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Oceanography

1 INTRODUCTION

At the onset of this new millennium increased awareness of the stresses being placed on the Earth system, often induced by human activities, has intensified the need for information on the present state of the Earth system and for enhanced capability to assess its evolution such as associated with environmental pollution, natural resource management, sustainable development, and global climate change.

This realisation has resulted in increased political and legal obligations on governments and on national and regional agencies to address Earth system topics of global concern. These obligations are often encapsulated within international treaties, whose signatories have explicit requirements placed upon them.

Many of these treaties call for systematic observations of the Earth to increase our understanding of its processes and our ability to monitor them:

– The UN Framework Convention on Climate Change (FCCC).
– The UN Convention to Combat Desertification in those Countries experiencing Serious Drought and/or Desertification.
– The Montreal Protocol of the Vienna Convention on the Protection of the Ozone Layer.
– Agenda 21 and the UN Commission on Sustainable Development.
– The Intergovernmental Panel on Climate Change.

These commitments require substantial economic, technical and scientific resources for their execution, and action at many levels, including significant programmes of global observations. In this context it is recognised that Earth observation satellites provide an important and unique source of information. The most well established international forum for coordinating the operational provision of data is without doubt the World Weather Watch (WWW) of the World Meteorological Organisation (WMO). Another prominent, though less established forum is the Global Ocean Observing System (GOOS). Common for these international forum and observing system is the role of satellite Earth Observation. Here it is worth mentioning the World Climate Research Program (WCRP), International Geopshere and Biosphere Program (IGBP) and Intergovermental Oceanographic Committee (IOC) in which research projects and observing systems highlight the importance of continuous and regular access to Earth Observation data.

Earth observations from satellite are highly complementary to those collected by *in-situ* systems. Whereas *in-situ* measurements are necessary for underwater observations, for high accuracy local observations, for the calibration of observations made by satellite and as input to numerical models, satellite observations provide an inherent wide area unique capability to obtain regular quantitative information of surface variables and upper layer phenomena at global, regional and local scales.

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Johnny A. Johannessen, Stein Sandven and Dominique Durand. The Century of Space Science, 1585–1622

Present-day applications of satellite data are widespread and cover research, operational and commercial activities. These activities are of interest in the global context and in the regional, national, and local context where Earth observation data are successfully applied in support of a range of different sectors, including (not exclusive):

– climate change research,
– stratospheric chemistry, particularly related to the ozone hole,
– weather forecasts based on Numerical Weather Prediction (NWP),
– agriculture and forestry services,
– resource mapping,
– hazard monitoring and disaster assessment,
– sea ice monitoring,
– coastal zone management,
– oceanographic applications.

Details of the outstanding scientific advances made possible by satellite observations of the ocean and the associated societal benefits are provided by Halpern (2000).

The number of Earth Observing satellites are growing rapidly for both scientific research and operational application within fields of land, atmosphere, marine meteorology and oceanography including sea ice covered regions. International investment in satellite platforms, instruments and associated ground segments is already substantial, and more investment is planned over the coming decade. There are currently over 45 missions operating, and around 70 more missions, carrying over 230 instruments, planned for operation during the next 15 years by the world's civil space agencies (CEOS 97). In addition, Space Agencies are currently implementing their new strategy for defining future Earth Observing satellites, both dedicated research missions and continuous (operational) monitoring missions.

In this article we will review the status of satellite oceanography at the onset of the new millennium. In Section 2 the principal methods, instruments and basic measured surface parameters are addressed. Examples of contribution to climate monitoring and operational oceanography are then given in Section 3 and Section 4, respectively. In Section 5 an outlook towards the near future satellite observing system is provided followed by a summary in Section 6.

2 PRINCIPAL METHODS, INSTRUMENTS AND SURFACE CHARACTERISTICS

Satellite oceanography is primarily using three domains within the electromagnetic (EM) spectrum, notably radiation in: – the visible/near-infrared (VNIR); – the thermal infrared (TIR); – and the microwave bands of the EM spectrum (Figure 1).

