and incremented in intervals measured in latitude and longitude.
- The pixel size is normally larger than the native pixel in the level 2 image.
- A level 3 composite image is often global in coverage, or extends over a large region such as a single ocean. Its extent is almost always larger than can be encompassed by a single swath of the sensor that supplies its data.
- It contains data from several different overpasses, ascending and descending.
- It is based on data from overpasses within a given timespan.
- It should ideally contain statistical measures of the variability of the level 2 data sources from which it was constructed, such as the number and standard deviation of independent level 2 samples whose average becomes the composite value.

The advantage of a composite image is that it extends the geographical scope of an image dataset beyond the confines of a synoptic image derived from a single overpass, and it is able to fill up over time the gaps that are left either by the inadequate coverage pattern of swaths on a single day, or by the localized dropout of data because of cloud cover or other problems.

However, the construction of a composite image requires the application of rules which can impact the character of the resulting dataset. The main question to address is what to do with multiple entries into a single cell of the composite image. Given the downscaling that is normally involved, several pixels in one of the contributing level 2 images will contribute to a single cell of the composite. Normally the values from each of these should be averaged. Over the integration time for the composite several further level 2 datasets may contribute. These also will be added to the averaging process. However, in some cases only the latest value may be selected, or only those values which, in processing to apply the geophysical calibration, generated the highest confidence value. There is evidently the possibility that subtle changes of the rules for constructing the composite could significantly change the result.
Other design choices for a level 3 composite relate to the size of the pixels and the integration time. In some situations different composites are created from ascending and descending overpasses, which would be appropriate when, for a sensor on a Sun-synchronous satellite, these correspond to different local times of day such as day and night. The balance of advantage between different options depends on the use to be made of the composite data. It should always be remembered that a composite does not contain all the information present in the original level 2 datasets, although careful construction of a composite should enhance the applicability of the data. Composites are particularly useful for observing large-scale ocean phenomena that may be missed by the limited geographical coverage of individual level 2 datasets, and this will be examined in more detail in Section 6.2.1. The user should note carefully what rules have been applied in its production, and consider what implications these may have for interpreting and applying the data. In principle a level 3 composite should contain only data from contributing level 2 images, with the likelihood that some gaps will remain in the end product because of persistent cloud or other sampling limitations. Nonetheless users of composite datasets should check whether they do, in fact, contain inputs from more than the supposed contributing data. For example, persistent data gaps may have been filled by interpolation, or by substitution of the data from the previous integration period, or even from a climatological-averaged dataset.

Strictly, as soon as data from other sources are introduced, or spatial gaps are filled by interpolating adjacent pixels or even simply by smoothing the data field, the outcome should be referred to as a level 4 dataset. Whereas a level 3 product should be based only on what the satellite has measured, a level 4 product is created with the intention of providing the user with a complete data field that contains no gaps. It therefore represents the result of an analysis to estimate what the spatial distribution of a particular ocean variable was at a particular period of time, and so may be referred to as an "analysed field". Although it may be primarily derived from satellite data, we should expect it to contain more than just measurements from a single satellite. It may contain information from different types of satellite sensors, and be blended with in situ observations.

The processes used to fill gaps and smooth them will have constrained them to conform to certain expectations of what they should be like. Therefore, although for most applications it is far more helpful for users to be given a seamless and smooth distribution of an ocean variable rather than a sparse patchwork of what was actually observed, its interpretation must always bear in mind the assumptions that went into its construction. For example, if the production of a level 4 analysis of an ocean variable implicitly smooths observed data spatially or temporally with a low-pass filter, it would be inappropriate to use it for investigating high-frequency variability in that ocean property. The most useful level 4 products should be accompanied by confidence values for each pixel that will alert the user to issues where misinterpretation might occur. For example, it is to be expected that the confidence value will be lower for those pixels where no observations have been available to influence the analysis. This will be discussed further in Section 14.4.2 which explains...
how analyzed sea surface temperature data products are prepared for operational applications.

### 2.4 SENSOR TYPES FOR OBSERVING THE OCEAN

#### 2.4.1 Using the electromagnetic spectrum

All sensors employed on ocean-observing satellites use electromagnetic (EM) radiation to view the sea. The capability of particular sensors to measure certain properties of the ocean and how well they can view through the atmosphere or penetrate clouds depends critically on which part of the EM spectrum they use (for a fuller discussion see *MTOFS*, chapter 2). Figure 2.10 shows regions of the electromagnetic spectrum that are of relevance to remote sensing and in particular the bands occupied by the four broad classes of satellite sensors used for viewing the ocean. The types of sensor found in these broad classes, the primary measurement they make, and the parameters that can be derived from them are summarized in Figure 2.11.

![Figure 2.10. The electromagnetic spectrum, showing the regions exploited by typical remote-sensing instruments.](image)
Figure 2.10 also shows how transmittance of the atmosphere varies with EM wavelength, which accounts for why sensors are only found in certain wavebands. For much of the EM spectrum the atmosphere is opaque and therefore unusable for remote sensing of the ocean. However, there are a number of “window” regions where most of the radiation gets through although it may be attenuated to some extent. One of these windows extends from the visible part of the spectrum (between 400 nm and 700 nm, used by the human eye) into the near-infrared (IR). This is used by radiometers that observe sunlight reflected from the ground and the ocean. In the context of this book these are termed ocean color sensors (Section 2.4.2 outlines how they work). Between wavelengths of about 3.5 μm and 13 μm are found a number of narrow windows exploited by IR radiometers. This is called the thermal IR part of the spectrum because much of the radiation has been emitted by surfaces according to their temperature. In ocean remote sensing it is used for measuring sea surface temperature (SST).

At much longer wavelengths, greater than a few millimeters, the atmosphere becomes almost completely transparent. This is referred to as the microwave part of the spectrum, normally differentiated spectrally in terms of frequency rather than wavelength. Because it is widely exploited for many technological aspects of modern civilization, including radio and TV broadcasts, telecommunications, mobile telephony, and so on, certain parts have to be specially reserved for remote sensing and are allocated by international regulation. They are found as discrete narrow-frequency bands within the broad regions indicated in Figure 2.10 for microwave radiometry and radars. Microwave radiometers are passive sensors, simply measur-
ing the naturally ambient radiation that is emitted by the ocean, atmosphere, and land surfaces. Radars are active microwave devices which emit pulses and measure echoes from the sea surface, in order to gain information about some aspect of the surface. Radar frequency bands are sometimes referred to by the letters indicated in Table 2.2.

There is a variety of different types of radar, which can be distinguished by the direction in which they point, the length and modulation of the emitted microwave pulse, and the way the echo from the sea surface is analyzed. Radars can be classed as either viewing straight down at the nadir point below the platform, or viewing obliquely to encounter the surface at an incidence angle between 15° and 60°. The nadir sensors measure surface height or slope and are called altimeters. Those viewing obliquely measure a property called $\sigma_0$, the normalized radar backscatter cross-section, which is related to surface roughness at lengthscales comparable with the radar wavelength.

In the rest of Section 2.4 the measuring capability of each sensor class is summarized in a separate subsection. The principles of operation of each sensor type is outlined, while the most important elements of the data-processing and interpretation tasks are highlighted. It is also important to recognize that in general the direct measurements made by a satellite sensor, even after applying atmospheric corrections when necessary, are not themselves oceanographic parameters. Apart from the example of sea surface temperature, the primary quantity observed by most sensors requires a further stage of interpretation to generate derived parameters that are useful for oceanographers, as summarized by the bottom row of Figure 2.11

### Table 2.2. Definition of common radar bands used for ocean remote sensing.

<table>
<thead>
<tr>
<th>Band</th>
<th>Frequency (GHz)</th>
<th>Wavelength</th>
</tr>
</thead>
<tbody>
<tr>
<td>L</td>
<td>0.390–1.55</td>
<td>19.35–76.9 cm</td>
</tr>
<tr>
<td>S</td>
<td>1.55–4.20</td>
<td>7.14–19.35 cm</td>
</tr>
<tr>
<td>C</td>
<td>4.20–5.75</td>
<td>5.22–7.14 cm</td>
</tr>
<tr>
<td>X</td>
<td>5.75–10.9</td>
<td>2.75–5.22 cm</td>
</tr>
<tr>
<td>$K_u$</td>
<td>10.9–22.0</td>
<td>1.36–2.75 cm</td>
</tr>
<tr>
<td>$K_a$</td>
<td>22.0–36.0</td>
<td>8.33–13.6 mm</td>
</tr>
</tbody>
</table>

#### 2.4.2 Ocean color radiometers

The basic principle of ocean color remote sensing is straightforward. The light measured by an ocean color sensor pointing towards the sea comes originally from the Sun. Some photons of light emitted by the Sun, with energies that place them in the visible part of the spectrum, enter the sea where they are either absorbed or scattered depending on what is contained in the seawater. Those of the scattered photons that emerge again give the sea its apparent color. This is quantified by a satellite ocean color sensor which measures the amounts of different wavelengths of light reaching it. A *multispectral radiometer* typically samples a limited number of narrow wavebands, chosen to capture the main structure of the spectral shape of incoming light (see Figure 2.12). An *imaging spectrometer* samples in full detail across the spectrum, but such instruments have so far been used mainly onboard
From the relative magnitude of the water-leaving radiance detected by the different spectral channels of a radiometer, methods have been developed to estimate the concentration of those water constituents which give the sea its color. The term *ocean color* is used loosely in remote sensing to refer to both the magnitude and the spectral composition of the light leaving the water. In practice it is the spectral radiance at the top of the atmosphere that is measured from a satellite. As shown in Figure 2.13, this consists of light reflected by the atmosphere, the sea surface, and (in very shallow water) the sea bed, as well as backscattered by seawater constituents. The retrieval of useful oceanographic quantities from top-of-atmosphere measurements is a challenging task, requiring careful separation of

Figure 2.12. A typical spectrum of Earth-leaving radiance in the visible and near-infrared part of the spectrum, observed from a satellite above the atmosphere, as spectrally sampled by (a) a multispectral radiometer and (b) an imaging spectrometer with a spectral resolution of about 5 nm.

Figure 2.13. Factors which affect the light reaching an ocean color sensor.
atmospheric scattering and surface reflection from true water-leaving radiance. The rest of this section briefly outlines how this is tackled, but for a more comprehensive introduction to the physical principles underlying ocean color remote sensing, and the detailed methodology which has matured over more than 20 years, the reader should consult *MTOFS* (chapter 6 and the references therein to the scientific literature).

**Atmospheric correction**

The greater part of the measurement of visible wavelengths of light made looking down from a satellite orbit comes from light scattered by the atmosphere into the field of view of the sensor, which may contribute more than 90% of total measured radiance. A proportion of the light leaving the sea surface is also scattered out of the field of view and so the atmospheric correction procedure must account for both these factors in order to estimate water-leaving radiance for each of the spectral bands recorded by the sensor.

That part of atmospheric correction due to scattering by air gas molecules themselves can be calculated directly for each pixel in the field of view of an imaging sensor, from knowledge of the relative positions of the Sun, the satellite, and the pixel. Scattering is also caused by larger particles of aerosols in the atmosphere. These may be water vapor or dust particles, but unlike atmospheric gases their concentration and distribution in the atmosphere is unknown and impossible to predict. Fortunately a technique has been developed which allows the variable contribution of scattering from aerosols to be estimated from measurements made by the sensor.

The key is to use radiance measured in two spectral bands from the near-infra-red part of the spectrum. Because the sea readily absorbs almost all solar near-infrared radiation incident upon it, any light in these wavebands measured at the top of the atmosphere must have been scattered by the atmosphere or reflected at the surface. This can be used to estimate how much aerosol scattering has occurred in visible channels where the water-leaving radiance is not zero, and so the correction is accomplished. Thus near-infrared channels are essential for an ocean color sensor, although care must be taken to select the correct wavebands that do not overlap gas absorption lines in the spectrum.

