Chapter 5

50 Years of Satellite Remote Sensing of the Ocean

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ABSTRACT

The development of the technologies of remote sensing of the ocean was initiated in the 1970s, while the ideas of observing the ocean from space were conceived in the late 1960s. The first global view from space revealed the expanse and complexity of the state of the ocean that had perplexed and inspired oceanographers ever since. This paper presents a glimpse of the vast progress made from ocean remote sensing in the past 50 years that has a profound impact on the ways we study the ocean in relation to weather and climate. The new view from space in conjunction with the deployment of an unprecedented amount of in situ observations of the ocean has led to a revolution in physical oceanography. The highlights of the achievement include the description and understanding of the global ocean circulation, the air–sea fluxes driving the coupled ocean–atmosphere system that is most prominently illustrated in the tropical oceans. The polar oceans are most sensitive to climate change with significant consequences, but owing to remoteness they were not accessible until the space age. Fundamental discoveries have been made on the evolution of the state of sea ice as well as the circulation of the ice-covered ocean. Many surprises emerged from the extraordinary accuracy and expanse of the space observations. Notable examples include the determination of the global mean sea level rise as well as the role of the deep ocean in tidal mixing and dissipation.

1. Introduction

The development of the technologies of satellite remote sensing of the ocean began in the 1970s. This paper reviews the progress made in the past 50 years. The goal of this review is not to provide an exhaustive account of all aspects of ocean remote sensing, but to focus on the progress enabled by remote sensing in the advancement of our understanding of the ocean circulation and ocean–atmosphere interaction in relation to weather and climate, a central theme of AMS. Given the massive amount of progress in the field, the materials covered in the review are inevitably subjective and eclectic.

Weather and climate are governed by the processes of coupled ocean–atmosphere interaction. Ocean is the major reservoir of water, heat, and greenhouse gases on Earth. For a comprehensive account of the history of the understanding of ocean circulation, the reader is referred to the chapter by Wunsch and Ferrari (2019, in this volume). On global scales, the ocean has absorbed more than 90% of the heat from global warming since the industrial revolution (Levitus et al. 2009). The calculations of Earth’s radiation balance suggest that Earth’s surface temperature would reach 67°C without the ocean (Philander 2004). The ocean is also an integral part of the global water cycle, playing key roles in regulating the hydrological cycles on Earth and affecting water availability to human society. On basin scales, the ocean plays significant roles in climate variability phenomena such as El Niño–Southern Oscillation (ENSO) that have regional to global ramifications. On more regional scales, the ocean provides heat and water to fuel storms over the ocean, the study of which is essential to understanding severe weather like hurricanes and nor’easters. Yet the observations of the global ocean have been challenging because of its vast domain, posing great difficulties in sampling its ever-changing properties on all spatial and temporal scales.

The beginning of the space age in the 1950s–60s marked a transition of observing the ocean from an in situ,
slow, shipboard effort to remote sensing in orbit to sample the global ocean in a short period of time. The era of satellite oceanography began taking shape in the 1970s after a seminal conference convened at Williamstown, Massachusetts (Kaula 1969), outlining many key concepts of remote sensing of Earth from orbiting satellites, including oceanography and geodesy. The first measurement of sea surface temperature (SST) from space was made in the late 1960s by the TIROS weather satellite program. A notable discovery from the early infrared observations of SST made by weather satellites was the tropical instability waves in the Pacific Ocean (Legeckis 1977), illustrating the power of satellite observations of large-scale oceanographic phenomena. Although the infrared sensors have been improved by the Advanced Very High Resolution Radiometer (AVHRR) Program since the late 1970s, their limitation by cloud covers is a major issue for continuous monitoring of oceanic processes. Nevertheless, blended SST products based on AVHRR, passive microwave sensors (all-weather capability), and in situ measurements (e.g., Reynolds et al. 2007) have made significant contributions to ocean and climate research.

The technologies for dedicated ocean observations were first demonstrated by the Skylab missions (Krishen 1975), using microwave sensors with all-weather observing capability. This mission demonstrated the feasibility of radar altimetry, scatterometry, and microwave radiometry, laying the foundation for the follow-on missions to observe the ocean from space. Following Skylab’s demonstration of measuring the shape of sea surface with radar altimetry, the Geodetic Earth Orbiting Satellite-3 (GEOS-3) satellite provided the first survey of the relief of ocean surface reflecting Earth’s gravity anomalies and surface geostrophic currents (Huang et al. 1978).

The first satellite dedicated to observing the ocean using a comprehensive suite of sensors was the Seasat mission (Born et al. 1979). Although the mission lasted only slightly over a period of 100 days, it accomplished its goal to prove the concept of satellite oceanography to observe the global ocean with synoptic frequency. The primary impact of Seasat was microwave remote sensing with radar instruments: radar altimeter, scatterometer, and synthetic aperture radar (SAR). These measurements showed the feasibility of mapping the ocean surface geostrophic circulation, vector wind field, surface and internal gravity waves, sea ice, and a host of air-sea interaction processes. The success of Seasat had led the way to today’s capability of observing the global ocean with active sensors.

In addition to radar active remote sensing, another important development was passive microwave radiometry and advanced visible and infrared sensors for measuring ocean surface properties like temperature, salinity, wind speed, and ocean color (in relation to biological processes). The Scanning Multifrequency Microwave Radiometer (SMMR) carried by Seasat as well as Nimbus-7, which also carried the Coastal Zone Color Scanner (CZCS), provided the basis for the development of all-weather passive remote sensing in the ensuing decades.

Before the advent of satellite remote sensing, one had to resort to sparse data collected in different years and decades to reach a broad-brush view of the global ocean that is hardly sufficient to gain quantitative understanding of its physics for addressing questions on weather and climate. Arguably, the most powerful impact of Seasat was the demonstration of the feasibility of mapping the variability of the sea surface height (SSH) field globally from space in a matter of weeks, using the radar altimeter. Simultaneously, the surface vector wind driving the ocean current was also measured with the scatterometer. The microwave radiometer measured SST, wind speed, and columnar water vapor that allowed the later development of a methodology to estimate evaporation and latent heat (Liu 1986; Liu et al. 2000). The new capability of Seasat had fundamentally changed the paradigm of the study of the global oceans.

The first wave of dedicated ocean observations from space was a series of satellite radar altimeters launched by various international agencies: Geosat (U.S. Navy), ERS-1 [European Space Agency (ESA)], and TOPEX/Poseidon [NASA and the French Centre National d’Etudes Spatiales (CNES)] with ever increasing precision and accuracy (Fu 2001; Fu and Cazenave 2001). This series of measurement was continued by the Franco-American Jason-1, Jason-2, and Jason-3, as well as by ESA’s Environmental Satellite (Envisat), U.S. Navy’s Geosat Follow-On, the France-India collaborative mission of AltiKa, and ESA’s CryoSat (Fu and Morrow 2013), leading to a continuous record of ocean surface topography since 1992. The new capability of satellite oceanography underscored the timeliness of the development of international programs of global observation of the ocean and atmosphere in the 1990s (see Davis et al. 2019, this volume), the World Ocean Circulation Experiment (WOCE; Siedler et al. 2001), as well as the Tropical Ocean and Global Atmosphere (TOGA; McPhaden et al. 1998), leading to today’s Climate and Ocean: Variability, Predictability and Change (CLIVAR; http://www.clivar.org/) Program. The combined efforts of satellite remote sensing and in situ observing systems marked the beginning of the era of well-sampled global ocean and atmosphere.

Highly complementary to radar altimetry observations was the development of the global international in situ Argo Program (Roemmich et al. 2009) and the U.S.–Germany Gravity Recovery and Climate Experiment (GRACE) mission (Tapley et al. 2004). Beginning
in the early 2000s, Argo has evolved into an international program that has deployed nearly 4000 floats in the global ocean sampling the temperature and salinity of the upper 2000 m of the water column. This measurement, coupled with ocean surface topography from altimetry, has provided essential information on the geostrophic circulation of the world’s oceans, as well as the density field of the ocean.

Recognized in the Williamstown conference as well as the planning for the TOPEX/Poseidon mission in the early 1980s, there was a fundamental need for the information of the geoid, which is a surface of constant gravitational potential, to determine the geostrophic velocity of ocean circulation from radar altimetry measurement of the sea surface height. After a long lead development to overcome the technological challenges, the GRACE mission was launched in 2002 for the measurement of Earth’s time-varying gravity field. GRACE has not only provided the geoid for determining the ocean circulation but also the ever-changing mass of water on Earth with applications to oceanography, geodesy, and hydrology. The combined system of altimetry, Argo, and GRACE has allowed the study of global sea level change caused by climate variability. This ocean observing system has established a framework to assess, understand, and possibly predict the sea level change in the future.

Seasat has also inspired follow-on missions to measure the vector wind field. Most notable was the QuikSCAT mission launched in 1999. It provided the first decade-long high-resolution wind field. While the scatterometer was originally designed as a wind sensor, the backscatter it measured had long been recognized to be more closely related to the ocean surface stress that drives ocean circulation. Oceanographers derive stress mainly from wind through an empirically determined drag coefficient that is often problematic. The unique stress-measuring capability of the scatterometer has shed new light to the atmospheric forcing of ocean at the mesoscale, particularly under the strong winds of tropical cyclones where stress does not increase with wind as expected, and over ocean eddies, where current rotation and temperature feedbacks strongly drive the vorticity and surface divergence (Chelton et al. 2004; Liu and Xie 2017).