The visible and infrared channels utilize intervals of the EM spectrum with high atmospheric transmission, such as in the bands from 0.4–2.5 µm, 3.5–4.0 µm and 10–13 µm (Figure 1, top). The EM waves in these bands do not generally penetrate clouds, so remote-sensing observations of the Earth's surface in these bands can only be done satisfactorily under cloud free conditions. As such this is posing severe limitations in regions where clouds are frequently present. In the microwave area, on the other hand, at

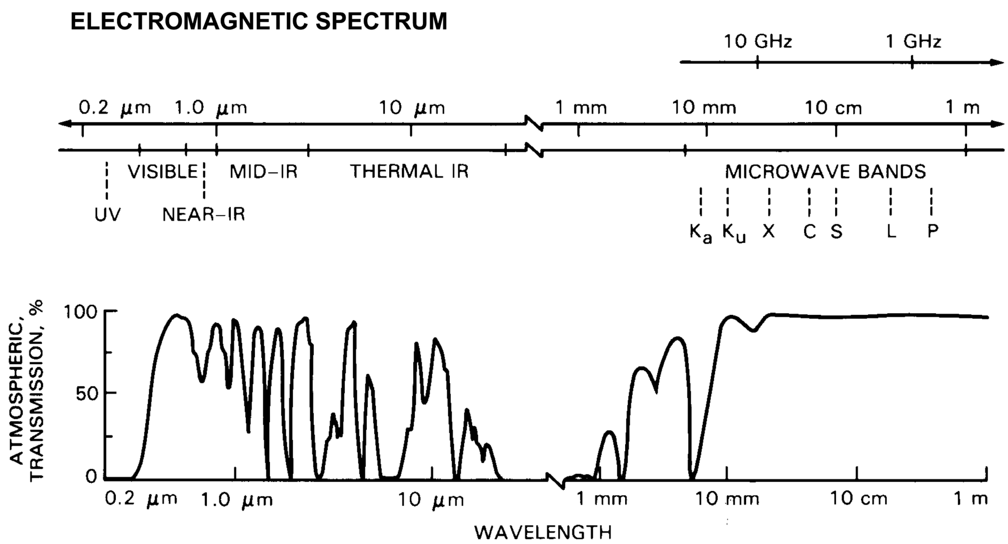

Figure 1 The electromagnetic (EM) spectrum showing the bands used in remote sensing together with the operating area for some sensors (upper graph). The atmospheric transmission of the EM spectrum is shown in the lower graph. Note that the operating areas for satellite oceanography are located in parts of the spectrum where atmospheric transmission is high.

wavelengths above 0.3 cm, EM waves generally penetrate clouds which makes it feasible to obtain regular, daily observations of ocean and sea ice surfaces (Figure 1, bottom). These characteristic spectral domains are further addressed in the following.

2.1 Visible/near infrared

The basic quantity observed in the VNIR domain is the albedo or alternatively the fraction of the incident sunlight that has been scattered and/or reflected in the atmosphere/ocean system. The incident solar radiation undergoes a number of interactions (absorption and scattering) with molecules and particles in the atmosphere and in the water, in addition to the reflection that occurs at the air – sea interface.

Only a fraction of the incident radiation penetrates the water body. Absorption by water molecules becomes critical at wavelength greater than 700 nm. Therefore water appears black at such wavelength, except when a high load of suspended particles is present near the surface. The visible light (400–700 nm) may propagate in the water medium and interact with water molecules, organic and inorganic particles in suspension, dissolved optically active substances, and possibly the sea-floor in optically shallow-waters. The penetration depth depends upon the wavelength and the water column absorption properties. Only a small fraction of the visible-light spectrum is scattered upwards to the surface giving rise to the so-called water-leaving radiance, L_w. This is expressed as:

$$L_w = (L_s - L_a - \alpha L_r)/\alpha \tag{2.1}$$

where L_s is the radiance reaching the sensor, L_a is the atmospheric radiance, L_r is reflected radiation from the sea surface and α is the atmospheric diffuse transmittance. In most oceanic waters, L_w represents less than 10% of the total signal measured by a spaceborne sensor. Typically, 90% of radiation has been scattered in the atmosphere, without interaction with subsurface waters. Therefore any quantitative estimation of water-column optically active constituents depends upon an adequate retrieval of the water-leaving radiance L_w emanating from the water column. This requires reliable correction of the remotely sensed signal scattered in the atmosphere and reflected at the air–sea interface (Gordon, 1997).