**Reflection from surfaces**

Part of the measured signal is reflected directly from the sea surface and this has no value for quantifying the water content. If the satellite detects direct specular reflection from the Sun it will dominate all other parts of the signal. Sun glitter prevents further analysis of the signal and must be avoided if at all possible, by careful selection of the orbit and sensor geometry in relation to the position of the Sun, and by tilting the sensor away from the Sun, bearing in mind that the extent of the Sun glitter region on the sea surface varies with the roughness of the sea.

Reflection from the sea surface of sky light (i.e., sunlight already scattered by the atmosphere) cannot be avoided. Instead corrections can be made for it within atmo-
spheric correction. This means that knowledge of sea state (based on wind speed) is also needed for atmospheric correction.

The other surface which may contribute reflections to the measured signal is the sea bed. Since sunlight is fairly rapidly attenuated with depth in seawater, by both absorption and scattering in the water column, reflection from the sea bed is only a problem when the sea is both shallow and clear. In most circumstances it is not an issue, but it can create problems when interpreting ocean color data in tropical, shallow seas. Where the water is very clear then bottom reflections in shallow seas can allow the color signal to be used to detect either the depth of the sea, the character of the sea bed (sand, coral, vegetation, etc.), or both but this is not a major application for global ocean color sensors.

**Interpreting ocean color**

The color of the sea is not of itself an ocean variable of particular interest for most marine scientists. However, the factors which influence ocean color, such as the presence of phytoplankton, the concentration of pigments associated with primary production or dissolved organic material, and the concentration of suspended particulates are all of considerable oceanographic importance. Measurements of these properties can be derived from the color.

When atmospheric correction has been successfully applied to satellite ocean color data, the result is an estimate of the water-leaving radiance in each spectral channel in the visible waveband, normalized to reduce dependence on the Sun’s elevation and the viewing incidence angle. Effectively, normalized water-leaving radiance should represent what a sensor would measure if looking straight down from an orbit that carries it just above the sea surface at the bottom of the atmosphere. This is what our eyes would detect as the color and brightness of the sea, ignoring any light reflected from the surface. The primary challenge of ocean color remote sensing is to derive quantitative estimates of the type and concentration of those materials in the water that affect its apparent color.

Photons from the Sun, with EM energy corresponding to visible wavelength, enter the sea and eventually interact with molecules of seawater or its contents. The outcome will be either that the photon is scattered, in which case it may change its direction with a chance of leaving the sea and contributing to what the sensor measures, or it will be absorbed. The probability of scattering or absorption depends on the wavelength of the light and the material which it encounters. The molecules within seawater tend to preferentially scatter shorter wavelengths of light (the blue part of the spectrum) and preferentially absorb longer wavelengths (the red end). This is why pure seawater with little other content appears blue.

The pigment chlorophyll-α which is found in phytoplankton has a strong and fairly broad absorption peak centered at 440 nm in the blue, but not in the green. Therefore, as chlorophyll concentration increases, more blue light is absorbed while green light continues to be scattered, and so from above the seawater looks greener. This is the basis for many of the quantitative estimates of seawater content derived
from satellite ocean color data. The typical form of an algorithm to estimate the concentration of chlorophyll \( C \) or phytoplankton biomass is:

\[
C = A \left( \frac{R_{550}}{R_{490}} \right)^B,
\]

where \( A \) and \( B \) are empirically derived coefficients; and \( R_\lambda \) is reflectance (radiance coming out of the sea towards the sensor, normalized by incoming irradiance) over a spectral waveband of the sensor centered at wavelength \( \lambda \). This is described as the green/blue ratio. In the open sea it is possible to estimate \( C \) to an accuracy of about 30\% by this means. Most algorithms presently in use are somewhat more complex than (2.3) but are still closely related to it. If the sample data from which the coefficients \( A \) and \( B \), etc. are derived is representative of many different open-sea situations then such algorithms can be applied widely in many locations.

Other substances which interact with the light and so change the apparent color of the sea are suspended particulate material (SPM) that has a fairly neutral effect on color except in the case of highly colored suspended sediments, and colored dissolved organic matter (CDOM, sometimes called “yellow substance”) which absorbs strongly towards the blue end of the spectrum. Both of these affect the light along with the chlorophyll “greening” effect when there is a phytoplankton population. However, because the chlorophyll, CDOM, and SPM all co-vary within a phytoplankton population the green–blue ratio effect dominates the color and each of these materials can be quantified by an algorithm such as (2.3), as long as phytoplankton are the only major factor other than the seawater itself that affects the color. Such conditions are described as being Case 1 waters, and it is here that ocean color algorithms work fairly well to retrieve estimates of \( C \) from satellite data.

However, if there is SPM or CDOM present from a source other than the local phytoplankton population (e.g., from river runoff or resuspended bottom sediments), then we can no longer expect any simple relationship between the concentrations of these and \( C \). In this situation green–blue ratio algorithms do not perform very well, if at all, and it becomes much harder to retrieve useful quantities from ocean color data using universal algorithms. These are described as Case 2 conditions. Unfortunately it is not easy to distinguish between Case 1 and Case 2 waters from satellite data alone. This can result in very degraded accuracy with errors of 100\% if standard chlorophyll algorithms are applied in Case 2 waters. It is prudent to classify all shallow-sea areas as Case 2, particularly where there are riverine and coastal discharges or strong tidal currents stirring up bottom sediments, unless in situ observations confirm that Case 1 conditions apply.

Another useful measurement that can be derived from ocean color is the optical diffuse attenuation coefficient, \( K \), usually defined at a particular wavelength such as 490 nm (i.e., \( K_{490} \)). This is also inversely correlated with the blue–green ratio because the less the attenuation coefficient, the deeper the light penetrates before it is scattered back out, the more of the longer wavelengths are absorbed, and the bluer the water appears. The algorithms for \( K \) are similar in form to (2.3) and are somewhat less sensitive to whether the conditions are Case 1 or Case 2.
**Ocean color sensors and products**

Table 2.4 in Section 2.5 lists the important ocean color sensors that have been flown. Although visible wavelength radiometers were among the very first Earth-observing sensors used in the 1970s, a gap of 18 years to the next launch has resulted in the development of ocean color sensors being less mature than that of other methods of satellite oceanography. The Sea-viewing Wide Field-of-view Sensor (SeaWiFS) launched in 1997 provided the first reliable, long-term, fully operational delivery of ocean color data products. Since then a number of sensors with finer spectral resolution have been flown. All the sensors in Table 2.4 fly in low (~800 km) Sun-synchronous polar orbits, providing a resolution at nadir of about 1.1 km and almost complete Earth coverage in 2 days. MERIS is also capable of a high-resolution mode with 300 m square pixels. Other color sensors have also been flown by individual countries offering less comprehensive coverage and poorer data availability than those listed.

All of the sensors listed in Table 2.4 are supported by a calibration and validation program and their data are worked up by the responsible agency into derived oceanographic products at level 2 and in some cases level 3. In all cases some measure of $C$ is produced and an estimate of $K$ is derived globally. These are generally reliable products and $C$ approaches the target accuracy of 30% in open-sea Case 1 waters. However, great care must be taken when using the products over coastal water and shelf sea water (possibly Case 2) where the potentially large errors could give misleading information. Some agencies offer a number of other products estimating the concentration of certain material in the seawater such as SPM and CDOM but these are yet to be proven.

In addition, some allied products are offered by some agencies providing ocean color data. These may be partially derived from the ocean color sensor but have input from other sources: satellites, *in situ* measurements, or models. In that respect they could be classed as level 4 products. Examples of these are surface solar irradiance (SSI) and photosynthetically available radiation (PAR).

### 2.4.3 Thermal infrared radiometry for measuring sea surface temperature

Whereas visible waveband radiometers rely on reflected sunlight and can operate only during the local daytime, in the thermal infrared and microwave parts of the spectrum most observed radiation will have been thermally emitted by the sea surface. In this way infrared and microwave radiometers can be used directly to measure the radiation temperature of surfaces. Given knowledge about the emissivity of the sea surface this can be used to estimate the physical temperature of the water. For infrared measurements there is a close relationship between emitted infrared radiation and sea surface temperature (SST). The challenges facing the satellite oceanographer are how to remove the effect of the atmosphere from measured infrared radiance and still derive SST estimates to an accuracy within a few tenths of a Kelvin, and how to relate temperature measured by a radiometer from
above with what a thermometer placed in contact with the sea would measure. These are the issues introduced in this section.

**Physical principles**

An infrared sensor records the radiance detected at the top of the atmosphere in specific wavebands, $\lambda_n$. The individual measurements in each channel, $n$, can be expressed as an equivalent black-body brightness temperature, $T_{bn}$. That is the temperature at which a black body (a surface with 100% emissivity) would emit the measured radiance. At a particular wavelength, black-body emission is defined by the Planck equation:

$$L(\lambda, T) = \frac{C_1}{\pi \lambda^5 \left[ \exp\left(\frac{C_2}{\lambda T}\right) - 1\right]},$$

where $L$ is spectral radiance, per unit bandwidth centered at $\lambda$, leaving the unit surface area of the black body, per unit solid angle (W m$^{-2}$ m$^{-1}$ str$^{-1}$); $\lambda$ is the wavelength (m); $T$ is the temperature (K) of the black body; $C_1 = 3.74 \times 10^{-6}$ W m$^{-2}$; and $C_2 = 1.44 \times 10^{-2}$ m K. This must be integrated with respect to wavelength over the measured waveband and convolved with the spectral sensitivity of the sensor in order to represent radiance intercepted by a particular spectral channel. The spectral shape of Equation (2.4) and how it varies with temperature are shown in Figure 2.14. Black-body radiation is an ideal theoretical concept. The actual radiation emitted from the sea surface is a fraction, $\varepsilon$, called the emissivity, which is close to one for seawater and IR radiation.

Figure 2.14 also indicates the “atmospheric window” regions of the thermal infrared spectrum in which radiation passes through the atmosphere with only a small amount of attenuation, in the absence of cloud. These are found between about 3.5 µm and 4.1 µm (referred to as the 3.7 µm window), and between 10.0 µm and 12.5 µm. The latter is often used for two separate wavebands, 10.3 µm to 11.3 µm and 11.5 µm to 12.5 µm, generally referred to as the “split window” channels. All the main sensors used to measure SST by infrared radiometry use these three channels, including the two main IR sensor series referred to in this book. These are the Advanced Very High Resolution Radiometer (AVHRR) developed by the National Oceanographic and Atmospheric Administration (NOAA) in the U.S.A. and the Along-track Scanning Radiometer (ATSR) a British-designed sensor developed and flown by the European Space Agency (ESA).

To obtain $T_{bn}$ from the digital signal $S_n$ recorded by the sensor for waveband $n$ requires direct calibration of the sensor using two onboard black-body targets of known temperatures which straddle the range of ocean surface temperatures being observed. This is the method adopted by the ATSR class of sensor, whereas the AVHRR uses the simpler but less accurate alternative of a single onboard black body with a view of cold space serving as an alternative to the second black body. Calibration targets are viewed once for every scan across the swath.
Atmospheric correction

Ideally we wish to measure the radiance leaving the water surface, which is determined by the skin temperature of the sea, \( T_s \), and by the emissivity of seawater. In the thermal infrared this is greater than 0.98, but a small contribution to satellite-detected radiance comes from the reflected sky radiance, for which allowance must be made. Because of absorption by gases in the atmosphere \( T_{bn} \) is cooler than \( T_s \) by an amount which varies in time and place, mainly with the amount of atmospheric water vapor. It is the task of the atmospheric correction procedure to estimate \( T_s \) given top-of-atmosphere measurements of \( T_{bn} \).