In recent years, a new frontier of satellite oceanography has been establishing to measure sea surface salinity (SSS) of the global ocean using L-band radiometry [the Soil Moisture Ocean Salinity (SMOS), Aquarius, and Soil Moisture Active Passive (SMAP) missions]. The synergy of different satellite measurements such as wind, SSH, SST, SSS, and ocean color observations has led to advances in the understanding of ocean–atmosphere interaction in relation to climate, ocean dynamics, and marine biogeochemistry (e.g., Lee et al. 2012).

Through decades of technological advancement, the integrated satellite observing systems for the ocean are capable of monitoring the global ocean and air–sea interaction processes not only on the ocean basin scales but on the mesoscales as well. Ocean observing satellites, together with in situ ocean observing platforms and observationally constrained ocean state estimation systems, are providing quantitative characterization of the four-dimensional ocean circulation, its roles on climate and weather processes, its linkages with the global water cycle, and its impacts on ocean biogeochemistry. These observations are also playing a pivotal role in improving climate and weather predictions.

In the following, the paper first discusses the contributions of satellite remote sensing to the modern description and understanding of the ocean circulation, followed by the air–sea fluxes driving the coupled ocean–atmospheric systems, which are most prominently illustrated in the topical oceans. The new findings in the polar oceans are discussed in terms of sea ice dynamics and the circulation of the ocean under ice cover. The paper ends with two important, originally unexpected breakthroughs: the determination of the global mean sea level rise and the ocean internal tides and their role in tidal dissipation and the mixing of the deep ocean.

2. Ocean circulation

The movement of ocean water, transporting mass, heat, energy, momentum, nutrients, and chemical properties, determines to a large extent the state of the hydrosphere of Earth consisting of water, air, and ice. Before the 1960s, when modern electronics enabled the collection of long time series of ocean currents, the notion of large-scale ocean circulation was that of a steady laminar fluid. The discovery of the rapidly changing currents associated with the oceanic mesoscale eddies called for the need to map ocean currents with sufficient spatial and temporal resolutions. Oceanographers were impressed by the first global map of the variability of the SSH derived from only one-month’s altimetry data from Seasat (Cheney et al. 1983; see section 2c). Because of their large signals and short scales, the oceanic mesoscale eddies were the first features of ocean circulation detected by satellite observations (see section 2c).

The large-scale ocean circulation, including its variability and mean state, is more difficult to observe from space. For example, the determination of large-scale SSH change requires precision orbit determination that has taken a long period of time to develop. The advancement of satellite altimetry from 1980s to the
present has advanced the knowledge of the global ocean circulation from the mesoscale to the basin scale and revolutionized the field of physical oceanography.

### a. Ocean general circulation

In the open ocean away from the equator, the time-averaged state of the large-scale ocean circulation is in geostrophic balance, the Coriolis force exerted on the moving fluid is balanced by the horizontal pressure gradient as shown by the following equations:

\[
f_u = -\frac{1}{\rho} \frac{\partial p}{\partial y}, \tag{5-2}
\]

\[
f_v = \frac{1}{\rho} \frac{\partial p}{\partial x}, \tag{5-1}
\]

\[
0 = -\frac{\partial p}{\partial z} - g \rho, \tag{5-3}
\]

where \( p \) is pressure; \( \rho \) is the density of seawater; \( g \) is Earth’s gravity acceleration; \( u \) and \( v \) are the zonal and meridional velocity components, respectively; \( f \) is the Coriolis parameter defined as \( f = 2\Omega \sin \phi \), where \( \Omega \) is Earth’s rotation rate \( (7.292 \times 10^{-5} \text{ rad s}^{-1}) \); and \( \phi \) is the latitude. This balance allows the computation of the velocity at the surface relative to a deep level from the horizontal pressure gradient as shown by the following equation:

\[
\rho u = -\frac{g}{f} \int_{z_o}^z \frac{\partial p}{\partial z} dz + u_o, \tag{5-4}
\]

where \( z_o \) is a reference level for the integration and \( u_o \) is the meridional velocity at the reference level. A similar equation can be obtained for \( u \). This “dynamic method” based on the “thermal wind” equation has been the approach to the determination of the ocean general circulation from shipboard measurements of the ocean density field (see Wunsch and Ferrari 2019, this volume). However, the determination of the velocity at the reference level has been an elusive goal. The idea of making measurement of the height of sea surface from space plus independent determination of the geoid for determining the surface ocean general circulation was part of the original rationale for developing satellite altimetry and gravimetry. Once the surface geostrophic velocity is known then the geostrophic velocity at deep levels can be determined from Eq. (5-4).

The challenge of determining the ocean general circulation from space is the accuracy of orbit determination and the gravity field. The first satellite mission that had the required accuracies was TOPEX/Poseidon (T/P; Fu et al. 1994). Shown in Fig. 5-1 (top) [adapted from Rapp et al. (1996)] is the ocean surface topography, defined as the difference between the T/P SSH and the geoid from the Joint Gravity Model (Nerem et al. 1994), expressed by the orthonormal expansion to degree 14, corresponding to a wavelength of approximately 2800 km. This represents a significant step of direct observation of the global ocean general circulation from space, as represented by the horizontal gradient of the ocean surface topography.

The technique of orbit determination and the associated gravity field has been significantly improved since the mid-1990s, especially via the gravity missions of GRACE and Gravity Field and Ocean Circulation Explorer (GOCE, an ESA mission). Shown in Fig. 5-1 (lower panel) (P. Knudsen 2018, personal communication, produced with permission) is the ocean surface topography based on the mean sea surface from a constellation of altimetry satellites and the geoid based on a combination of the gravity fields from the GOCE and GRACE missions (e.g., Fecher et al. 2017). It represents the state of the art of determining the surface ocean general circulation from space. The resolution has been improved by a factor of 100, resolving features of scale less than 30 km. There has been a great deal of evolution in this effort in the past two decades. Earlier versions of space observations were combined with oceanic in situ data to produce global fields with enhanced resolution (e.g., Rio et al. 2011; Maximenko et al. 2009). However, the result of Knudsen was obtained solely from space observations. It reveals a high degree of details: not only the midlatitude gyres and boundary currents, but also the structure of the high-latitude circulation.

### b. Basin-scale variability

It has been known from the dynamics of the planetary-scale fluids that large-scale variability exhibits waves influenced by Earth’s rotation such as Rossby and Kelvin waves. Early evidence of these waves was revealed from shipboard and island observations (e.g., White and Bernstein 1979; Wunsch and Gill 1976). The spatial and temporal structures of the variability were not well resolved by these observations to reveal their basin-scale patterns. Satellite observations of SSH over the past decades have provided sufficient spatial and temporal sampling to allow significant advances in the knowledge of the basin-scale waves.

Patterns of Rossby waves were prominent in the expendable bathythermograph (XBT) observations of the North Pacific (White 1982), but a comprehensive spectral description of these waves was not possible until the satellite altimetry observations from T/P. Shown in Fig. 5-2 (from Fu and Chelton 2001) is the frequency–wavenumber spectrum of SSH at a number of locations
in the global oceans. The variance of SSH variability is concentrated over a band of negative wavenumbers, indicating westward propagation. The band of wavenumbers and frequencies is thus subject to interpretations from the theories of Rossby waves.

The dispersion relation computed from the conventional theory in the absence of a mean current is shown by the white symbols, which do not fit the band well. This discrepancy is also manifested as phase speed that is faster than predicted from the conventional theory. Numerous studies have been inspired to explain the observations. A plausible theory was to include the effects of the vertical shear of a mean current in the calculation of the dispersion relation (shown by black symbols). Comparison of the two dispersion relations provides evidence for the superiority of the revised theory. However, other interpretations of the apparently “nondispersive” linear relationship between dominant frequencies and wavenumbers exist (see Wunsch 2010; Chelton et al. 2011a). See section 2c for more discussion.

In the tropics, the dynamics is complicated by the rapidly diminishing Coriolis effects approaching the equator, leading to the rich structure in the frequency–wavenumber spectrum of SSH shown in Fig. 5-3, computed from altimetry data in a latitude band of 7°S–7°N across the tropical Pacific (Farrar 2011). The region of the dominant frequencies and wavenumbers is underscored by the dispersion relation of Kelvin waves (the black straight line in the side of positive wavenumbers), and those of the mixed Rossby–gravity waves, the first and second mode of the baroclinic Rossby waves (the black curves from the top to the bottom in the side of negative wavenumbers). In addition to the large variance of the Kelvin waves and the baroclinic Rossby waves at large wavelengths of 9°–30°, and low frequencies, a prominent band is enclosed by the white rectangle covering periods of ~33 days and wavelengths of

![Figure 5-1](image-url)
12°–17°, representing the tropical instability waves (TIW). These waves were first detected from satellite observations of sea surface temperature (Legeckis 1977). Satellite altimetry observations have provided an unprecedented database for advancing the understanding of these waves (Lyman et al. 2005). The dispersion relation derived for the TIW by Lyman et al. (2005) from a stability analysis is shown by the gray line in the rectangle, with the white circle denoting the most unstable mode. On the energetics of the TIW, Farrar (2011) demonstrated from SSH observations that the TIW radiates energy to the mid latitudes in the form of barotropic waves, a mechanism of the influence of the tropics on the midlatitude ocean circulation.