Furthermore the capability of deriving the accurate quantity of a particular water quality parameter depends upon the complexity of the water column in terms of number and properties of components that interact with the electromagnetic signal. According to the optical complexity of the water body, two types of water are defined, namely case I and case II waters (Morel and Prieur, 1977).

Case I waters are natural open-ocean water bodies, for which water-leaving radiance measured by remote sensors are only dependent on chlorophyll pigment concentration. In such waters, which represent about 90% of the World ocean, the variation in the color of the upper water column could be related to the variation of the concentration of chlorophyll pigments contained in phytoplankton cells. The color of oceanic case I waters shifts from deep blue in oligotrophic waters (very low chlorophyll concentration) to dark green in eutrophic waters (high concentration). This shift results from the strong absorption by algae pigments (chlorophyll pigments and carotenoids) in the blue part of the visible spectrum, with a maximum around 445 nm, compared with the weak absorption around 550–580 nm (Morel, 1998). In case I type water, quite robust empirical relationships can be derived, linking the chlorophyll concentration to the ratio of water leaving radiance (and/or the reflectance of the sea) at these wavelengths.

Case II waters are more optically complex. In such waters the satellite derived water-leaving radiance, in addition to being dependent upon chlorophyll pigments and derived products, is also sensitive and modified by at least one other optically-active component, e.g., suspended sediment and/or colored dissolved organic matter. Most coastal waters are classified as case II waters. Furthermore the various optically active components do not display typical linear relationship. Therefore simple empirical models may no longer be used, and more sophisticated approach, such as inverse modeling must be considered.

2.2 Thermal infrared

For TIR (and PMW) the measured quantity is the emitted energy as function of surface temperature and emissivity. The emissivity is a dimensionless coefficient, e, and can be computed from the complex dielectric constant (or the relative permettivity) $e = e' - ie''$, which characterizes the electrical properties of the media. e' is referred to as the dielectric constant and e'' as the dielectric loss factor. Alternatively, e can be estimated from the complex index of refraction such as $n^2 = e$.

In the thermal infrared part of the spectrum, the surface signal expressed as the radiance observed by remote sensing can be used as input to Planck's law of radiation to find the sea surface temperature (SST) if the emissivity e of the surface is known. For water, the value of e in the most used thermal spectral band of 10 μm–12 μm is very high and stable, about 0.99. At a given wavelength λ the blackbody radiance, L, can be expressed as:

$$L = (2\pi hc^2/\lambda^5)/(hc/e^{\lambda kSST} - 1) \tag{2.2}$$

The Planck's law is usually expanded in a Taylor series from which the linear term is maintained:

$$Ln(L) = aT_b^{-1} + b \tag{2.3}$$

where T_b is the brightness temperature in the TIR part of the spectrum, and the coefficients a, b are constant values for each spectral band.

2.3 Passive microwaves

In the microwave domain the brightness temperature T_b provides the measurement of the microwave emission from the surface. The brightness temperature is defined by the real surface temperature T_s and the emissivity e by the relation

$$T_b = T_s * e \tag{2.4}$$

Spatial variations in T_b observed over the surface of the Earth are due primarily to variations in the emissivity of the surface material and secondly to variations in surface temperature. For the most frequently used frequencies between about 6 GHz and 90 GHz, the emissivity of ice, snow and water show large variations allowing observations of a wide range of multidisciplinary parameters spanning land, ocean, cryosphere and atmosphere.