A well-established method of atmospheric correction is to make use of differential attenuation in different wavebands. This is illustrated schematically in Figure 2.15. In case A, water vapor is assumed to be less than in case B, causing less absorption in both channels \( i \) and \( j \) for case A than for B, and requiring a larger correction to be applied to B than to A, although we cannot say directly how big either correction should be. However, if channels \( i \) and \( j \) respond differently to water vapor they will record a difference \( \Delta_{i,j}(T_b) \) between the top-of-atmosphere brightness temperatures \( T_{bi} \) and \( T_{bj} \) measured simultaneously by each channel. This spectral difference is also related to the number of absorbing gases in the atmospheric path and so \( \Delta_{i,j}(T_b) \) will be smaller for case A than for B. Thus the size of \( \Delta_{i,j}(T_b) \) provides a measure of the degree of atmospheric attenuation. This is utilized by

Figure 2.14. Infrared emission spectra of black bodies at temperatures between \(-10^\circ C\) and \(40^\circ C\). The gray bands show the location of atmospheric windows. At other wavelengths the atmosphere is opaque.
algorithms of the form:

\[ T_s = aT_{bi} + b(T_{bi} - T_{bj}) + c , \]  

(2.5)

where \( a \), \( b \), and \( c \) are coefficients to be determined. During the day, the algorithm uses only split window channels, while at night the 3.7 \( \mu \)m channel can also be used. The latter is corrupted by reflected solar radiation and so cannot be used in the daytime. A number of nonlinear variants of this basic form have also been developed (Barton, 1995).

Common to these algorithms for AVHRR is the requirement for coefficients to be determined by a best fit between satellite predictions and coincident observations of SST from a number of drifting buoys. The match between buoys and satellites has a variance of more than 0.5 K, applicable only to regions populated by buoys. The same algorithms are assumed to apply to parts of the ocean where there are no buoys, although the validity of this assumption needs to be quantified. Regional algorithms matched to local data can achieve greater accuracy.

Although the instantaneous distribution of water vapor and aerosols in the atmosphere are not known, the radiation transfer physics of the atmosphere is well understood and can be modeled with some confidence in fine spectral detail. It is therefore possible to simulate \( T_b \) for a given combination of \( T_s \), atmospheric profile and viewing angle for the spectral characteristics, and viewing geometry of each channel of a particular sensor. This offers an alternative strategy for atmospheric correction in which an artificial dataset of matching \( T_s \) and \( T_{bi} \), \( T_{bj} \), etc. is created using a wide variety of typical atmospheric water vapor and temperature profiles. The coefficients for an equation of form similar to Equation (2.5) are generated by a regression fit to the artificial dataset. The resulting algorithm should be applicable to all atmospheric circumstances similar to those included in the modeled dataset, leading to an estimate of the skin SST. It is independent of coincident in situ measurements, although those are needed for validation. This is the approach adopted for the ATSR.

The ATSR scans conically to observe a forward view at about 60° incidence angle and a near-nadir view of the same patch of sea about 2 min to 3 min later (as illustrated in Figure 2.16). Using the same three spectral channels as AVHRR for
each of the views, it thus acquires six measures of brightness temperature (four during the daytime). The different pathlengths for forward and nadir views provide extra information leading to a more robust algorithm. The single-view approach failed temporarily when large volumes of volcanic dust were suddenly injected into the stratosphere by the eruption of Mt. Pinatubo in 1991 (Reynolds, 1993), whereas for ATSR a reworking of the semiphysical model by including stratospheric aerosols in the radiation model led to algorithms which cope well with volcanic ash or similar problems (Merchant and Harris, 1999; Merchant et al., 1999).

Cloud detection

Atmospheric correction algorithms produce maps of SST at fine resolution (about 1.1 km) for each overpass. However, atmospheric correction methods cannot retrieve SST when cloud wholly or partly obstructs the field of view. Therefore at this stage cloud must be detected using a variety of tests (Saunders and Kriebel, 1988), so that only cloud-free pixels are retained for oceanographic applications, such as assimilation into models. The most difficult cloud contamination to identify is that by subpixel-size clouds, thin cirrus, or sea fog where only small deviations of temperature occur. Failure to detect cloud leads to underestimation of the SST and can produce cool biases of order 0.5 K. Thus confidence in the cloud detection procedure
is just as important as atmospheric correction for achieving accurate SST. Where uncertainty remains in cloud detection, this should be flagged in the error estimate fields attached to SST products. Cloud detection is generally more successful during daytime, when visible and near-IR image data can be used, than at night.

**Which sea surface temperature does a satellite measure?**

There is an additional pitfall for interpreting satellite-derived SST data, consequent upon the thermal structure in the top few meters of the ocean. Two distinct factors create near-surface vertical temperature gradients, illustrated schematically in Figure 2.17. First, on sunny calm days a diurnal thermocline tends to develop above which an upper layer is found, a meter or so thick and up to about 1 K warmer than below (although exceptionally it can be several Kelvins warmer). At night the warm layer collapses. Second (and independently of the first effect), the top skin layer of the sea, a fraction of a millimeter thick, tends to be a few tenths of a Kelvin cooler than the water immediately below.

The problem for understanding remotely sensed SST is that there are various methods of measuring SST samples at different levels of the near-surface thermal structure (shown in Figure 2.17). Thus the term “SST” means different things for the thermometer on a buoy’s hull, for a sensor in a ship’s cooling water intake, for an infrared radiometer, for a microwave radiometer, and for an ocean model. These differences are important when accuracies of a few tenths of a Kelvin are required. They may also vary considerably during the day so that if a single daily sample is

![Figure 2.17. Schematic diagram showing characteristic temperature profiles at the sea surface for (a) nighttime conditions or daytime with moderate to strong winds and (b) daytime calm to light wind conditions and direct solar heating.](image-url)
used to characterize SST the result may be aliased depending on the time in the
diurnal cycle at which it is sampled.

Therefore the practice is now being adopted of distinguishing between skin SST,
the temperature in the top few microns, and subskin SST a short distance (of order
1 mm) below the surface. These are separated by the thermal skin layer where heat
transport is restricted to molecular conductivity because of the suppression of
turbulence close to the surface. The subskin is typically a few tenths of a degree
warmer than the actual skin. Infrared radiometers measure skin SST whereas
microwave radiometers, penetrating deeper, approximately measure subskin SST.

In addition a new term, foundation SST, has been coined to describe the
temperature on which diurnal warming (if any) is built each day. It is most clearly
specified as the temperature of the well-mixed layer found just below the skin layer at
dawn, when any diurnal thermocline structure from the previous day has collapsed.
At this time of day it is equivalent to subskin SST. Foundation SST is defined on a
daily basis. It corresponds to what oceanographers generally mean when they refer
to the temperature of the upper mixed layer of the ocean. When in situ measurements
are made from buoys or ships they sample the structure shown in Figure 2.17 at an
indeterminate depth, typically somewhere between subskin and foundation SST.
This creates additional uncertainty for using in situ observations to calibrate SST
atmospheric correction algorithms.

**SST data products**

SST level 2 image datasets, at a resolution of about 1 km and arranged in satellite
scan co-ordinates, are produced within an hour of AVHRR overpasses and made
available for a variety of applications. Because raw AVHRR data are broadcast
directly, local receiving stations around the world can obtain highest resolution
data and produce SST products in near-real time. Because of the complexity of
processing ATSR data, having to resample the forward and nadir views from their
curved scan lines onto a rectangular grid before applying the atmospheric correction
algorithm, the SST products from ATSR are produced only by ESA, but are still
made available within, at most, a few hours of the overpass. Infrared scanners on
geostationary satellites generate image datasets every 15 min or 30 min. The most
recent geostationary sensors are well calibrated and use multiband radiometry for
accurate atmospheric correction, capable of generating SST maps of comparable
accuracy to AVHRR and ATSR. The disadvantage of these products is that their
coverage does not extend to high latitudes and their spatial resolution is poorer,
typically 4 km or 5 km pixels. A summary of oceanographically important infrared
sensors is given in Table 2.5 (to be found in Section 2.5).

Until recently, oceanographers looking for a satellite-based record of global or
ocean-wide SST distribution and how it evolves with time would go to one of the
global composite datasets constructed from a particular infrared sensor (as outlined
in Section 2.3.6). A number of different composite SST products have been pro-
duced, with lengthscales varying from 1/2° to 1/6° (about 50 km to 16 km at the
Equator) or finer, and time steps from 4 days to 1 week or 1 month. From these,
annual climatologies were created allowing SST anomalies to be quantified relative to the seasonal norm (see Section 6.2.1) The most widely used were those derived from AVHRR, such as multichannel sea surface temperature (MCSST) (Walton et al., 1998) or Pathfinder SST (Vazquez et al., 1998), which is a reprocessing of the archived pixel-level AVHRR data with algorithms incorporating the best knowledge of sensor calibration drift and making full use of the available drifting buoy dataset (Kilpatrick et al., 2001). The global composite product from ATSR is called averaged SST (ASST) which was reprocessed using the more robust atmospheric algorithms (Merchant and Harris, 1999; Merchant et al., 1999).

However, most oceanographic users looked for the best possible estimate of SST, rather than a product from a particular sensor. Reynolds and Smith (1994) developed a level 4 analysis product that was an optimal interpolation of both in situ and satellite data, and should therefore provide better climatological continuity with pre-satellite SST records from before 1980. The most significant development in SST monitoring since 2002 has been an international collaborative initiative to exploit the complementarity between different sensors rather than leaving users to choose one of the competing SST data products. This is the GODAE High Resolution SST Pilot Project—GHRSSST-PP (Donlon et al., 2007). Its main achievement has been to persuade Earth observation agencies to produce level 2 maps of SST in a common format with essential ancillary data, which has enabled new level 4 SST analyses to be created that draw from all available level 2 products. The primary aim is to improve the usefulness of satellite data for assimilation into ocean-forecasting models (as discussed further in Chapter 14), and for creating robust SST records for climate time series. In Section 2.6 pointers are given for accessing the datasets mentioned above.

2.4.4 Microwave radiometry

It was noted in the previous subsection that radiation in the microwave as well as the thermal infrared parts of the spectrum is thermally emitted by the sea surface and can be used directly to measure the radiation temperature of surfaces. However, unlike the infrared, the microwave brightness temperature of a surface is not related so directly to its physical temperature and depends also on other properties of the surface. This has created a different set of challenges for the development of ocean measurement capabilities using microwave radiometry. Although in many ways the use of passive-microwave radiometers for measuring SST is inferior to using infrared sensors, the technique does have the very significant advantage of being able to see through cloud.

Physical principles of microwave radiometry

The physics of microwave emission from a surface of temperature \( T \) is deceptively simple. Ideal black-body emission is expressed as a linear dependence on
temperature in a simplified form of the Planck equation:

\[ B_f = \frac{2kT}{c^2}, \]  

where \( B_f \) is spectral radiance expressed as radiance per unit frequency interval. Because of this linear relationship, microwave radiation is itself often referred to as brightness temperature (i.e., the temperature of the black-body source that would generate the measured radiance).