The decades-long global observations from space have revealed long-term change of ocean circulation. For example, the basinwide Pacific decadal oscillation (PDO) of the ocean and atmosphere has an imprint on the change of the Kuroshio, a major western boundary currents of the world’s oceans. Qiu and Chen (2011) discovered a relationship between the index of PDO and the characteristics of the Kuroshio Extension jet east of Japan. During the positive phase of the PDO (positive index in the right panel of Fig. 5-4), when the sea surface of the tropical Pacific is warm and that of the northwest Pacific is cool, the jet becomes stronger, as exhibited by positive SSH difference across the jet (the left panel). During the negative phase of the PDO, the jet becomes weaker. Qiu and Chen (2011) also found that the position of the jet moved northward during the positive phase of the PDO and southward during the negative phase. This finding suggests that the Kuroshio might be part of a basinwide ocean–atmosphere coupled system of decadal variability of the Pacific Ocean.

On time scales longer than a decade, satellite observations have shown changes associated with long-term

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**Fig. 5-2.** Frequency–wavenumber spectra computed from the T/P altimeter SSH data along the 24° latitude in both hemispheres. The longitude ranges are labeled on each panel. The dispersion relation computed from the eigenvalue problems for the standard theory with zero mean background flow and the extended theory that includes the baroclinic background mean flow are shown by the open and solid circles, respectively. [From Fu and Chelton (2001); reprinted with permission from Academic Press/Elsevier.]
climate change. Satellite observations of sea surface temperature (e.g., Merchant et al. 2012) have shown a well-sampled trend of the rise of sea surface temperature with the warming climate. An important accomplishment of satellite altimetry is the capability of detecting the long-term change of sea level. The topic of the global mean sea level will be discussed later (section 6). On regional scales, it is a challenge to distinguish between natural variability from a long-term trend. The fact that the linear trend of SSH over a period of nearly 20 years (Fig. 5-5a, from Hamlington et al. 2014) has a high degree of spatial variability is a surprise. Most of the small-scale features are probably caused by the residual variability of the mesoscale eddies. Most of the large-scale patterns are related to those of the PDO (Fig. 5-5b). After the PDO patterns are removed, the residuals can be interpreted as caused by the long-term trend of climate change, which shows a prominent rise of sea level in the region of the western Pacific warm pool (Fig. 5-5c). This is an important finding of the spatial variability of the rise of sea level that stores heat from the warming climate.

Another contributor to sea level change, with more potential of raising the global sea level from the current warming climate is the melting of land ice, from both glacial ice caps and polar ice sheets. The gravity measurement made by the GRACE mission allows the determination of the sea level change caused by the added mass to the ocean from the melting ice from solving the sea level equation driven by the mass change associated with the gravity measurement (Adhikari et al. 2016). Displayed in Fig. 5-6 is such an estimate (Adhikari and Ivins 2016). It shows that the sea level change near the source regions of the Greenland Ice Sheet and the Antarctic Ice Sheet was falling as a result of the decrease in local gravity and the rise of land, both from the loss of mass from ice melt. The mass of meltwater appears in the open ocean remote from its sources. This result illustrates the impact of satellite observations in aiding the prediction of the geographic variability of sea level change from ice melt, a great concern of the coastal population and infrastructure of the world.

c. Mesoscale eddies

An early impact of satellite altimetry was the depiction of the global variability of SSH caused by the mesoscale eddies. Displayed in the upper panel of Fig. 5-7 is the standard deviation of SSH from only one month’s worth of Seasat altimeter data (Cheney et al. 1983). It reflects the spatial pattern of eddy energy over the world’s oceans. Nearly 20 years later, the combination of T/P with ERS-1 altimeter observations produced a more accurate and detailed map of the variability (the lower panel of Fig. 5-7, from Ducet et al. 2000). These are iconic satellite images of the ocean dynamics that provide a test of the performance of numerical ocean models to be discussed later.

Since the launch of T/P, there have always been at least two satellite altimeters orbiting Earth, making mapping the SSH possible at the mesoscale scales larger than about 150 km in wavelength (Ducet et al. 2000). Using a technique of detecting closed-form eddies, Chelton et al. (2011b) estimated the resolution of two satellite altimeters to be about 200-km wavelength at midlatitudes. They tracked individual eddies in a 17-yr record of SSH data from two simultaneous operating satellite altimeters. They have documented a wide range of properties of the global ocean eddy field. Approximately 75% of the eddies propagate westward at the speed of nondispersive long Rossby waves. The remaining 25% are found in the strong eastward currents in the western boundary of the midlatitude gyres and the Antarctic Circumpolar Current (also see Fu 2009). They are drifted eastward with the currents.

A significant fraction (~20%) of the eddies have lifetimes longer than 16 weeks traveling over an average distance ~550 km. Their average amplitude and radius
is 8 cm and 90 km, respectively. Figure 5-8a depicts the trajectories of the long-lived eddies. There is a roughly equal number of cyclonic and anticyclonic eddies. The lack of eddies in the tropical region is primarily owing to their large and complicated size, as well as fast speed, making the tracking technique ineffective. Outside the tropics, the strong eddies with amplitudes larger than 20 cm are concentrated in the region of strong currents as expected (Fig. 5-8b). The size of eddies decreases with latitude as the Rossby radius of deformation (Fig. 5-8c).

The predominantly westward propagation has created a misleading impression of Rossby waves. In many analyses based on data with coarse resolution, the eddies would appear as large-scale, nondispersive waves propagating at speeds somewhat higher than that of long Rossby waves (Chelton and Schlax 1996; Cipollini et al. 1997). As shown in Fig. 5-8d, the degree of nonlinearity, measured by the ratio of the current velocity associated with the eddy to the phase speed of the long Rossby waves, is larger than 2 over most of the oceans. This suggests that the governing dynamics of the eddies is highly nonlinear, a drastic departure from the classic interpretation of the westward propagation noted in both in situ and satellite observations. The dominance of eddies in the observed westward propagation is a factor of the “nondispersive” nature of the spectra shown in Fig. 5-2.

The identification of eddy locations in the altimetry data record allows the collocation of eddy features in other observed variables of the ocean. Gaube et al. (2015) analyzed collocated sea surface wind vectors (from satellite scatterometer) and temperature (from satellite infrared and microwave radiometers). Displayed in Fig. 5-9 are patterns of sea surface temperature and wind speed over identified ocean eddies, demonstrating the corresponding eddy-like patterns of wind and temperature. This illustrates a process of the ocean–atmosphere interaction at the eddy scales. The eddy currents create structures of sea surface temperature, which in turn affects the patterns of wind that drives ocean currents via Ekman pumping. Such self-sustained processes play a significant role in the dynamics of ocean circulation. Moreover, Chelton et al. (2011a) showed

FIG. 5-4. (a) Time series of the SSH height difference across the KE jet averaged from 140° to 165° E (the mean difference is removed). (b) SSH anomalies along the zonal band 32°–34° N from the satellite altimeter data. (c) PDO index from the Joint Institute for the Study of the Atmosphere and Ocean website (http://jisao.washington.edu/pdo/PDO.latest). [From Qiu and Chen (2011).]
that the westward propagation of the chlorophyll anomalies was caused by the stirring of the ambient chlorophyll field by ocean eddies, rather than by trapping and transport of chlorophyll within nonlinear eddies.

d. Ocean dynamics, modeling, and state estimation

Long time series of ocean measurement became available in the late 1960s owing to the development of instrument electronics. The frequency spectra from these measurements revealed the richness of the temporal scales of oceanic variability, with most of the variance taking place at the mesoscale. Arrays of moored instruments were deployed in the 1970s to study the spatial structure of the mesoscale eddies in selected locations. The launch of satellite altimeters enabled the spatial structure of the ocean variability to be sampled on global scales for the study of its wavenumber spectrum, showing the richness of the spatial scales of oceanic variability. The 1-month SSH data from the repeat track of Seasat were analyzed (Fu 1983) to explore the global variability of the SSH wavenumber spectrum. The resulting spectra exhibited drastically different regional characteristics depending on the level of variance. In the high-variance regions near strong currents, the spectra exhibited a power law as \( k^{-3} \), where \( k \) was wavenumber, at wavelengths shorter than the energy-containing wavelength of about 250 km. This was consistent with the notion that the ocean dynamics in this regime of spatial scales was governed by the geostrophic turbulence (Charney 1971), which dictated a \( k^{-3} \) velocity spectrum, or a \( k^{-5} \) SSH spectrum.

In regions of low SSH variance, however, the SSH wavenumber spectra displayed a \( k^{-1} \) power law over wavelengths of 100–1000 km. This was puzzling, because it implied a “blue” spectrum (\( \sim k \)) of the geostrophic velocity, not explainable by ocean dynamics. This puzzle was not resolved until longer and more accurate data had been collected by later altimeter missions. Using seven years of SSH data from Jason-1, which was the follow-on to T/P, Xu and Fu (2012) produced new estimates of the SSH wavenumber spectrum over the global oceans (Fig. 5-10), showing more details than previous studies. Near the ocean surface, theories (e.g., Held et al. 1995) suggest that the spectral power laws for SSH are governed by the surface quasigeostrophic (SQG) dynamics and should follow \( k^{-11/3} \), instead of \( k^{-5} \). Le Traon et al. (2008) suggested that the SSH wavenumber spectrum in the regions of strong currents was consistent
with the SQG theory. Xu and Fu (2012) concluded that only in the core of the major current systems the spectral slopes were steeper than $k^{-11/3}$, approaching $k^{-5}$. The so-called SQG regime occurred at the fringe of the major currents, whereas the vast majority of the ocean interior poleward of the 20° latitudes has SSH spectral slopes between $k^{-11/3}$ and $k^{-2}$. Within a relatively narrow latitude band around the equator, the SSH spectral slopes are flatter than $k^{-2}$. These new findings have posed challenges to theories and models in elucidating the geographic nuances of ocean dynamics.