While e for calm water can be calculated quite accurately from the electric properties (Stogryn, 1971), the value and variation of e for the various forms of ice and snow is less accurately known, and therefore often has to be empirically measured. For sea ice, the dielectric constant e' is relatively constant with frequency above 1 MHz, but e'' is not. There is a minimum in e'' at 3–8 GHz with higher value for lower and higher frequencies.

For firstyear ice at 283 K and 8‰ salinity, the minimum e'' is approximately 0.3. As temperature decreases e'' will increase because precipitated salt will go back into solution. Furthermore, e'' will decrease with decreasing salt content. Multiyear ice has a lower e'' than firstyear ice and its temperature dependence is weaker. Thus, microwave radiation penetrates deeper into multiyear ice than firstyear ice.

2.4 Active microwaves

Radar instruments provide their own source of illumination in the microwave portion of the EM spectrum, at wavelengths on the order of 10^4 longer than those in the visible part of the spectrum (Figure 1). Because of this, radars can operate independent of solar illumination, cloud cover and precipitation conditions.

All radar measurements can be described by a basic equation, which relates transmitted power, distance, reflectivity and antenna characteristics. The equation can be formulated as

$$P_R = \frac{P_t}{4\pi R^2} G \frac{\sigma}{4\pi R^2} A \tag{2.5}$$

where P_R is power received, P_t is power transmitted, G is the gain of the antenna, σ is the radar cross section, and A is antenna area. The energy of the outward propagating wave, which is spherically expanding, is given by the first ratio. This spherically expanding wave is focused down to an angular beamwidth by the antenna so that the fluxes becomes higher by a factor of G over that of a spherically expanding wave. The focused energy impinges on an object which has a radar cross section σ, which is defined as the equivalent of a perfectly reflecting object of a given area which reflects isotropically (spherically) as shown by the second ratio. Finally, the antenna area, A, term intercepts a portion of the reflected wave so that this portion of the flux defines the power received by the antenna.

The basic radar equation is general; it can be applied to any object of any shape or composition. For imaging over areas of terrain or ocean, a reflection coefficient is defined, σ_0, which is the radar cross section, σ, per unit area. The radar equation (2.5) can then be expressed as

$$P_R = \frac{\lambda^2}{(4\pi)^3} \int \frac{P_t G^2 \sigma_0}{R^4} dA \tag{2.6}$$

The averaged received power for a radar can then be determined by examining the integral radar equation for distributed targets. The radar scattering coefficient, σ_0, also called backscatter coefficient, expresses a measure (usually in dB units) of the energy scattered back towards the antenna. It is a function of frequency, incidence angle, polarization and the scattering characteristics of the illuminated area.

Radar frequencies are identified by letter designations, and the most commonly used are K-band (30 GHz, 1 cm), X-band (9.4 GHz, 3.2 cm), C-band (5.3 GHz, 5.7 cm), L-band (1.25 GHz, 23.5 cm) and P-band (450 MHz, 62 cm). At these wavelengths the EM-waves are not appreciably attenuated by clouds, precipitation or the Earth's atmosphere (see Figure 1). Therefore, good quality radar data can be obtained during all kind of weather and light conditions. The three main classes of satellite radars (altimeter, synthetic aperture radar, and scatterometer) are further described in the following sections.

2.4.1 Radar altimeter

The radar altimeter measures the transit time and backscatter power of individual transmitted pulses. The transit time is proportional to the satellite's altitude above the ocean, land or ice surfaces. The pulse propagates toward the surface at time t_1 with a speed of light c, is backscattered by the surface, and an echo is received by the sensor at a time t_2. The time difference $t_d = t_2 - t_1$ is equal to the round trip distance to the reflecting surface divided by the propagation speed

$$t_d = \frac{2h}{c} \tag{2.7}$$

Springer

http://www.springer.com/978-0-7923-7196-0

The Century of Space Science
Bleeker, J.A.; Geiss, J.; Huber, M. (Eds.)
2001, XLIX, 1846 p., Hardcover
ISBN: 978-0-7923-7196-0