However, the use of passive-microwave radiometers to measure SST is more complex than Equation (2.6) might lead us to believe. Unlike thermal IR wavebands, where \( \varepsilon \approx 0.98 \) and the emitted radiation is dominated by skin temperature, in the microwave region the \( \varepsilon < 0.5 \) for the sea surface. It turns out that emissivity depends on the viewing incidence angle relative to the local surface slope, and the dielectric constant of seawater as well as temperature. Since the dielectric constant also depends on temperature, the dependence of microwave brightness temperature on subskin SST is not linear. Moreover radiation may change if the mean square surface slope or surface salinity changes, even if the SST remains constant. While this is clearly a drawback for measuring SST, it does offer the possibility of using microwave radiometers for detecting sea surface roughness, another of satellite oceanography’s primary observable quantities. It also implies that salinity may be considered to be a fifth primary observable quantity, and so it has been added to the third row of Figure 2.11 although, at the time of writing, this has still to be demonstrated from a satellite sensor. Figure 2.18 is a cartoon summarizing the various different environmental factors which influence microwave emission from the sea surface and its passage through the atmosphere.

**Elements of microwave radiometers**

Microwave radiometers are passive devices. Unlike radars they do not create their own coherent source of energy and so are incapable of many of the complex signal-processing techniques used to enhance the resolution of active instruments. Radiometers measure the power of the continuous, incoherent, electromagnetic radiation incident upon their detectors. They sample within specific narrow-frequency bands and some radiometers are capable of differentiating between the power in different polarization orientations. Radiometers are restricted to specific microwave wavebands, not because of atmospheric windows as for infrared and visible radiometers, but because the microwave spectrum is used extensively by modern telecommunications and broadcasting infrastructure. These signals would swamp background radiation from natural sources, but by international agreement certain bands are not permitted for use by radio sources in order to preserve them for passive radiometry.

Some instrument designs use a parabolic reflector to focus the ground view onto the detector. By rotating the reflector about a vertical axis the field of view scans across the ground in a circular arc, maintaining the same incidence angle for all samples. Calibration is achieved by pointing at a calibration source on the satellite.
In a different type of radiometer design the signal is recorded from each element in an array of unfocused detectors. By integrating these signals with different timelags the effective field of view can be steered to obtain a spatially resolved field of brightness temperature. However, whether mechanical or electrical focusing is used, the spatial resolution is between one and two orders of magnitude poorer than for an IR radiometer. Because of poor focusing, microwave radiometers are not reliable within about 100 km of land, because of the occurrence of stray signals and brightness contrasts between land and sea which leak through sidelobes in the power pattern of antennas.

**Retrieving geophysical quantities from microwave radiometers**

It is possible to distinguish between different contributions to the brightness temperature of SST, surface roughness, and salinity, as well as to identify atmospheric contamination by liquid water, because each factor differentially affects different microwave frequencies. For example, SST strongly affects wavebands between 6 GHz and 11 GHz whereas the effects of salinity are found only at
frequencies below about 3 GHz. Surface roughness effects influence frequencies at 10 GHz and above, and are also polarization-specific. Thus a multifrequency and multipolarization radiometer can, in principle, be used to measure SST, surface wind, and precipitation (see chapter 8 of Robinson, 2004). The retrieval of useful oceanographic measurements is based mainly on using empirical algorithms, developed from matchups between in situ observations and satellite data.

**Microwave radiometry missions useful for oceanography**

Table 2.6 (see Section 2.5) lists the microwave radiometers used by oceanographers and the data products that they have delivered. Since the mid-1980s a series of Special Sensor Microwave Imagers (SSM/I) have flown on a U.S. defense satellite program, delivering mainly meteorological products (Wentz, 1997), and used by oceanographers to detect sea ice or for wind speed measurements. However, serious consideration of microwave measurements of SST from space started only when a sensor having a 10.7 GHz channel was flown on the Japanese–U.S. Tropical Rainfall Measuring Mission. Called the TRMM microwave imager (TMI) it has a spatial resolution of 0.5° (about 50 km) and because it oversamples it is capable of mapping mesoscale eddies quite effectively using a gridscale of 25 km. It lacks the preferred SST waveband of 6.6 GHz, but its 10.7 GHz channel is sensitive to SST in tropical water temperatures. It covers only latitudes lower than 40°.

In 2002 the Japanese Advanced Microwave Scanning Radiometer (AMSR-E) was launched into a near-polar orbit on the NASA Aqua satellite. This sensor includes a channel at 6.6 GHz, which is sensitive over the full range of sea temperatures, and has opened the way for routine, high-quality, global mapping of SST by microwave radiometry. AMSR-E is now providing global cloud-free SST to an accuracy of ~0.4 K derived from oversampled 76 km resolution data. Composite daily, weekly, and monthly SST products are supplied on a 1/4° grid.

In the context of the GHRSST Pilot Project mentioned in Section 2.4.3, SST retrievals from microwave radiometry are now routinely used to complement infrared radiometers for producing SST analysis (level 4) products, their strong contribution coming from their ability to sample daily with few data dropouts caused by atmospheric conditions. It is therefore of considerable concern to operational oceanographers that the continuity should be maintained of microwave radiometers with 6.6 GHz capability when the existing sensors reach the end of their life.

There are, however, experimental sensors planned to test the capability of ocean salinity measurement from satellites. The European Space Agency’s Soil Moisture and Ocean Salinity (SMOS) satellite was launched in late 2009. Its main payload is an electronically focusing synthetic aperture microwave radiometer operating in L band (1.4 GHz). NASA are preparing another L-band radiometer for launch on their Aquarius mission in 2010, focused on measuring global ocean surface salinity.
2.4.5 Altimetry for measuring surface slope, currents, and wave height

A more comprehensive introduction to the scientific principles and the detailed methods of satellite altimetry, and a description of 30 years of evolving altimeters, can be found in chapter 11 of *MTOFS*.

**The principle of altimetry over the ocean**

A satellite altimeter is a nadir-viewing radar which emits regular pulses and records the travel time, the magnitude, and the shape of each return signal after reflection from the Earth’s surface. Travel time is the essential altimetric measurement, leading to determination of the ocean surface topography at length scales longer than about 100 km. Ocean surface topography contains information about ocean dynamical and geophysical phenomena. If the travel time can be measured to a precision of $6 \times 10^{-11}$ s then, knowing the speed of light, the distance can be calculated to a resolution of 1 cm. Corrections have to be made to allow for the changed speed of light through the ionosphere and the atmosphere, and for delays associated with reflection from a rough sea surface (Chelton *et al.*, 2001). It is generally agreed that for these corrections to approach the target accuracy of 1 cm a dual-frequency altimeter must be used (to determine ionospheric refraction), and a three-channel microwave radiometer is needed to sound water vapor in the atmosphere.

The altimeter is not an imaging sensor. Viewing only the nadir point below the satellite, it simply records measurements of distance between the satellite and the sea surface along the ground track. As discussed in Section 2.2, the spatial and temporal-sampling characteristics therefore depend entirely on the exact orbit repeat cycle of the satellite. This was chosen to be about 10 days for the TOPEX/Poseidon (T/P) and Jason altimeters which fly on platforms dedicated to the altimetric mission, although for other altimeters it has ranged between 3 days, 17 days, 35 days, and longer. The longer the revisit interval the finer the spatial-sampling grid. Typically, ocean topography data are interpolated onto a geographical grid and composited over the period of an exact repeat cycle, to produce “images” which are comparable with global SST or ocean chlorophyll composite images although produced in a completely different way.

By itself, knowing the distance $R_{alt}$ between the ocean surface and a satellite is of limited value. Figure 2.19 shows what else needs to be defined or measured for this to yield an oceanographically useful property. First of all, when the height of the satellite, $H_{sat}$, is known relative to a reference level, then the height, $h$, of the sea above the reference level can be determined. The reference level is a regular ellipsoid-shaped surface defined within a frame of reference fixed in the rotating Earth. It is chosen to match approximately the shape of the Earth at sea level, and provides a convenient datum from which to measure all other heights.

Several physical factors contribute to $h$, which is called the ocean surface topography. The first is the distribution of gravity over the Earth, as represented by the geoid, at height $h_{geoid}$ above the reference ellipsoid in Figure 2.19. The geoid is the equipotential surface, at mean sea level, of the effective gravitational field of the Earth which incorporates Earth rotation forces and the gravitation of the solid
Earth, the ocean itself, and the atmosphere. By definition it is normal to the local effective gravity force, and if the ocean were everywhere in stationary equilibrium relative to the Earth, its surface would define the geoid.

Another factor which contributes to $h$ is $h_{\text{tide}}$, the instantaneous tidal displacement of the sea surface relative to its tidally averaged mean position, including the contribution of the Earth tide. A third is the local response, $h_{\text{atm}}$, of the ocean to the atmospheric pressure distribution over the ocean, approximated by the inverse barometer effect in which an increased pressure of 1 mbar lowers sea level by 1 cm. The remaining factor is displacement of the sea surface associated with the motion of the sea, called the ocean dynamic topography $h_{\text{dyn}}$. Thus:

$$h = h_{\text{dyn}} + h_{\text{geoid}} + h_{\text{tide}} + h_{\text{atm}}.$$ (2.7)

The dynamic topography is the property which is of most relevance for ocean modeling since it contains information about ocean circulation. Rearranging (2.7) and substituting $h = H_{\text{sat}} - R_{\text{alt}}$ yields:

$$h_{\text{dyn}} = H_{\text{sat}} - R_{\text{alt}} - h_{\text{geoid}} - h_{\text{tide}} - h_{\text{atm}}.$$ (2.8)

The accuracy and precision of the estimated ocean dynamic height depends not only on the altimetric measurement itself but also on the other four terms in (2.8). For dedicated altimetry missions flying at a height of about 1,340 km where atmospheric drag is minimal, the height of the satellite in orbit, $H_{\text{sat}}$, can now be predicted to a precision of 2 cm using a combination of laser and microwave-tracking devices and an orbit model using precise gravity fields. The tidal contribution has been evaluated along the repeat orbit track by tidal analysis of the altimeter record spanning several years. Because tidal frequencies are very precisely known the response to each constituent can be evaluated to an accuracy better than 2 cm in the open ocean, even though the sampling interval of about 10 days is longer than most tidal periods. This is only possible when the precise period of the repeat cycle is chosen to avoid any serious aliasing with one of the major tidal constituents. For this reason a Sun-synchronous orbit, which aliases the $S_2$ (solar semidiurnal) tidal signal, should not be
used. Over shelf seas where tides are very high and can vary rapidly over short
distances it is not so easy to remove the tides and so the estimate of dynamic
height is less accurate. Atmospheric pressure correction is based on the output of
atmospheric circulation models.

**Evaluating sea surface height anomaly**

At the time of writing, the geoid has been measured independently, but not yet to a
very high precision, and so oceanographers must be content with measuring the
combined \( h_{\text{dyn}} + h_{\text{geoid}} \). Of these, the typical magnitude of the spatial variability of
\( h_{\text{geoid}} \) is measured in tens of meters, about 10 times greater than that of \( h_{\text{dyn}} \), which is
why until recently the time-mean ocean topography from altimeters provided geo-
physicists with the best measure of the geoid. However, \( h_{\text{geoid}} \) does not vary with
time, at least not sufficiently to be detected by an altimeter over tens of years,
whereas the time-variable part of \( h_{\text{dyn}} \) is comparable in magnitude with the mean
component, of order meters over a few months.

Therefore the time variable part of \( h_{\text{dyn}} \), called the sea surface height
anomaly, SSHA, can be separated from the measured \( h_{\text{dyn}} + h_{\text{geoid}} \) by simply sub-
tracting from it the time-mean sea surface height over many orbit cycles
(\( \text{MSS} = \text{Mean}\{h_{\text{dyn}} + h_{\text{geoid}}\} \)). To enable a time-mean to be produced, the orbit
track must be precisely repeated to within a kilometer and the data must be accu-
culated from several years of a 10-day cycle. For this reason it is essential to fly a
new altimeter in precisely the same orbit as its predecessor so that the mean surface
topography of the earlier mission can be used straight away. Then SSHA can be
calculated from the first orbit cycle of the new altimeter, without having to wait
another few years to build up a new mean topography for a different orbit track.