As predicted by ocean dynamic theories (Rhines 1979), ocean circulation energy tends to cascade from both small and large scales toward the deformation radius of the first baroclinic mode (10–100 km), where the energy is converted into barotropic energy (vertically uniform structure), which then cascades toward larger scales, the so-called inverse cascade. Observations of SSH by satellite altimeters have offered an opportunity to test the theory in the real oceans. Scott and Wang (2005) made the first attempt and showed an inverse cascade at scales larger than the deformation radius. Scott and Arbic (2007) investigated the problem further and demonstrated that the inverse cascade is dominated by baroclinic modes in a quasigeostrophic model simulation. Later it was found that the SQG turbulence theory could explain the inverse cascade of upper-ocean baroclinic energy (Capet et al. 2008). Satellite observations have thus provided credible support for a prevailing theory of ocean dynamics. However, some more recent views of the SQG theory have raised questions on its universal validity (e.g., Lacasce 2012).

Numerical simulations of the ocean circulation began in the 1970s. The first eddy-permitting global ocean general circulation models were developed in the late 1980s and early 1990s, partly inspired by satellite...
3. Air–sea fluxes

Global climate change is caused by a small yet persistent imbalance between the amount of sunlight absorbed by Earth and its thermal radiation emitted back to space. The radiation balance has to respond to air–sea fluxes in an integrated Earth system. The ocean interacts with the atmosphere through the fluxes of momentum, heat, and moisture. The ocean heat exchange with the atmosphere has four components, the sensible heat, the latent heat, the longwave radiation and the shortwave radiation. The exchange of water between the ocean and the atmosphere stems from the difference between evaporation $E$ and precipitation $P$. Momentum is exchanged through surface stress. Latent heat flux is usually estimated from evaporation observations (Semtner and Chervin 1992) and the World Ocean Circulation Experiment. The performance of the models was demonstrated by comparison with satellite altimeter data (McClean et al. 1997, Fu and Smith 1996; Smith et al. 2000). Displayed in Fig. 5-11 is comparison of the standard deviation of SSH from a simulation by an ocean general circulation model with observations from satellite altimetry (from Fu 2009). The model was developed by a group at the Massachusetts Institute of Technology (MIT) led by John Marshall (Marshall et al. 1997). The simulation was performed by a version of the model with horizontal resolution of 18 km and 50 vertical levels (Menemenlis et al. 2005). Continuous development of the model toward even higher resolution is ongoing for fully resolving ocean eddies and even tides and internal waves.

The global data stream from various satellites as well as in situ observations has motivated the development of model-based systems for estimating the state of the global oceans (Stammer et al. 2002; Wunsch and Heimbach 2013). Using a state-of-the-art ocean general circulation model, constrained by various atmospheric and oceanic observations including satellite altimeter observations of SSH, scatterometer observations of vector wind, and hydrography observations from Argo and the World Ocean Circulation Experiment, the project of the Estimation of the Circulation and Climate of the Ocean (ECCO) has produced estimates of the global state of the ocean for the past two decades. Displayed in Fig. 5-12 (from Wunsch and Heimbach 2014) are the heat storage of the global ocean (left) and its change from 1993 to 2011. Although the fact that the Atlantic Ocean is the warmest and the Pacific Ocean is the coolest has been known from in situ observations, satellite observations provide a detailed global view of the heat distribution in the ocean. The change of the heat storage over decadal scales is a powerful indicator of the ocean’s role of absorbing heat from the warming climate, showing where the heat is going. Note that the large increase of heat storage in the tropical western Pacific is consistent with the observational analysis by Hamlington et al. (2014) shown in Fig. 5-5.

FIG. 5-8. Global maps of eddy characteristics that were tracked for 16 weeks or longer in satellite altimeter observations of SSH (October 1992–December 2008): (a) trajectories of cyclonic and anticyclonic eddies (blue and red lines, respectively); (b) mean amplitude for each $1^\circ$ square; (c) mean scale for each $1^\circ$ square, defined to be the effective radius at which the rotational speed averaged around an SSH contour is maximum; and (d) mean nonlinearity parameter $U/c$ (see text). [From Fu et al. (2010) based on Chelton et al. (2011b; reprinted with permission from Elsevier).]
multiplied by the factor of latent heat of vaporization. Surface stress $\tau$, sensible heat $H$, and evaporation $E$ and the related latent heat are transported by atmospheric turbulence. Global synoptic observations of these variables had not been possible until the advent of the space age. There has been effort over decades to estimate $\tau$ and $E$ from space-based measurements, and this is the focus of this section, which is not intended for an exhaustive summary of the past progress, but focused on the evolution of the estimation of these parameters from space.

Atmospheric turbulence at the ocean surface is created by the small-scale random processes associated with wind shear and buoyancy (vertical gradient of wind and density). Turbulence is difficult to measure directly. Historically, it is characterized and inferred from the mean parameters measured on ships, for example, wind $u$ and specific humidity $q$ of air, temperature $T_s$, and current $u_s$ at sea surface, through transfer coefficients in the bulk formula:

$$\tau = \rho C_D (u - u_s)(u - u_s), \quad \text{(5-5)}$$

$$E = \rho C_E (q - q_s)(u - u_s), \quad \text{(5-6)}$$

$$H = \rho c_p C_H (T - T_s)(u - u_s), \quad \text{(5-7)}$$

where $c_p$ is the isobaric specific heat, and $\rho$ is the air density. Bold symbols represent vector variables. Being a function of the vector difference between wind and ocean current, the surface stress has direction and magnitude. The humidity at the interface $q_s$ is usually taken to be the saturation humidity at $T_s$ multiplied by a factor of 0.98 to account for the effect of salt in the water. The coefficients $C_D$, $C_E$, and $C_H$, or the bulk parameters, are empirically determined by measurements at very limited sites for short durations (e.g., Large and Pond 1981; Katsaros et al. 1987; Smith 1989). The measured surface current should include both the geostrophic and ageostrophic components. Liu et al. (1979) determined the fluxes from the bulk parameters by solving the flux–profile relations (or similarity functions), including stability dependence and surface molecular constraints. The method was later improved by Fairall et al. (1996). The parameterization is valid in the

**FIG. 5-9.** Composite averages of anomalies of (a),(b) SST and (c),(d) wind speed in midlatitude eddies. Eddies are segregated according to the meridional direction of the background SST gradient, either southward or northward, and the rotational sense of the eddies, either clockwise or counterclockwise. The composite averages were constructed by rotating the coordinate system for each eddy realization to align the background SST gradient to a polar angle of either ±90°. The magnitude of the asymmetry between the primary and secondary poles of the anomalies is labeled as the value $r$ in each panel. The x’ and y’ coordinates of the composite averages are normalized by the eddy radius scale. The contour intervals of the SST and wind speed composites are 0.05°C and 0.025 m s$^{-1}$, respectively. [From Gaube et al. (2015).]
atmospheric surface (constant flux) layer. Secondary effects, such as the sea state, contribute to the uncertainties of the coefficients. Ocean surface current was often neglected in deriving the transfer coefficients and in the application of the bulk formula, based on the assumption that the speed of current is much smaller than that of wind. Neglecting the surface current altogether may introduce significant errors as discussed in section 3a.

Since the 1970s, the estimation of turbulent fluxes from satellite measurements has largely been made from estimating the bulk parameters from the measurements. There has been skepticism on the notion that the best approach to estimating turbulent fluxes from space-based observations is constrained by the requirement to reproduce ship-based meteorological and oceanic measurements, which usually are not flux measurements. For example, ocean surface stress has to be determined from ocean surface vector wind from scatterometer whose model function was established via comparison with wind measurements by shipboard sensors. Direct retrieval of fluxes has been proposed. For example, direct retrieval of latent heat flux from satellite measurements of radiance, bypassing the use of the bulk parameters, was proposed by Liu (1990). Error analysis on the fluxes derived from bulk parameterization was usually performed by comparing the bulk parameters retrieved from space-based sensors with buoy and ship measurements that are more familiar to the oceanographers. Direct retrieval of fluxes, however, is not easy to validate by in situ observations. There are fewer in situ measurements of fluxes than bulk parameters. The development of the conventional bulk parameterization method and alternative methods, including direct retrieval, for stress and evaporation are discussed in sections 3a and 3b, respectively. The main sensors for estimating stress and evaporation are operational scatterometers and microwave radiometers, respectively. Other types of present and future sensors are discussed in section 3c.

a. Wind and stress

Starting with the launch of Seasat in 1978, microwave radar scatterometer has been the best-established instrument dedicated to the measurement of surface wind vector. Table 5-1 lists the past and current scatterometers whose data are available to the public. A radar scatterometer sends microwave pulses to Earth’s surface and measures the power backscattered from the small centimeter waves (including capillary waves), which are believed to be in equilibrium with the surface stress. The initial geophysical model functions relate measured normalized radar cross section $\sigma_o$ to the frictional velocity $u_o=\left(\frac{\tau}{\rho}\right)^{1/2}$, representing kinematic stress (Jones and Schroeder 1978). The data products of the scatterometer on Seasat were calibrated against measured stress (Liu and Large 1981).
Because the public is more familiar with wind than stress, and there are more wind measurements than stress measurements for calibration and validation, the equivalent neutral wind $u_N$ has been used as the geophysical product. By definition, $u_N$ has an unambiguous relation with surface stress, while the relation between actual wind and surface stress depends on the atmospheric stability. This is a theoretical definition because the atmosphere is rarely neutral (when buoyancy exactly balances wind shear in turbulence generation). A practical way to derive $u_N$ from in situ wind measurements by removing the stability effect was described by Liu et al. (2000), and the result depended to some extent on the formulation of the stability correction. For many decades, scatterometer has been used as a wind sensor and $u_N$ has been assimilated into numerical weather prediction (NWP) models as the 10-m wind. The data have made significant contributions in operational NWP, ocean storm analysis, and understanding air–sea interaction (O'Brien 1999).

Several wind and stress datasets over the global oceans are available (e.g., Risien and Chelton 2008; Desbiolles et al. 2017). Displayed in Fig. 5-13 (upper panels) are maps of the global climatology of stress vorticity (the vertical component of the curl of wind stress) derived from eight years of QuikSCAT data by Risien and Chelton (2008). The wind stress curl field of January and July estimated from scatterometer are compared with those from the NCEP climatology products (lower panels of Fig. 5-13). The comparison has revealed many small-scale features in the scatterometer maps that are missing, not only from the climatology shown, but also from many synoptic wind fields in all the NWP products from various data centers (e.g., Chelton et al. 2004; Chelton and Xie 2010).

Oceanographers may use $u_N$ provided by the scatterometers, neglect the ocean surface current, and simply apply a drag coefficient $C_D$, for neutral stability, in the bulk formula to obtain the stress as the forcing of ocean circulation. There are shortcomings of such bulk parameterization procedures in estimating stress from the scatterometer wind. The shortcomings are related to ocean current, ocean temperature, and flow separation under the strong wind of tropical cyclones (TC), as discussed in sections 3a(1), 3a(2), and 3a(3), respectively. They lead to suggestions of alternative methods, including direct flux retrieval, to bypass bulk parameterization.

1) THE EFFECT OF OCEAN SURFACE CURRENT

There have been studies demonstrating the effects of including ocean current from geostrophic approximation in deriving ocean surface stress from wind, using the bulk formula. Pacanowski (1987) demonstrated the current effect on momentum transport in ocean general circulation model three decades ago. Several numerical experiments showed that the stress computed with the bulk formula with the inclusion of surface current reduces the overall kinetic energy transfer from the wind.
to the ocean (e.g., Dewar and Flierl 1987; Duhaut and Straub 2006; Hughes and Wilson 2008; Hogg et al. 2009; Gaube et al. 2015), Significant errors may be introduced by neglecting the ageostrophic components of surface current. The difficulty in modeling the total surface current was discussed during the establishment of the Ocean Surface Current Analysis–Real Time (OSCAR) project to map surface current (e.g., Lagerloef et al. 1999). By using actual surface current measurements (including both geostrophic and ageostrophic) by Lagrangian drifters, Liu et al. (2007) and Liu and Xie (2008) demonstrated that the scatterometer measurements of wind vector $\vec{u}_N$ had rotation (vorticity) in opposite direction to that of surface current over the meanders of the Agulhas and Kuroshio Extension currents (as exhibited by their mesoscale anomalies). Example of the results at Kuroshio Extension is shown in Fig. 5-14. To focus on the eddy scales, large-scale gradients were removed by a two-dimensional filter. Park et al. (2006) detected “deflections” of scatterometer winds by the Gulf Stream rings and explained the deflection as vector difference between the prevailing wind and the cyclonic and anticyclonic ring currents (geostrophic assumption). Based on these results, Liu et al. (2007) and Liu and Xie (2008) inferred that the scatterometer measures the vector difference between wind and current, which is the stress measurement according to Eq. (5-5). The results also suggest that, unlike the prevailing winds, the ocean surface stress induced by the motion of ocean currents relative to the overlying atmosphere spins down the mesoscale eddies and reduces the kinetic energy into the ocean.

2) THE EFFECT OF SEA SURFACE TEMPERATURE

The spatial coherence between scatterometer measurements and SST [also represented as $T_s$ in Eq. (5-7)] was observed by Xie et al. (1998) in the propagation of the signals of the tropical instability waves. Since then, the coherence has been observed over many locations and under various atmospheric conditions, for example, the western boundary currents, the Antarctic Circumpolar Current, marginal seas during cold air outbreaks, warm and cold ocean eddies, the intertropical convergence zone, and typhoon wakes (see Liu et al. 2000 for a review). Large magnitudes of $\vec{u}_N$ are found over warm water and small magnitudes are found over cool water.
Quantitative relations between wind speed and SST perturbation were derived (e.g., O’Neill et al. 2005, 2012). The positive correlation between SST and wind speed over mesoscale eddies is opposite to the negative correlation found in the seasonal large-scale variations (e.g., Liu et al. 1994). The convergence/divergence centers are in quadrature with SST, located at the steepest SST gradient in downwind direction between the warm and cold centers. Figure 5-15, showing the pattern in the Kuroshio Extension (Liu and Xie 2008) is an example (also see Fig. 5-9).

Explanations of the ubiquitous coherence between wind and SST field through atmospheric boundary layer dynamics have not generally been successful despite many attempts (e.g., Small et al. 2008), because the coherence occurs under very different atmospheric conditions. There have been intense debates between proponents of the vertical momentum mixing mechanism (VMM), attributed to Wallace et al. (1989) and the pressure adjustment mechanism (PAM) attributed to Lindzen and Nigam (1987) to explain the wind and SST relation. VMM is supported by the scatterometer observations of the pattern of divergence anomaly centers in quadrature of the SST centers. The boundary layer wind profiles measured in the tropical Pacific appear to support VMM (Hashizume et al. 2002). PAM is supported by modeling results showing the collocation of the divergence centers with the SST centers. However, bringing momentum from the top of the boundary layer down to the surface, as in VMM, should also be influenced by the pressure gradient force, which would align the convergence centers with the low pressure (warm) centers.

An alternative explanation given by Liu and Xie (2008) on the spatial coherence of scatterometer data with SST is that the small-scale turbulence (stress and latent heat flux) at the surface, unlike the wind aloft, does not depend on the factors (e.g., pressure gradient force, Coriolis force, baroclinicity, clouds) that govern the dynamics of the boundary layer higher up. Buoyancy-induced turbulence has to be coherent with SST, regardless of the atmospheric circulation aloft. Farther up from the surface, pressure gradient force becomes increasingly important, with winds converging to low pressure (warm) center and diverging from high pressure (cold) center.

The lack of wind measurement above the mesoscale eddies made the validation of the alternative hypothesis difficult. Most of our information of wind nowadays comes from the NWP products. Operational NWP centers have been assimilating scatterometer measurements as the 10-m wind, and their surface wind products have almost identical spatial patterns in relation to SST as the scatterometer measurements. Nevertheless, Wang and Liu (2015) were able to detect that the NWP winds above the surface layer (away
from the effect of assimilation of scatterometer winds) have divergence centers more aligned with the centers of SST anomalies.

To characterize the influence of the SST-stress coherence on the forcing of the ocean, we need to examine the stress vorticity with respect to the SST gradient. Liu and Xie (2008) assumed a uniform westerly wind blowing across the observed SST anomalies and computed the $\mathbf{u}_N$ vorticity using similarity functions and demonstrated the crosswind distribution of the vorticity center (Fig. 5-16b). The stress vorticity induced by SST may modify the vorticity induced by ocean surface current of the mesoscale eddies. Such effects (crosswind distribution) are not prominent in the vorticity observed by QuikSCAT (Fig. 5-14b). The SST-induced crosswind vorticity distribution does not appear to be as large as the current-induced vorticity, as detected from the data. The thermodynamics and the oceanic feedback remain to be clarified.

3) TROPICAL CYCLONES

There is a long history of using scatterometers to study TC (e.g., Hawkins and Black 1983; Jones et al. 1999). However, the bulk parameterization does not work under the high wind of TC. The failure to characterize the drag coefficient and the difficulty of retrieving accurate high winds are the problems. A viable solution is the direct retrieval of stress from backscatter measured by the scatterometers.

We do not have sufficient stress measurement to confirm the characteristic of $C_D$ in TC. Under a moderate range of wind speed, $C_D$ is found to increase linearly with wind speed while $C_H$ and $C_E$ are relatively constant (Liu et al. 1979). Emanuel (1995) challenged the application of conventional transfer coefficients to the high wind regime. To attain the wind strength of a TC, the energy dissipated by drag cannot keep increasing while the energy fed by sensible and latent heat does not increase with wind speed. Several studies, using different methods to infer stress, produced a large spread of values and behaviors of $C_D$, as reviewed by Soloviev et al. (2014) and Liu and Tang (2016).