It is important to remember that the SSHA, which is widely used for oceano-
graphic analysis and is assimilated into dynamical ocean models, does not contain
any information about the dynamic height of the ocean associated with the mean
circulation. Global maps of SSHA do not display the dynamic topography signa-
tures of the strong ocean currents at all, apart from the fact that the eddy-like
activity is strongest where the major currents tend to meander.

There are presently three families of altimeters in operation, as listed in Table 2.7
in Section 2.5, with details of their attitude, orbit repeat, and approximate accuracy
(root mean square) of an averaged SSHA product. The T/P–Jason family is a joint
French/U.S.-dedicated altimetry mission in a high non–Sun-synchronous orbit. In
contrast the Geosat and ERS series are on lower Sun-synchronous platforms for
which orbit prediction accuracy would be, on their own, much poorer. However,
because these satellites cross over each other’s orbit tracks it is possible, over an
extended timespan, to significantly improve their orbit definitions by cross-
referencing to the better known T/P or Jason orbits (Le Traon *et al.*, 1995; Le
Traon and Ogor, 1998). The accuracy quoted for the SSHA applies only after this
procedure has been performed, and would otherwise be much worse for the ERS and
Geosat families. The specification of errors for an altimeter must be handled with
care because the error magnitude relates very much to the time and spacescale over
which it is being averaged. The lower error attached to larger scale/longer period
averaging must be offset against the lesser utility of the averaged SSHA field, 
especially in the context of operational oceanography.

The data products from altimeters are presented first as along-track values of
SSHA, wind speed (determined from the peak height of the echo), and significant
wave height (from the pulse shape—see below and Chapter 8). These level 2 products
are sampled every second along track, and are contained in the Geophysical Data
Record (GDR), which also includes ancillary information about the various correc-
tions applied. Whilst each agency publishes the GDR for their own altimeter,
scientific users of the data may find it most helpful to work with data where cross-
referencing between different altimeters has been performed in a consistent way,
referred to as the Data Unification and Combination System (DUACS), ensuring
that there should be little if any bias between the SSHA from different satellites.
Data are also resampled onto a $1/3^\circ \times 1/3^\circ$ Mercator grid, integrated over a period
of time that relates to the orbit repeat interval. Details of these vary between the
different agencies producing products (see Table 2.10).

Variable currents from sea surface height anomaly

To determine an estimate of the time-variable part of ocean surface currents,
geostrophic equations are used:

$$
\begin{align*}
fv &= g \frac{\partial h_{SSHA}}{\partial x} \\
fu &= -g \frac{\partial h_{SSHA}}{\partial y}
\end{align*}
$$

where $(u, v)$ are the east and north components of geostrophic velocity; $f$ is the
Coriolis parameter; $g$ is the acceleration due to gravity; and $x$ and $y$ are distances
in the east and north direction, respectively.

From a single overpass, only the component of current in a direction across the
altimeter track can be determined, but where ascending and descending tracks cross
each other the full vector velocity can be estimated. Because Equation (2.9) assumes
geostrophic balance, if there is any ageostrophic surface displacement it will lead to
effects in $(u, v)$. However, ageostrophic currents should not persist for longer than
half a pendulum day ($1/f$) before adjusting to geostrophy. Thus the spatially and
temporally averaged SSHA maps produced from all the tracks acquired during a
single repeat cycle (10, 17, or 35 days depending on the altimeter) should represent a
good approximation to a geostrophic surface that can be inverted to produce surface
geostrophic currents.

Close to the Equator the SSHA cannot be interpreted directly in terms of surface
currents since here $f$ is very small and the geostrophic equations (2.9) cannot be
applied.
New altimeter data products containing independent geoid data

In the relatively near future it is hoped that the lack of knowledge about the geoid can be remedied. What is needed is a means of measuring $h_{\text{geoid}}$ without using altimetry, and this is provided by measurement of the gravity field above the Earth from satellites. Both the presently operating Gravity Recovery And Climate Experiment (GRACE) and the Gravity and Ocean Circulation Explorer (GOCE) mission which was launched in mid-2009 (although first results are still awaited as this book goes to press) measure elements of the gravity field from which it is possible to recreate the sea level geoid. At the required accuracy of about 1 cm GRACE can achieve this only at a lengthscale longer than several hundred kilometers, but it is hoped that GOCE can do so once and for all down to a lengthscale of about 100 km. This will allow steady-state ocean currents to be derived from archived altimetric data and greatly improve the capacity to utilize altimetric data in near-real time.

In anticipation of the eventual availability of a high-quality, independent geoid from GOCE, a hybrid mean dynamic topography (MDT) was produced (Rio and Hernandez, 2004) using the following approach. The absolute dynamic topography of the sea surface, $h_{\text{dyn}}$, is the sum of the SSHA and MDT. Eventually MDT should be determined precisely by subtracting an independent measure of $h_{\text{geoid}}$ from MSS. This was done approximately using the EIGEN-GRACE03S geoid, evaluated to spherical harmonic degree 30, which implies that it contains little useful information on geoid variability at lengths less than 400 km but is quite well defined for lengthscales above 660 km. To improve the accuracy of MDT at shorter lengthscales it was fitted to the dynamic height associated with in situ measurements of steady currents using an inverse technique. The in situ data were buoy velocities from the WOCE–TOGA program, corrected for mesoscale variability using coincident SSHA. Comparison with independent velocity observations show differences to be globally less than 13 cm/s r.m.s. From MDT a new altimetry product is produced (Table 2.10) called the absolute dynamic topography (ADT = MDT + SSHA) from which absolute currents can be estimated using standard geostrophic retrieval, following Equation (2.9) with $h_{\text{ADT}}$ replacing $h_{\text{SSHA}}$.

Measuring significant wave height from altimeters

When an altimeter measures the time for an emitted pulse to return, it tracks in detail the shape of the leading edge of the echo, from which it is possible to make a very good estimate of significant wave height, $H_{1/3}$, within the pulse-limited footprint illuminated by the altimeter. For a perfectly flat, calm surface the return echo has a very sharp edge. If there are large waves, several meters in height from trough to crest, then the return signal starts to rise earlier, as the first echoes are received from the crests, but takes longer to reach its maximum, when the first echoes are received from the wave troughs. The rising edge of the echo is modeled by a function in terms of the root mean square ocean wave height, so that by matching the observed shape to the model function it is easy to gain an estimate of $H_{1/3}$. This method has delivered robustly accurate measurements of $H_{1/3}$ for more than 20 years from...
different altimeters (Cotton and Carter, 1994) and comparison with buoys shows root mean square differences of only 0.3 m (Gower, 1996), which is the limit of buoy accuracy. Applications of this method to wave monitoring and forecasting are discussed further in Chapter 8.

2.4.6 Oblique-viewing radars for measuring sea surface roughness

Active microwave devices provide their own energy, in the form of radar pulses which are emitted from spacecraft, reflected from the sea surface, and received back at the sensor again. Since the amount of energy reflected depends largely on the short-scale profile of the surface at length scales comparable with the radar wavelength, most radars provide information about sea surface roughness as the primary observable quantity.

Interpreting radar backscatter measurements

The magnitude of the radar echo reflected from the sea surface is expressed as a variable called the normalized radar backscatter cross-section, usually referred to by the symbol $\sigma_0$. After calibration, the data from a radar yield estimates of $\sigma_0$, either as single averages for a given field of view or as an array of many samples mapped over the sea surface. The size of $\sigma_0$ depends on surface roughness and in particular on the amplitude of short waves on the sea surface propagating in the radar ground range direction and having a wavelength of $n\lambda/(2 \sin \theta)$—where $n$ is 1, 2, etc.; $\lambda$ is the radar wavelength; and $\theta$ is the radar incidence angle. This is the Bragg resonance mechanism, as a consequence of which different radar frequencies produce different magnitudes of echoes from the same sea surface and same incidence angle.

Figure 2.20 illustrates broadly how $\sigma_0$ varies with incidence angle under different wind conditions or sea states. The behavior of $\sigma_0$ can be separately characterized in three ranges of incidence angle. At low incidence angles (a) specular reflection appears to be the dominant process. For a very calm sea there is a very narrow angular response giving a very high return at $0^\circ$ incidence which rapidly drops off as the incidence angle increases. For a somewhat rougher surface under moderate winds the nadir-viewing response is weaker, but does not decay so rapidly with increasing viewing angle, so that within a few degrees from normal incidence it is reflecting more power than the flat surface. The very high sea state continues the trend, with an even lower $\sigma_0$ at $0^\circ$ but very little dropoff with incidence angle. This is the way in which the magnitude of altimeter pulses responds to surface roughness.

In a central region of the diagram (b) at incidence angles between about 20$^\circ$ and 70$^\circ$, appropriate for most oblique-viewing radars, the behavior of $\sigma_0$ is quite simply described. At a given $\theta$ it increases with sea state, while there is an approximately linear reduction with increasing viewing angle, except for a calm sea that is already very low. Finally at incidence angles greater than 70$^\circ$ (c) the value of $\sigma_0$ appears to drop off more rapidly with $\theta$, reaching very low values at grazing incidence approaching 90$^\circ$. The broad dependence on sea state, albeit different in different bands of viewing angle, is what makes $\sigma_0$ such a useful parameter for marine remote
Note that in practice $\sigma_0$ depends also on other parameters such as frequency and polarization, so Figure 2.20 is not intended to be precise. Sections 9.3 and 9.4 of *MTOFS* (Robinson, 2004) provide a much fuller discussion of radar backscatter from the ocean, with many references to the wide literature on this subject.

**Scatterometers**

A scatterometer is the simplest type of radar used for remote sensing. It is an oblique-viewing radar pointed towards the sea from aircraft or satellites at incidence angles normally between $20^\circ$ and $70^\circ$. The receiver simply measures the backscattered power from the field of view in order to determine $\sigma_0$. There is no attempt to preserve phase information after demodulation of the microwave signal. Therefore it does not resolve variations of $\sigma_0$ in range or azimuth in a detailed way and cannot generate a high-resolution image. By measuring the average $\sigma_0$ over a wide area of sea (with a spatial resolution typically 20–50 km) it uses this to estimate the wind speed.

The interpretation of scatterometer measurements of backscatter relies on an empirically derived model of the relationship between $\sigma_0$, wind speed, incidence angle, and the direction of the wind relative to the radar azimuth. As long as $\sigma_0$ at each point on the ground is measured at least twice in close succession, viewing from different directions, there is in principle enough information to be able to retrieve an estimate of wind speed and direction using this model. Scatterometers
deployed to provide operational meteorological measurements have swaths spanning about 1,500 km and can view the global ocean surface twice in two days. A much fuller discussion of scatterometers and the principles of scatterometry can be found in sections 9.6 and 9.7 of MTOFS (Robinson, 2004). Chapter 9 of the current book shows how wind fields measured by scatterometers are contributing to oceanographic applications.

Imaging radars

With an active device, there is scope to measure not only the energy flux of the reflected signal, but also its detailed amplitude and phase, depending on how complex a measuring device is used, and how much data can be sampled and transmitted back to the ground station. Thus the timing of the return signal can be used to resolve between patches of sea surface at different distances from the radar. Moreover, the detailed shape of the return pulse can be compared with the pulse that was originally transmitted and, for example, Doppler shifts can be detected. When suitably analyzed, such information can be made to yield further information about the sea surface, and in particular to improve the spatial resolution of detection making it possible to generate detailed images of surface roughness. Instruments that collect such detailed information are known as imaging radars. Most imaging radars on satellites belong to a class known as synthetic aperture radars (SARs) because of the way they process data to recover detailed spatial resolution in the azimuth direction. These are described in detail in chapter 10 of MTOFS (Robinson, 2004).