Scatterometers cannot make wind measurements with sufficient accuracy under strong winds, when the operational scatterometer measurements (at Ku- and C-band frequencies) are saturated as illustrated in Fig. 5-17, from Liu and Tang (2016). The data of North Atlantic hurricanes in four seasons (2005–08), excluding those with over 10% chances of rain, were examined. QuikSCAT $\sigma_0$ from its two beams (at different incident angles and polarizations) are plotted against collocated $H^*$wind speed, at bin size of 1 m s$^{-1}$. $H^*$wind is not actual wind measurements but is the product of the Hurricane Research Division real-time hurricane wind...
analysis system (Powell et al. 2003). Figure 5-17 shows that, in moderate winds ($U < 30 \text{ m s}^{-1}$), the logarithm of $\sigma_o$ (in dB) increases almost linearly with the logarithm of wind speed. At strong winds ($U > 30 \text{ m s}^{-1}$), however, $\sigma_o$ increases at a much slower rate with increasing wind speed. When the model function developed over the moderate wind range is applied to the strong winds, an underestimation of wind speed occurs. Intense efforts have been made to adjust the model function (the slope in Fig. 5-17) in strong winds but there are not sufficient in situ measurements available to give credible results. The variations caused by the change of azimuth angle should be a major part of the error bars.

Liu and Tang (2016) bypassed the problems of wind retrieval and $C_D$ by retrieving stress directly from the backscatter in the high winds of TC. The relation between backscatter and surface roughness (waves) was developed from theory and wind tunnel experiments with no consideration of aerodynamics. The relation does not distinguish different weather systems. The same relation should apply in and out of TC. The difficulty of wind retrieval in TC lies in the different characteristics of wind and stress. Liu and Tang (2016) established an algorithm to retrieve stress magnitude over moderate winds, where data are far more abundant and the $C_D$ is better established. Using this algorithm, they retrieved a large ensemble of stress from QuikSCAT backscatter. The large dataset of stress, directly retrieved from QuikSCAT, shows a steep decrease in the $C_D$ with wind speed at winds stronger than 25 m s$^{-1}$. This new $C_D$ implies less stress and less resistance to storm intensification in high winds than estimation using previous $C_D$. Less increase of stress in high winds may mean less cooling of the ocean and more heat into the TC than expected previously.
b. Water and latent heat flux

The equation of water balance in the atmospheric column is

\[
\frac{\partial W}{\partial t} + \nabla \cdot \mathbf{\Theta} = E - P = F, \tag{5-8}
\]

where \( \mathbf{\Theta} = \int_0^L qU dp \) is the moisture transport integrated over the depth of the atmosphere, and \( W = 1/g \int_0^L q dp \) is the precipitable water, or column integrated water vapor. In these equations, \( p \) is the pressure, \( p_s \) is the pressure at the surface, and \( F \) is the freshwater exchange between the ocean and the atmosphere. The first term is the change of storage. Because the residence time of water in the atmosphere is short, for periods longer than a few days, it is negligible. There is a balance between the divergence of the transport \( (\nabla \cdot \mathbf{\Theta}) \) and \( F \). The balance leads to two ways of estimating the freshwater flux. One is to estimate \( E \) independent of \( P \), as summarized in section 3b(1). The other is to estimate \( \mathbf{\Theta} \), with \( E \) being derived as \( (\nabla \cdot \mathbf{\Theta} - P) \). The first method, through the small turbulent-scale processes, has been called the “supply side” estimation; the water is supplied by transport from the ocean. The second method has been called the “demand side” estimation; the large-scale atmospheric circulation demands the water flux from the ocean (Dobson et al. 1982). Most of the past efforts were on the supply side, using the bulk parameterization as summarized in the first part of section 3b(1). Direct retrieval of the flux from the radiances of microwave radiometer is discussed in the latter part of section 3b(1). The efforts on the demand side are discussed in section 3b(2).

1) Evaporation and Latent Heat Flux: The Supply Side

Most productions of space-based evaporation datasets in the past were based on the bulk parameterization [Eq. (5-6)]. Note that the latent heat flux is related to \( E \) by the nearly constant value of the latent heat of vaporization. Discussion in this section on \( E \) applies equally to the latent heat flux. The computation of \( E \) by the bulk parameterization requires SST, \( u \), and \( q \). Over the ocean, microwave radiometers provide measurements of \( u \), \( W \), and SST under both clear and cloudy conditions, but not \( q \). The feasibility of estimating \( E \) using satellite data was demonstrated by Liu (1986), based on an empirical relation between \( W \) and \( q \) on a monthly time scale over the global ocean. The physical rationale is that the vertical distribution of water vapor through the whole depth of the atmosphere is coherent for periods longer than a week (Liu et al. 1991). This relation has been scrutinized in a number of studies and many variations of this method have been proposed to improve the estimation (see Liu and Xie 2014 for a review). In early applications, the latent heat flux data were combined with surface radiative fluxes to examine the ocean’s responses to surface thermal forcing over the global oceans (e.g., Liu et al. 1994).

There has been continuous effort to improve the accuracy and resolution of space-based estimation of the bulk parameters. Liu (1990) suggested the incorporation of the information on the vertical distribution of humidity provided by atmospheric sounders. Jackson et al. (2009) have adopted this suggestion. Various statistical algorithms were developed to retrieve \( q \) directly from the radiance measured by microwave radiometers (e.g., Liu et al. 1991; Schulz et al. 1993; Schlüssel et al. 1995; Jackson et al. 2009; Roberts et al. 2010; Bentamy et al. 2017). The problem was illustrated by Liu and Xie (2014) by comparing \( q \) directly retrieved from the Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E) with 30,000 randomly selected measurements made by buoys. The root-mean-square difference of 1 g kg\(^{-1}\), which is only 5% of the range of 20 g kg\(^{-1}\), appears to be a good result. However, it is a 20% error on \( q - q_s \), which has a smaller range typically of 4 g kg\(^{-1}\). Yu and Weller (2007) combined bulk parameters from space-based observations with model outputs. The errors in \( q \) retrieval have also been attributed to the change of weather regimes. Surface evaporation, as small-scale turbulence, is not affected by the weather dynamics aloft, as the bulk parameters are [see discussion in section 3a(2)].

The measurement of SST from space is critical to estimating air–sea exchanges. Microwave radiometers at certain frequencies have the advantage of measuring SST uninterrupted by cloud and water vapor contamination (Wentz et al. 2000; Chelton and Wentz 2005). The microwave measurements were used to estimate \( E \) through bulk parameterization. Continuous measurements over the tropical oceans were provided by the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) between 1997 and 2015. TMI used only the 11-GHz channel to retrieve SST; this channel is less sensitive to SST in the cold water of the high-latitude oceans [Gentemann et al. 2010]. AMSR-E, on board NASA’s Aqua satellite, provided global coverage between 2002 and 2011, with an additional channel at 7 GHz to optimize SST retrieval in cold water. It was replaced by a second-generation AMSR-2 on the Japanese Global Change Observation Mission–Water (GCOM-W1) in 2012.

All the bulk parameters used in the traditional method could be derived from the radiance (in terms of brightness temperatures) measured by a microwave
radiometer, leading to the estimation of $E$. Liu (1990) demonstrated the direct retrieval of $E$ from radiance. The direct retrieval method may improve accuracy in two ways. The first is the use of a single $C_E$ to derive $E$ (in training the statistical model). The second is to mitigate the magnification of error caused by multiplying inaccurate wind speed with inaccurate humidity ($q$ and $q_e$) in the bulk formula. Based on the comparison with 30,000 randomly selected buoy measurements, Liu and Xie (2014) showed that the direct retrieval reduced the root-mean-square error from 19% to 9.7% of the range. Beside the disadvantage of direct flux retrieval described in the introductory part of section 3, the radiance used to retrieve $E$ has to come from the same sensor, while the bulk parameters from different sensors may be combined to estimate $E$.

2) THE DIVERGENCE OF MOISTURE TRANSPORT: THE DEMAND SIDE

Attempts have been made to estimate $\Theta$ from space-based measurements using advanced statistical techniques (e.g., Xie et al. 2008). The computation of $\Theta$, as defined in Eq. (5-8), requires the vertical profile of $q$ and $u$, which are not measured by space-based sensors with sufficient resolution. $\Theta$ can be viewed as the column of water vapor $W$ advected by an effective velocity $u_e$, so that $u_e = \Theta/W$, where $u_e$ is the depth-averaged wind velocity weighted by humidity and is related to winds at the surface and above the boundary layer. $W$ has been derived from microwave radiometer measurements with good accuracy. Statistical models have been developed and validated to estimate $\Theta$, using cloud drift winds at 850 mb, scatterometer winds and $W$. Xie et al. (2008) showed that $\Theta$ derived from their statistical model agreed with $\Theta$ derived from 90 rawinsonde stations from synoptic to seasonal time scales and from equatorial to polar oceans. Hilburn (2010) found very good agreement between this dataset and the data computed from the Modern-Era Retrospective Analysis for Research and Applications (MERRA) over the global ocean. The $\Theta$ data were further validated through the mass balance of the oceans and continents, using GRACE to monitor the change of water storage and climatological river discharge (Liu and Xie 2014).

Three datasets of $E$, as 3-yr (2003–05) averages, are compared in Fig. 5-18. The first set is retrieved directly from the radiance measured by AMSR-E, as described in section 3b(1). The other two are $E$ produced from the bulk formula: the Hamburg Ocean Atmosphere Parameters and Fluxes from Satellite Data (HOAPS 3; Andersson et al. 2010) and the objectively analyzed air–sea fluxes (OAFlux; Yu and Weller 2007). All the $E$ data products are combined with $P$ from TRMM merged data 3B42 in order to compare with $\nabla \cdot \mathbf{E}$. There are general agreements in the magnitude and geographical distribution.

c. Future perspectives

There has been almost three decades of effort on remote sensing of turbulent fluxes between the ocean and the atmosphere through the retrieval of surface wind vector, SST, and near-surface humidity. Space-based measurements of ocean surface current have not been realized except for its geostrophic component. There are plans for continuous deployments of C-band and Ku-band scatterometers, which are sensitive to the shear between wind and ocean current, allowing direct retrieval of stress. Retrieval of wind vector from scatterometers at L-band on the Aquarius and SMAP satellites has been explored (Fore et al. 2015) and the results do not show high-wind saturation in tropical cyclone as discussed in section 3a(3).

Besides the scatterometers listed in Table 5-1, a polarimetric radiometer, WindSAT, launched in 2003 also demonstrated the sensitivity to wind direction in additional to speed (Gaiser et al. 2004). In spite of early problem and deficiencies, the data have been reprocessed and used as consistent calibration standard (Ricciardulli and Wentz 2015). For continuation, a new polarimetric radiometer, the Compact Ocean Wind Vector Radiometer (COWVR) has been built for deployment in orbit (Brown et al. 2017).

A radiometer to continue and improve AMSR-2 is planned on the GCOM-W1 follow-on mission, which is in formulation phase. Space-based measurements of bulk parameters related to the estimation of evaporation/latent heat flux will be continued. While incremental improvements on $E$ have been achieved, validation studies show that accuracy and sampling problems make it difficult to meet the expected requirements (e.g., Brunke et al. 2011; Prytherch et al. 2015). Direct retrieval from the radiance of microwave radiometers with the necessary frequencies should be pursued. Other methods based on conservation principle, similar to the example described in section 3b(2) are being explored.

4. The large-scale effects of the tropical ocean on the global climate

The circulation of the tropical oceans, transporting heat comparable to or larger than does the tropical atmosphere, has significant influence on Earth’s climate. The tropical oceans also host a suite of large-scale oceanic waves, for example, oceanic Kelvin and Rossby waves. These waves transmit the oceanic response to wind forcing zonally across the tropical oceans, thereby modulating SST and providing
feedback to the atmosphere. Ocean–atmosphere interactions in the tropics result in several modes of interannual climate variability, such as ENSO, which originates in the tropical Pacific sector. The tropics are also intimately linked to decadal climate variability such as the PDO. Such climate variability has significant regional to global influence (e.g., on weather patterns and biogeochemistry) through atmospheric and oceanic teleconnections.

The advance in ocean remote sensing has enabled fundamental progress in monitoring and understanding the tropical ocean circulation and its effect on climate variability. Here we highlight the major contributions of ocean remote sensing in improving the knowledge of the tropical ocean variability and its relationships with the interannual to decadal climate variability.

a. Ocean–atmosphere coupling on interannual time scales

For the interannual variability, the most fundamental knowledge gained from ocean-observing satellites is on the tropical interannual climate modes of the coupled ocean–atmosphere system. These include the measurements associated with ENSO in the tropical Pacific sector, the tropical Atlantic equatorial and meridional modes, and the Indian Ocean zonal/dipole mode (IOZDM). In the following, we highlight some major accomplishments in the understanding of these climate modes.

1) THE TROPICAL PACIFIC OCEAN

ENSO is a dominant mode of the interannual climate variability in the tropical Pacific sector with worldwide influence. It is characterized by large-scale ocean–atmosphere coupling in the tropical Pacific associated with the so-called Bjerknes feedback process (e.g., Bjerknes 1966, McPhaden 1999). This process involves the interannual variation of the strength of the Walker circulation and the associated changes of the tropical Pacific trade wind. The latter causes seesaw of the zonal slope of the thermocline or zonal SSH gradient across...
the basin, which induces variation in the zonal SST gradient that in turn provides positive feedback to the Walker circulation and the trade wind. Oceanic Kelvin and Rossby waves are fundamental to ENSO as they transmit the oceanic response to wind forcing zonally across the equatorial oceans, thereby modulating the zonal SST gradient and providing feedback to the atmosphere.

Satellite observations of ocean surface winds, SSH, and SST have revolutionized the monitoring and understanding of ENSO. The iconic example is the work by Picaut et al. (2002) in depicting and evolution of the 1997/98 El Niño and the underlying oceanic and coupled ocean–atmosphere processes (Fig. 5-19). During February–March 1997, westerly wind bursts in the western equatorial Pacific (Fig. 5-19a) weakened the easterly trade wind, triggering eastward propagation of downwelling equatorial Kelvin waves as reflected by the positive SSH anomalies (Fig. 5-19b) or anomalous depression of thermocline depth. This had the initial effect of warming the SST in the eastern equatorial Pacific because it reduced the influence of colder subsurface waters on the warmer surface layer in the eastern equatorial Pacific. Meanwhile, the edge of the western Pacific warm pool extended eastward (Fig. 5-19c) because of the advection by anomalous westward currents associated with the westerly wind anomalies. The changes in zonal SST structure provided positive feedback to the atmosphere by further weakening the trade wind in the central equatorial Pacific later in 1997 (e.g., June 1997, Fig. 5-19a), which in turn caused larger positive SSH anomalies and warmer SST in the central and
eastern equatorial Pacific. The ocean–atmosphere coupling led to one of the largest El Niño ever observed. During the peak of this El Niño event, the difference of SSH anomalies between the eastern and western equatorial Pacific reached approximately 50 cm, while the SST anomaly in the eastern equatorial Pacific exceeded $3^\circ$C (Picaut et al. 2002).

These satellite observations provided researchers with significant insights about the processes associated with El Niño development and challenged models to reproduce the observed changes. Moreover, they provided the necessary observations to test the ENSO theories, such as the delayed action oscillator theory associated with oceanic equatorial wave dynamics and the subsequent effect on ocean–atmosphere coupling (Suarez and Schopf 1988) and the recharge/discharge oscillator theory that involves the exchanges of heat between the equatorial and off-equatorial upper oceans (Jin 1997). The satellite observations (Fig. 5-19) showed that the delayed action oscillator mechanism was active during the onset of the 1997 El Niño, while both delayed oscillator and the recharge/discharge oscillator mechanisms were at work during the transition to the 1998 La Niña. Detailed discussion of the advance of ENSO theories are available in Battisti et al. (2019, this volume).

Satellite SSH observations have also revealed the importance of the combined effects of wind forced equatorial Kelvin and Rossby waves and their reflections at the eastern and western boundaries on the ENSO cycle (e.g., Delcroix et al. 2000). The reflection of Rossby waves at western boundaries is key to the termination of El Niño in the delayed action oscillator theory. Although such reflection was found to have 80%–90% efficiency based on altimeter SSH, the resultant upwelling Kelvin waves during the peak of the 1997/98 El Niño was insufficient to terminate the warming event without the upwelling Kelvin waves generated by easterly wind anomalies during the peak of that event (Boulanger et al. 2003, 2004).

Ocean surface current estimates derived from satellites, such as OSCAR (Bonjean and Lagerloef 2002; Dohan 2017), have greatly facilitated diagnostic analysis of ENSO. For example, surface current anomalies in the tropical Pacific Ocean are found to lead SST anomalies by approximately 3 months with a magnitude that scales with the SST anomaly magnitude (Fig. 5-20; Lumpkin et al. 2013). These features provide observational evidence for the importance of surface current in regulating ENSO SST across much of the tropical Pacific, and are in support of the important role of the zonal advective feedback mechanism discussed in the context of models and theories (e.g., An et al. 1999).

Sustained satellite observations of SST and SSH have enabled the characterization of the diversity of ENSO events and multidecadal change of ENSO characteristics. Since 2000, there has been more frequent occurrence of the so-called El Niño Modoki (Ashok et al. 2007) or central Pacific El Niño (Kao and Yu 2009).
Different from classical eastern Pacific El Niño where large positive SSH and SST anomalies developed in the eastern equatorial Pacific, El Niño Modoki events are generally associated with positive SSH and SST anomalies and surface current convergence in the central equatorial Pacific because of the anomalous zonal wind convergence associated with El Niño Modoki (e.g., Ashok et al. 2007; Singh et al. 2011). The recharge/discharge of equatorial upper-ocean heat content as inferred from SSH is also different between the two types of El Niño events (Singh and Delcroix 2013). On multidecadal time scales, Lee and McPhaden (2010) found that the amplitude of SST anomalies associated with El Niño in the central equatorial Pacific has doubled from the 1980s to the 2010s. This was due to the more frequent occurrence of central Pacific El Niño as well as the larger amplitudes of central Pacific El Niño since the turn of the century. The increasing amplitude of El Niño in the central equatorial Pacific during the 1980s–2010s has contributed to the observed multidecadal warming trend of SST in the eastern part of the western Pacific warm pool (Lee and McPhaden 2010).