While short-scale surface roughness may not seem at first to be a very important oceanographic parameter, much important oceanographic information can be derived from it in addition to sea surface wind strength and direction which drives it in the first place. There are a number of upper-ocean phenomena and processes which modulate wind-driven, short surface ripples. These become much more evident using SARs that can resolve details down to 30 m. One source of modulation is variable surface tension caused by the presence of surfactant material, which allows SARs to reveal the presence of oil spills and sea surface slicks. Another major cause of modulation is by small-scale patterns of convergent and divergent currents at the sea surface. Driven by phenomena such as long-surface swell waves, internal waves, flow over undulating shallow topography, or ocean fronts and eddies, the hydrodynamic interaction between variable surface currents and the energy of Bragg ripples is able to generate signatures in the $\sigma_0$ field. Thus processes whose center of action may be tens of meters below the sea surface are “painted” on the radar images, providing an unexpected opportunity to gain new scientific understanding of subsurface phenomena. For example, Chapter 12 in this volume shows how much new knowledge about internal waves has come from the analysis of SAR image data. The capacity of SAR to contribute a unique perspective to the measurement of long-surface gravity waves is demonstrated in Chapter 8, while other, sometimes unexpected, oceanographic applications of SAR image data will emerge in other chapters.
2.5 PLATFORMS AND SENSORS FOR SATELLITE OCEANOGRAPHY

This section summarizes, mainly in tabular form, information about the main satellites and sensors that have been used for ocean remote sensing. Table 2.3 lists the more important satellites, indicating their type of orbit and the ocean-viewing sensors they have carried. The list is by no means complete, since many agencies from many countries have launched satellites that have delivered some useful ocean measurements. However, this is intended to serve as a pointer to those satellite series which today deliver the majority of data for routine ocean monitoring, to previous satellites whose data form a valuable body of archived ocean observations spanning 30 years, and also to those pioneering satellites that proved the concepts underlying the methods of ocean remote sensing. Where indications are given of ocean-viewing satellites and sensors later than 2009, which is when the tables were compiled, they are based on firm plans of space agencies but readers must confirm for themselves whether they have been successfully launched. A fuller list of most satellites with any relevance to ocean observation up to 2004 can be found in table 3.2 of *MTOFS*.

The rest of the section contains details of the main sensors used for the different ocean remote-sensing methods that were summarized in Section 2.4. Table 2.4 lists the ocean color sensors whose data are most widely used by ocean scientists. Table 2.5 lists the infrared radiometers used for measuring sea surface temperature. In both cases it must be pointed out that several other sensors have been, or still are, delivering data, but not routinely or with wide and fairly open access to the data. The sensors listed here are those that the reader is most likely to encounter in the scientific literature on ocean applications of remote sensing, or whose data can be readily obtained. Table 2.6 lists the third class of passive ocean-observing sensor, microwave radiometers. Note in this case the variety of different data products available from microwave radiometry.

The remaining tables identify the more important active-microwave sensors. Table 2.7 lists the altimeters, Table 2.8 the high-resolution imaging radars (synthetic aperture radars—SARs), and Table 2.9 the scatterometers. Note that more information can be found in Chapter 8 about the various sensors used for measuring ocean surface waves, and in Chapter 9 about sensors used to measure the wind over the sea, while Table 7.1 lists some high-resolution visible and near-infrared sensors used for land mapping but which have potential for seabed mapping in tropical coastal ecosystems.

2.6 SATELLITE OCEAN DATA PRODUCTS

In this final section of the chapter, our attention focuses on the ocean data products that are now readily available for users wishing to apply them to some specific operational monitoring or forecasting task, for ocean scientists to use as observational research data, or simply for the curious who are fascinated to discover how the ocean is behaving. The current trend in the dissemination of satellite data is towards providing users with fully processed end products, validated estimates of an actual
Table 2.3. Satellites carrying important ocean-viewing sensors. Entries in bold refer to series.

<table>
<thead>
<tr>
<th>Satellite</th>
<th>Period of useful operation</th>
<th>Agency</th>
<th>Orbit type</th>
<th>Ocean observing sensors or sensor types deployed</th>
</tr>
</thead>
<tbody>
<tr>
<td>Landsat-1 to -7</td>
<td>1972-present</td>
<td>NASA/USGS</td>
<td>Leo, P, SS</td>
<td>MSS, TM, ATM, ETM+</td>
</tr>
<tr>
<td>Meteosat-1 to -7</td>
<td>1977-present</td>
<td>ESA/Eumetsat</td>
<td>Geo</td>
<td>VISSR</td>
</tr>
<tr>
<td>Seasat</td>
<td>July-Sep, 1978</td>
<td>NASA</td>
<td>Leo, NSSS</td>
<td>MMR, Scat, SAR, RA</td>
</tr>
<tr>
<td>Nimbus-7</td>
<td>1978–1986</td>
<td>NASA</td>
<td>Leo, P, SS</td>
<td>CZCS, SMMR</td>
</tr>
<tr>
<td>TIROS-N</td>
<td>1978–1981</td>
<td>NOAA</td>
<td>Leo, P, SS</td>
<td>AVHRR</td>
</tr>
<tr>
<td>NOAA-7 to -18</td>
<td>1981–present</td>
<td>NOAA, U.S.A.</td>
<td>Leo, P, SS</td>
<td>AVHRR/2, AVHRR/3</td>
</tr>
<tr>
<td>Geosat</td>
<td>1985(86)-90</td>
<td>NASA</td>
<td>Leo, P, 17-day ERM RA altimeter</td>
<td></td>
</tr>
<tr>
<td>SPOT-1 to -4</td>
<td>1986–present</td>
<td>CNES, France</td>
<td>Leo, P, SS</td>
<td>DORIS, HRVIR, Vegetation</td>
</tr>
<tr>
<td>DMSP-F8 to -F15</td>
<td>1987–present</td>
<td>DoD, U.S.A.</td>
<td>Leo, P, SS</td>
<td>SSM/I ocean meteorology</td>
</tr>
<tr>
<td>ERS-1</td>
<td>1991–1999</td>
<td>ESA, Europe</td>
<td>Leo, P, SS</td>
<td>RA, AMI (SAR-Scat), ATSR</td>
</tr>
<tr>
<td>TOPEX-Poseidon</td>
<td>1992–2005</td>
<td>NASA/CNES</td>
<td>1,336 km, 10-day ERM non-SS</td>
<td>DORIS, Poseidon-1, TOPEX altimeter</td>
</tr>
<tr>
<td>GOES-8 to -12</td>
<td>1994–present</td>
<td>NOAA, U.S.A.</td>
<td>Geo</td>
<td>GOES I-M Imager</td>
</tr>
<tr>
<td>ERS-2</td>
<td>1995–present</td>
<td>ESA, Europe</td>
<td>Leo, P, SS</td>
<td>RA, AMI, ATSR-2, PRARE</td>
</tr>
<tr>
<td>Radarsat-1</td>
<td>1995–present</td>
<td>Canada</td>
<td>Leo, P, SS</td>
<td>SAR</td>
</tr>
</tbody>
</table>

(continued)
### Table 2.3 (cont.)

<table>
<thead>
<tr>
<th>Satellite</th>
<th>Period of useful operation</th>
<th>Agency</th>
<th>Orbit type</th>
<th>Ocean observing sensors or sensor types deployed</th>
</tr>
</thead>
<tbody>
<tr>
<td>ADEOS</td>
<td>1996-1997</td>
<td>NASDA, Japan</td>
<td>Leo, P, SS</td>
<td>OCTS, Scat</td>
</tr>
<tr>
<td>Seastar</td>
<td>1997–present</td>
<td>NASA</td>
<td>Leo, P, SS</td>
<td>SeaWiFS ocean color sensor</td>
</tr>
<tr>
<td>TRMM</td>
<td>1997–present</td>
<td>NASA, NASA</td>
<td>Leo, non SS</td>
<td>TMI</td>
</tr>
<tr>
<td>Geosat FO</td>
<td>1998–present</td>
<td>U.S. Navy</td>
<td>Leo, P, 17-day ERM</td>
<td>RA altimeter</td>
</tr>
<tr>
<td>Quikscat</td>
<td>1999–present</td>
<td>NASA</td>
<td>Leo, P, SS</td>
<td>SeaWinds Scatterometer</td>
</tr>
<tr>
<td>Terra</td>
<td>1999–present</td>
<td>NASA</td>
<td>Leo, P, SS</td>
<td>MODIS imaging spectrometer</td>
</tr>
<tr>
<td><strong>Jason-1, -2</strong></td>
<td>2001–present</td>
<td>CNES/NASA</td>
<td>Leo, 10-day ERM non-SS</td>
<td>Poseidon-2 altimeter</td>
</tr>
<tr>
<td>Envisat</td>
<td>2002–present</td>
<td>ESA</td>
<td>Leo, P, SS</td>
<td>ASAR, RA-2, AATSR, MERIS</td>
</tr>
<tr>
<td>Aqua</td>
<td>2002–present</td>
<td>NASA</td>
<td>Leo, P, SS</td>
<td>AMSR-E, MODIS</td>
</tr>
<tr>
<td>MSG</td>
<td>2002–present</td>
<td>Eumetsat ESA</td>
<td>Geo</td>
<td>SEVIRI radiometer</td>
</tr>
<tr>
<td>ADEOS-2</td>
<td>2002–2003</td>
<td>JAXA, Japan</td>
<td>Leo, P, SS</td>
<td>AMSR, GLI, POLDER, SeaWinds</td>
</tr>
<tr>
<td>Coriolis</td>
<td>2003–present</td>
<td>DoD, U.S.A.</td>
<td>Leo, P, SS</td>
<td>Windsat M/w radiometer</td>
</tr>
<tr>
<td>METOP-1</td>
<td>2006–present</td>
<td>Eumetsat/ESA</td>
<td>Leo, P, SS</td>
<td>AVHRR/3, HIRS/4ASCAT</td>
</tr>
</tbody>
</table>

Leo = low Earth orbit, Geo = geostationary orbit, P = near polar, SS = sun-synchronous, X d ERM = X-day exact repeat mission.
ocean variable accompanied by measures of accuracy and reliability. This widens access to satellite-derived data compared with the first two decades of ocean remote sensing when only raw or half-processed data were available, which required the user to acquire skills in remote-sensing data processing before she or he could draw useful information from them. While this trend is making for much greater usability of satellite ocean data, it is still important for users to understand the limitations of the data and how to get the best from them. This section guides the reader towards selecting ocean image datasets from the more reliable sources of ocean satellite data, and identifies some software tools for manipulating these image data.

As outlined in Section 2.3, published satellite data are normally categorized as being processed to a certain level. Most of the useful sources provide ocean data products at level 2 or above (see Table 2.1) and users must consider which level is appropriate for their needs. If they are interested in observing fairly high–resolution ocean phenomena as they occur in a particular region (e.g., tracking the position of a front, or monitoring the development of a localized phytoplankton bloom), then level 2 products would be appropriate. To avoid being swamped by data they should restrict their geographical search area. At level 2 the data may be provided in satellite co-ordinates (with axes along-track and across-track) and so each image from a time sequence of the same nominal area may not necessarily overlay the others precisely. Ideally the geographical locations of each pixel will be specified in another array within the dataset. At level 2 there are likely to be gaps in each image caused by

<table>
<thead>
<tr>
<th>Sensor acronym</th>
<th>Platform</th>
<th>Full name of sensor</th>
<th>Agency</th>
<th>Dates</th>
<th>No. of narrow spectral channels</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Visible</td>
</tr>
<tr>
<td>CZCS</td>
<td>Nimbus-7</td>
<td>Coastal zone color scanner</td>
<td>NASA</td>
<td>1978–1986</td>
<td>4</td>
</tr>
<tr>
<td>OCTS</td>
<td>Adeos</td>
<td>Ocean color and thermal sensor</td>
<td>NASA</td>
<td>1996–1997</td>
<td>6</td>
</tr>
<tr>
<td>SeaWiFs</td>
<td>Sea Star</td>
<td>Sea-viewing wide-field-of-view sensor</td>
<td>NASA</td>
<td>1997–present&lt;sup&gt;a&lt;/sup&gt;</td>
<td>6</td>
</tr>
<tr>
<td>MODIS</td>
<td>TERRA</td>
<td>Moderate-resolution imaging spectrometer</td>
<td>NASA</td>
<td>2000–present&lt;sup&gt;a&lt;/sup&gt;</td>
<td>7</td>
</tr>
<tr>
<td>MERIS</td>
<td>Envisat</td>
<td>Medium-resolution imaging spectrometer</td>
<td>ESA</td>
<td>2002–present&lt;sup&gt;a&lt;/sup&gt;</td>
<td>8</td>
</tr>
<tr>
<td>MODIS</td>
<td>AQUA</td>
<td>Moderate-resolution imaging spectrometer</td>
<td>NASA</td>
<td>2002–present&lt;sup&gt;a&lt;/sup&gt;</td>
<td>7</td>
</tr>
<tr>
<td>GLI</td>
<td>MIDORI</td>
<td>Global imager</td>
<td>NASDA</td>
<td>2002–2003</td>
<td>12</td>
</tr>
</tbody>
</table>

<sup>a</sup> October 2009.
cloud or other factors depending on the type of data. Agencies that provide access to
data typically offer a search and selection Internet interface so that users can specify
a location, a period of time, and possibly a filter to eliminate images that have too
few cloud-free pixels.

Those users looking at larger scale phenomena and global maps are more likely
to prefer level 3 or level 4 data. Invariably these will already be gridded in a latitude
and longitude array. Although data dropout because of cloud may occur at level 3 it
will be less problematic than at level 2, while level 4 data are analyzed products from

### Table 2.5. Recent and current series of high-quality satellite infrared radiometers.

<table>
<thead>
<tr>
<th>Sensor acronym</th>
<th>Platform(s)</th>
<th>Full name of sensor</th>
<th>Agency</th>
<th>Dates (for series)</th>
<th>Main (^a) IR spectral bands ((\mu m))</th>
</tr>
</thead>
<tbody>
<tr>
<td>AVHRR/2</td>
<td>NOAA-7, -9, -11, -12, -14, METOP</td>
<td>Advanced very high-resolution radiometer, version 2</td>
<td>NASA/NOAA</td>
<td>June 1981–Mar 2001</td>
<td>0.725–1.10, 3.55–3.93, 10.3–11.3, 11.5–12.5</td>
</tr>
<tr>
<td>AVHRR/3</td>
<td>NOAA-15, -16, -17, -18, METOP</td>
<td>Advanced very high-resolution radiometer, version 3</td>
<td>NASA/NOAA</td>
<td>May 1998–present (^b)</td>
<td>0.725–1.10, 1.58–1.64, 3.55–3.93, 10.3–11.3, 11.5–12.5</td>
</tr>
<tr>
<td>ATSR-1, ATSR-2, AATSR</td>
<td>ERS-1 (^e), ERS-2 (^e), Envisat (^e)</td>
<td>Along-track scanning radiometer</td>
<td>ESA</td>
<td>Sep 2001–present (^b)</td>
<td>1.45–1.75 (^c), 3.55–3.85 (^d), 10.3–11.3, 11.5–12.5</td>
</tr>
<tr>
<td>SEVIRI</td>
<td>Meteosat second generation</td>
<td>Spinning enhanced visible and infrared imager</td>
<td>Eumetsat</td>
<td>Sep 2002–present (^b)</td>
<td>1.50–1.78, 3.48–4.36, 8.30–9.10, 9.80–11.80, 11.00–13.00</td>
</tr>
</tbody>
</table>

\(^a\) Most sensors have additional visible band(s) used for daytime cloud detection.
\(^b\) April 2008.
\(^c\) Day.
\(^d\) Night.
\(^e\) Each has a forward and a nadir view
several sources and should consist of completely full fields. Once a sensor, or a series of sensors, has been producing data for several years, climatologies ought to be available, as should anomaly maps (the production of anomaly datasets is explained in Section 6.2.1). Users should consider which of these they require for a particular application. The rest of this book provides examples of many ocean applications of satellite data and points towards the most appropriate dataset to use.

Oceanographers requiring satellite data for a particular application ought to approach the selection of a suitable source with care. They should consider

---

**Table 2.6. Recent and current series of satellite microwave radiometers.**

<table>
<thead>
<tr>
<th>Sensor acronym</th>
<th>Platform(s)</th>
<th>Agency</th>
<th>Dates (for series)</th>
<th>Channels</th>
<th>Main data products</th>
</tr>
</thead>
<tbody>
<tr>
<td>SSM/I (Special Sensor Microwave Imager)</td>
<td>DMSP: F8, F10, F11, F13, F14, F15</td>
<td>U.S. Dept. of Defense (DoD)</td>
<td>Sep 1987–present (^b)</td>
<td>Center frequency (MHz)</td>
<td>Wind speed(^a)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>19.35</td>
<td>V, H</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>22.235</td>
<td>V, H</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
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<td>37.0</td>
<td>V, H</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>85.5</td>
<td>V, H</td>
</tr>
<tr>
<td>TMI (TRMM Microwave Imager)</td>
<td>TRMM</td>
<td>NASA/ JAXA</td>
<td>Nov 1997–present(^b)</td>
<td>V, H</td>
<td>SST</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10.7</td>
<td>V, H</td>
</tr>
<tr>
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<td></td>
<td></td>
<td></td>
<td>19.4</td>
<td>V, H</td>
</tr>
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<td>21.3</td>
<td>H</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>37.0</td>
<td>V, H</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>85.5</td>
<td>V, H</td>
</tr>
<tr>
<td>AMSR-E (Advanced Microwave Scanning Radiometer)</td>
<td>Aqua</td>
<td>JAXA/ NASA</td>
<td>May 2002–present(^b)</td>
<td>V, H</td>
<td>SST</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>6.925</td>
<td>V, H</td>
</tr>
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<td></td>
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<td></td>
<td>10.65</td>
<td>V, H</td>
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<td>18.7</td>
<td>V, H</td>
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<td>23.8</td>
<td>V, H</td>
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<td></td>
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<td>36.5</td>
<td>V, H</td>
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<td>89.0</td>
<td>V, H</td>
</tr>
<tr>
<td>WindSat</td>
<td>Coriolis</td>
<td>U.S. DoD</td>
<td>Jan 2003–present(^b)</td>
<td>V, H</td>
<td>SST, wind speed and direction</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>6.8</td>
<td>FP(^c)</td>
</tr>
<tr>
<td></td>
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<td></td>
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<td></td>
<td>18.7</td>
<td>FP(^c)</td>
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<td></td>
<td></td>
<td></td>
<td>23.8</td>
<td>FP(^c)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>37.0</td>
<td>FP(^c)</td>
</tr>
</tbody>
</table>

\(^a\) Wind speed of 10 m.
\(^b\) September 2009.
\(^c\) FP = fully polarimetric (see sections 8.2.6 and 8.4.7 of MTOFS).
The methods of satellite oceanography

Table 2.7. Recent and current series of satellite altimeters.

<table>
<thead>
<tr>
<th>Altimeter</th>
<th>Agency</th>
<th>Dates</th>
<th>Height</th>
<th>Orbit</th>
<th>SSHA r.m.s. accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>TOPEX/Poseidon</td>
<td>NASA/CNES</td>
<td>1992–2005</td>
<td>1,336 km</td>
<td>9.92 day repeat</td>
<td>2–3 cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>non–Sun-synchronous</td>
<td></td>
</tr>
<tr>
<td>Poseidon-2 on Jason-1</td>
<td>NASA/CNES</td>
<td>2001–present</td>
<td>1,336 km</td>
<td>9.92 day repeat</td>
<td>~2 cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>non–Sun-synchronous</td>
<td></td>
</tr>
<tr>
<td>Poseidon-3 on Jason-2</td>
<td>NOAA/NASA/CNES</td>
<td>June 2008–present</td>
<td>1,336 km</td>
<td>9.92 day repeat</td>
<td>~2 cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>non–Sun-synchronous</td>
<td></td>
</tr>
<tr>
<td>Radar altimeter on ERS-1</td>
<td>ESA</td>
<td>1991–2000</td>
<td>780 km</td>
<td>3 &amp; 35 day repeat</td>
<td>~5–6 cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Sun-synchronous(RA)</td>
<td></td>
</tr>
<tr>
<td>RA on ERS-2</td>
<td>ESA</td>
<td>1995–2003</td>
<td>780 km</td>
<td>35 day repeat</td>
<td>~5–6 cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Sun-synchronous</td>
<td></td>
</tr>
<tr>
<td>RA2 on Envisat</td>
<td>ESA</td>
<td>2002–present</td>
<td>800 km</td>
<td>35 day repeat</td>
<td>3 cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Sun-synchronous</td>
<td></td>
</tr>
<tr>
<td>Geosat</td>
<td>U.S. Navy</td>
<td>1986–1989</td>
<td>800 km</td>
<td>17.05 day repeat</td>
<td>10 cm reanalysis</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Sun-synchronous</td>
<td></td>
</tr>
<tr>
<td>Geosat Follow-on</td>
<td>U.S. Navy</td>
<td>2000–present</td>
<td>880 km</td>
<td>17.05 day repeat</td>
<td>~10 cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Sun-synchronous</td>
<td></td>
</tr>
</tbody>
</table>

- The provider of the data. Is this a recognized agency with a reputation for high standards of quality and a professional reputation to maintain?
- The completeness of the data provided. Are there error estimates and/or confidence flags attached? If the datasets are not in a geographically gridded form can the latitude and longitude of every pixel be readily identified?
- Is the data format readily accessible? Normally the attachment of ancillary information on accuracy, location, etc. in addition to the primary data files requires the use of a structured but flexible format such as NetCDF or HDF for which data viewers are widely available. If the format is a proprietary one used by a particular agency then are data viewers available?
- Is there adequate information to be able to interpret the data? For example, are the units and the scale of the data clear—these should be included among the attributes in a NetCDF or HDF file.
- Is the process for generating ocean data products transparent? This means that it should be clear what algorithm or procedure has been used to generate the products. Ideally there should be a link with an algorithm theoretical basis document (ATBD) for each product. For level 3 data the rules for constructing the composite should be available, and for level 4 there should be a clear
Table 2.8. Recent and current satellite synthetic aperture radars.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>ERS-1 SAR</th>
<th>Radarsat</th>
<th>ERS-2 SAR</th>
<th>Envisat ASAR</th>
<th>Radarsat-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Agency</td>
<td>ESA</td>
<td>CSA</td>
<td>ESA</td>
<td>ESA</td>
<td>CSA</td>
</tr>
<tr>
<td>Altitude, km</td>
<td>780</td>
<td>800</td>
<td>780</td>
<td>700</td>
<td>800</td>
</tr>
<tr>
<td>Radar band</td>
<td>C</td>
<td>C</td>
<td>C</td>
<td>C</td>
<td>C</td>
</tr>
<tr>
<td>Polarization</td>
<td>VV</td>
<td>HH</td>
<td>VV</td>
<td>HH, VV</td>
<td>Multiple</td>
</tr>
<tr>
<td>Wavelength, cm</td>
<td>5.7</td>
<td>5.7</td>
<td>5.7</td>
<td>5.7</td>
<td>5.5</td>
</tr>
<tr>
<td>Incidence angle, deg</td>
<td>23°</td>
<td>20–50°</td>
<td>23°</td>
<td>17–50°</td>
<td>10–60°</td>
</tr>
<tr>
<td>Resolution, m</td>
<td>25</td>
<td>10–100</td>
<td>25</td>
<td>25–1,000</td>
<td>3–100</td>
</tr>
<tr>
<td>Swath, km</td>
<td>100</td>
<td>10–500</td>
<td>100</td>
<td>100–400</td>
<td>20–500</td>
</tr>
</tbody>
</table>

Table 2.9. Recent and current satellite scatterometers measuring wind speed and direction.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>AMI</th>
<th>NSCAT</th>
<th>SeaWinds</th>
<th>ASCAT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Agency</td>
<td>ESA</td>
<td>NASA/JAXA</td>
<td>NASA/JAXA</td>
<td>ESA/Eumetsat</td>
</tr>
<tr>
<td>Satellite(s)</td>
<td>ERS-1, ERS-2</td>
<td>ADEOS-1</td>
<td>QuikScat, Midori-2</td>
<td>METOP</td>
</tr>
<tr>
<td>Altitude, km</td>
<td>780</td>
<td>805</td>
<td>803</td>
<td>837</td>
</tr>
<tr>
<td>Radar band</td>
<td>C</td>
<td>Ku</td>
<td>Ku</td>
<td>C</td>
</tr>
<tr>
<td>Polarization</td>
<td>VV</td>
<td>VV and HH</td>
<td>VV and HH</td>
<td>VV</td>
</tr>
<tr>
<td>Frequency, GHz</td>
<td>5.3</td>
<td>13.995</td>
<td>13.4</td>
<td>5.255</td>
</tr>
<tr>
<td>Mode of operation</td>
<td>3 beams on one side</td>
<td>3 beams on both sides</td>
<td>Twin rotating beams</td>
<td>3 beams on both sides</td>
</tr>
<tr>
<td>Swath, km</td>
<td>500 km</td>
<td>2 × 600 km separated by 400 km</td>
<td>1,800 km includes 2 prime swaths of 450 km</td>
<td>2 × 500 km</td>
</tr>
<tr>
<td>Resolution, km</td>
<td>45 km</td>
<td>25–50 km</td>
<td>25 km</td>
<td>25 km</td>
</tr>
</tbody>
</table>
Table 2.10. Access to useful sources of satellite-derived ocean data products from the Web, in binary digital form.

<table>
<thead>
<tr>
<th>Ocean data product</th>
<th>Agency</th>
<th>Internet URL</th>
<th>For access to data select</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Sea surface temperature</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>From several sensors including AVHRR, GOES</td>
<td>NASA-JPL</td>
<td><a href="http://podaac.jpl.nasa.gov">http://podaac.jpl.nasa.gov</a></td>
<td>Tools &amp; Services/ftp Tools &amp; Services/POET</td>
</tr>
<tr>
<td>From AMSR-E, TMI microwave radiometry</td>
<td>REMSS</td>
<td><a href="http://www.ssmi.com/">http://www.ssmi.com/</a></td>
<td>Select AMSR or TMI Browse Data. Then select FTP or Download</td>
</tr>
<tr>
<td>Pathfinder (from AVHRR)</td>
<td>NOAA</td>
<td><a href="http://www.nodc.noaa.gov/sog/pathfinder4km/">http://www.nodc.noaa.gov/sog/pathfinder4km/</a></td>
<td>Select “Available Data”</td>
</tr>
<tr>
<td><strong>Ocean color and related products</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MERIS</td>
<td>Global gridded</td>
<td><a href="http://envisat.esa.int/level3/meris/">http://envisat.esa.int/level3/meris/</a></td>
<td>Select year from products table</td>
</tr>
<tr>
<td>MODIS, SeaWiFS, etc.</td>
<td>Global 4 and 9 km</td>
<td><a href="http://oceancolor.gsfc.nasa.gov/">http://oceancolor.gsfc.nasa.gov/</a></td>
<td>Level 3 Browser</td>
</tr>
<tr>
<td>MODIS, SeaWiFs etc</td>
<td>Local full resolution</td>
<td><a href="http://oceancolor.gsfc.nasa.gov/">http://oceancolor.gsfc.nasa.gov/</a></td>
<td>Level 1 &amp; 2 Browser</td>
</tr>
</tbody>
</table>
### Altimetry: surface height and related products

<table>
<thead>
<tr>
<th>Product</th>
<th>Source</th>
<th>URL</th>
<th>Additional Information</th>
</tr>
</thead>
<tbody>
<tr>
<td>SSHA</td>
<td>PO-DAAC (JPL)</td>
<td><a href="http://podaac.jpl.nasa.gov">http://podaac.jpl.nasa.gov</a></td>
<td>Tools &amp; Services/ftp</td>
</tr>
</tbody>
</table>

### Winds over the sea

<table>
<thead>
<tr>
<th>Product</th>
<th>Source</th>
<th>URL</th>
<th>Additional Information</th>
</tr>
</thead>
<tbody>
<tr>
<td>SSMI, Q-Scat, AMSR, TMI</td>
<td>REMSS</td>
<td><a href="http://www.ssmi.com/">http://www.ssmi.com/</a></td>
<td>Select required sensor</td>
</tr>
</tbody>
</table>
Table 2.11. Access to useful image data-viewing and manipulation tools.

<table>
<thead>
<tr>
<th>Name</th>
<th>Description</th>
<th>Sponsor</th>
<th>Policy on availability</th>
<th>Website URL</th>
</tr>
</thead>
<tbody>
<tr>
<td>BILKO</td>
<td>Bilko is a complete system for learning and teaching remote-sensing image analysis skills. Current lessons teach the application of remote sensing to oceanography and coastal management. PC-based image analysis software supports the main image data formats</td>
<td>UNESCO</td>
<td>Free to all who register their email address</td>
<td><a href="http://www.unesco.bilko.org">http://www.unesco.bilko.org</a></td>
</tr>
<tr>
<td>BEAM</td>
<td>A software system for viewing, analyzing, and processing remote-sensing data. Originally named as the Basic ERS &amp; Envisat (A)ATSR and MERIS toolbox, BEAM now supports a growing number of other sensors such as MODIS, AVNIR, PRISM, and CHRIS/Proba. Available for several platforms.</td>
<td>ESA/ Brockmann Consult</td>
<td>Free access under the terms of the GNU General Public License</td>
<td><a href="http://141.4.215.13/index.html">http://141.4.215.13/index.html</a></td>
</tr>
<tr>
<td>BEST</td>
<td>BEST (Basic Envisat SAR Toolbox) is a collection of executable software tools designed to handle ESA data products from both the Envisat ASAR instrument and the AMIs (Active Microwave Instruments) on ERS 1&amp;2. The latest version 4.2.0 solves the problem in handling ERS PGS products both in CEOS and in Envisat format, and also has ASAR WSS new functionalities</td>
<td>ESA</td>
<td></td>
<td><a href="http://envisat.esa.int/resources/softwaretools/">http://envisat.esa.int/resources/softwaretools/</a></td>
</tr>
</tbody>
</table>

The methods of satellite oceanography
specification of the different sources of data, with explanations of the rules for prioritizing them, bias adjustments, interpolation procedures, etc.

- Is the version of the dataset unambiguous? Some datasets differ between the near–real time product, which may lack some ancillary inputs, and a “consolidated” product that has been reprocessed once all the necessary information for geolocation and corrections has been acquired. In subsequent years the scientific and technical understanding of the processes may improve so that newer versions of the algorithm are introduced, leading to updated, more reliable, products. Similarly level 3 and 4 products may also be reprocessed or reanalyzed, using the reprocessed level 2 as input. It is vital for some applications that the user knows which version of data is being used.

- Are the products credible? Has there been adequate calibration of the sensor and the processes used to generate the products? Have the products themselves been validated against independent observations of the same ocean variable? Are there peer-reviewed publications which define the sensor calibration and processing algorithms and report the validation of the products?

This amounts to quite an extensive list of issues that need to be considered. Obviously, common sense should be applied. The casual user is keen to have a look at some ocean image data and will be most influenced by how easy it is to gain access to the data. In contrast the climate scientist, whose own reputation for careful analysis of climate trends is at stake, must be able to critically assess the quality of the data and justify her or his assessment by reference to peer-reviewed publications. All who use any type of data for scientific purposes need to consider these issues before committing their work to a dependence on data provided by a third party. They owe it to themselves to understand how the products they use have been generated so that they can ascertain whether the exciting “discoveries” from their analysis are genuine environmental phenomena or disappointingly an artifact of the data-processing chain! Moreover it is the user’s responsibility to keep track of product versions so as to avoid the embarrassment of publishing a paper about a sudden change in an ocean variable that turns out to correspond to a version switch of data products!

As a guide, Table 2.10 identifies by data product type some producers to whom users can look for data that should match up to many of the requirements listed above. Although the stated web addresses may go out of date, the agencies themselves should still be traceable for several years from the publication of this book. Most of these agencies provide data through the Internet at no cost to scientific users. In some cases the data are downloadable instantly. Others allow users to select subsets of data from an archive which are subsequently placed in an ftp site for download. Some agencies require users to register before allowing access to the data, so that they can differentiate between commercial usage that is charged and private scientific use that is not.

For the student new to satellite data on the web, it is worth pointing out the distinction between image datasets and digital pictures. A digital image dataset stores the data in a format that allows a user with suitable software to readily
access the measured ocean properties as digital values that can be extracted and
analyzed. The file format often has provision for ancillary data to be attached to 
the file with the primary image data. It also allows the image to be displayed and 
enhanced without losing track of actual scientific values. Examples of such formats 
are .hdf and .cdf files, or proprietary formats such as ESA’s .N1 data files. In contrast 
a digital picture file format such as .jpg, .png, .gif, .bmp contains digital values 
allowing a picture of the image data to be reproduced, but in general there is no 
way of reaching the true scientific values behind the picture. Such pictures may be 
satisfactory for illustrating a particular phenomenon, in which case they must be 
printed with the correct color scale, but cannot form the basis of any scientific 
analysis or manipulation. Some of the agencies that provide ocean data products 
have websites where images can be browsed at low or full resolution, and allow 
download by clicking on the pictures. However, this typically saves the files as a 
picture format. Scientific users need to explicitly download a .cdf or .hdf version if 
they wish to do more than simply copy the picture.

There are now a number of useful image display and image-processing 
software systems available for enhancing image data and performing more elaborate 
analytical procedures. Table 2.11 lists some of these. The BILKO system is a general 
purpose image analysis system for PCs which was developed by UNESCO as a basis 
for training in satellite data analysis. It is freely available and comes with tutorials 
that allow the user to discover its capabilities through examples of analyzing a 
variety of typical satellite ocean data products. It can handle standard image data 
formats like NetCDF and HDF, and has been extended to read ESA’s N1 file 
formats. The BEAM software has been developed specifically to allow scientific 
manipulation of ESA’s data products. Other agencies are also starting to provide 
software tools for selecting, enhancing, and manipulating their data products, 
sometimes online before the products are downloaded.

2.7 REFERENCES

100, 8777–8790.
In: L.-L. Fu and A. Cazenave (Eds.), Satellite Altimetry and Earth Sciences (pp. 1–131). 
Cotton, P. D., and D. J. T. Carter (1994), Cross calibration of TOPEX ERS-1 and Geosat wave 
Data Assimilation Experiment (GODAE) High Resolution Sea Surface Temperature Pilot 
Project (GHRSSST-PP), Bull. Am. Meteorol. Soc., 88(8), 1197–1213, doi: 10.1175/BAMS-
88-8-1197.
Gower, J. F. R. (1996), Intercalibration of wave and winds data from TOPEX/Poseidon and 


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