2) THE TROPICAL ATLANTIC OCEAN

For the tropical Atlantic sector, satellite observations have illustrated the two dominant modes of coupled ocean–atmosphere variability: 1) an equatorial (zonal) mode similar to ENSO in the tropical Pacific sector but with a weaker magnitude and 2) a meridional mode with opposite phases (e.g., in SSH and SST) between the northern and southern tropical Atlantic (e.g., Houghton and Tourre 1992; Chiang and Vimont 2004). The equatorial/zonal mode is characterized by changes in the east–west SSH (or thermocline) slope and SST gradient associated with changes of the trade wind (e.g., Arnault and Méléou 2012). The positive feedback between zonal wind, thermocline depth or SSH, and SST (i.e., Bjerknes feedback) help prolong this equatorial mode (Ding et al. 2010; Keenlyside and Latif 2007). The equatorial/zonal mode and the meridional mode are found to be linked through the changes in equatorial wind anomalies (Foltz and McPhaden 2010).

The potential influence of ENSO on the tropical Atlantic interannual climate modes has been investigated. Several studies have shown an influence of ENSO on the northern tropical Atlantic SST with warming in the tropical Atlantic lagging that in the tropical Pacific by a few months (Enfield and Mayer 1997; Huang and Hu 2007; Arnault and Méléou 2012). This ENSO related warming seems to have a signature on SSH in the Gulf of Guinea as well, with a longer lag of the latter (Andrew et al. 2006; Arnault and Méléou 2012). Other climate modes such as the IOZDM and the North Atlantic Oscillation (NAO) do not seem to influence the tropical Atlantic SSH (Andrew et al. 2006).

Because of the weaker variation of the trade wind in the tropical Atlantic than Pacific, the interannual anomalies of SSH in the tropical Atlantic Ocean are much smaller than that in the Pacific associated with ENSO. Given the small-magnitude SSH signals in the tropical Atlantic Ocean associated with natural variability, it might be easier to detect climate change related SSH signal in this region (e.g., Carson et al. 2015; Bilbao et al. 2015).

3) THE TROPICAL INDIAN OCEAN

Unlike the equatorial Pacific and Atlantic, the equatorial Indian Ocean does not have a trade wind and a cold tongue in the eastern part of the basin because of the seasonally varying winds associated with the Indian monsoon. The annual mean wind over the equatorial Indian Ocean is a weak westerly wind. Associated with this is a slightly deeper thermocline and warmer SST in the eastern equatorial Indian Ocean. On interannual time scales, the coupled ocean–atmosphere system in the tropical Indian Ocean sector is characterized by the IOZDM (e.g., Saji et al. 1999).

The IOZDM is associated with anomalous variations of winds in the southeast and equatorial Indian Ocean that induce changes in SSH or thermocline depth and SST in the southeastern and western tropical Indian Ocean, triggering a coupled ocean–atmosphere interaction across the equatorial Indian Ocean that is similar to the Bjerknes feedback process associated with ENSO in the Pacific sector (Webster et al. 1999). The strongest IOZDM event occurred during 1997/98 associated with the strong El Niño. During November 1997–May 1998, the difference of SSH anomalies between the southeastern and northwestern equatorial Indian Ocean reached nearly 50 cm, similar to that in the equatorial Pacific Ocean during the coincident El Niño event (Fig. 5-21; Yu and Rienecker 1999); the SST anomaly in the southeast equatorial Indian Ocean reached approximately −2°C. Negative IOZDM events (e.g., in 1996 and 1998) are associated with opposite changes of winds, SSH, and SST. While some IOZDM events are associated with ENSO events (most notably during 1997/98), studies have suggested that IOZDM can exist as an intrinsic mode of the Indian Ocean without ENSO influence (e.g., Ashok et al. 2003).

SSH observations in the equatorial Indian Ocean have been identified to be a predictor of IOZDM development as it is for ENSO in the Pacific sector. McPhaden and Nagura (2014) used SSH measurements and an analytical linear equatorial wave model to interpret the evolution of the IOZDM in the context of
the recharge–discharge oscillator theory. They found that, as in the Pacific, there were zonally coherent changes in SSH (heat content) along the equator prior to the onset of IOZDM events, an indication of a recharge oscillator being at work. These SSH changes are modulated by wind-forced westward-propagating Rossby waves at 5°–10°S, which at the western boundary reflect into Kelvin waves trapped to the equator. The biennial character of the IOZDM is affected by this cycling of wave energy between 5°–10°S and the equator. However, SSH changes are a weaker leading indicator of IOZDM-related SST anomalies than they are for ENSO. This is because IOZDM is also affected by ENSO through atmospheric teleconnection, in addition to the recharge oscillator process in the Indian Ocean.

The southwest Indian Ocean region (5°–15°S, 50°–70°E), referred to as the Seychelles–Chagos thermocline ridge (Hermes and Reason 2008), in association with a shallow thermocline, is a unique region with open-ocean upwelling as evidenced by enhanced phytoplankton concentration revealed by satellite ocean color data (Murtugudde et al. 1999; Schott and McCreary 2001). The shallowness of the thermocline makes the SST in this region easily affected by the fluctuations of the thermocline depth. This is well characterized by the highly correlated SSH and SST measurements (e.g., Xie et al. 2002; Rao and Behera 2005). Local Ekman pumping associated with the variability of IOZDM- or ENSO-related wind fields causes Rossby wave propagation into this region that influences the thermocline depth (and thus SSH), thereby inducing feedback to the atmosphere (including precipitation) through its effect on SST (Xie et al. 2002). These processes, referred to as coupled ocean–atmosphere Rossby waves, offer potential predictability for SST and tropical cyclones in the western Indian Ocean (Xie et al. 2002).

b. Ocean–atmosphere coupling on decadal and multidecadal scales

On decadal to multidecadal time scales, a dominant mode of coupled ocean–atmosphere climate variability...
in the Pacific sector is the PDO (Mantua et al. 1997; Newman et al. 2016). It has a dominant periodicity of approximately 20–30 years and a spatial structure akin to ENSO. During the positive phase of PDO, SST in the western tropical Pacific Ocean is lower than normal, while that in part of the eastern tropical Pacific Ocean is higher than normal, like a prolonged El Niño. This is reversed during the negative phase of PDO, similar to a prolonged La Niña. Coupled to these decadal SST changes are the variations of sea level pressure and ocean surface wind stress associated with the oscillation of the Walker circulation (http://research.jisao.washington.edu/pdo/). Positive (negative) PDO corresponds to weaker (stronger) trade winds over the tropical Pacific Ocean. While the PDO Index (see Fig. 5-4) is inferred from observations in the North Pacific, observations in the entire Pacific have also been used to characterize the pattern of basinwide decadal variability, referred to as the interdecadal Pacific oscillation (IPO), that is highly correlated to PDO (e.g., Power et al. 1999).

Altimetry data are fundamental to the characterization of SSH patterns associated with PDO and the associated changes in ocean circulation. SSH from altimetry and historical tide gauge data enabled the reconstruction of SSH back to the 1950s (e.g., Hamlington et al. 2011). The leading EOF of global SSH, explaining 41% of the SSH variance (Hamlington et al. 2013), is shown in Fig. 5-22. The spatial structure of SSH pattern of the leading EOF (Fig. 5-22a) is associated with higher SSH in the western tropical Pacific with a horseshoe pattern and lower SSH in the eastern tropical Pacific, generally higher SSH in the subtropics, and generally lower SSH in the subpolar oceans. The temporal variation of the principal component time series for this leading EOF (Fig. 5-22b) is highly correlated with the PDO index, defined as the principal component time series for the leading EOF of North Pacific SST (Mantua et al. 1997). Decadal SSH changes in many regions of the Indo-Pacific domain, including marginal seas, are associated with the PDO (e.g., Lee and McPhaden 2008; Merrifield 2011; Marcos et al. 2012; Zhang and Church 2012; Hamlington et al. 2014) (e.g., Fig. 5-5b). The PDO-related SSH variations dominated the overall basinwide SSH trends during the altimeter period (Hamlington et al. 2014). These regional variations of SSH have strong implications to the associated ocean circulation
based on geostrophic theory, as will be discussed later in this section.

On decadal time scales, the PDO-related SSH variations (Fig. 5-22) are well correlated with traditional ENSO indices (e.g., Zhang and Church 2012; Hamlington et al. 2013). However, decadal changes in ENSO characteristics can modify the regional spatial structure of the PDO-related SSH pattern inferred from SSH reconstruction from the 1950s to the 2000s (i.e., Fig. 5-22).

For the period of 1998–2007, the decadal trend of SSH in the tropical Pacific showed a different spatial pattern (Behera and Yamagata 2010): with SSH in the central tropical Pacific being higher than normal and those in the west and east being lower than normal [see Fig. 2A in Behera and Yamagata (2010)]. This is in contrast to the nodal structure seen in the decadal SSH pattern for the 1950–2010 period seen in Fig. 5-22a. The abnormal SSH behavior in the tropical Pacific during the 1998–2007 period was attributed to the frequent occurrence of central Pacific El Niño or El Niño Modoki (e.g., Ashok et al. 2007; Kao and Yu 2009) events during the early to mid-2000s (Behera and Yamagata 2010). The higher SSH in the central tropical Pacific during 1998–2008 was caused by the convergence of zonal wind anomalies in the central tropical Pacific associated with El Niño Modoki, in stark contrast to the westerly wind anomalies associate with the classical eastern Pacific El Niño.

An accelerated rise of western tropical Pacific SSH was observed for the period of 1993–2009 that is significantly higher than the rate of SSH rise in the preceding decades (e.g., Merrifield 2011). This was associated with a strengthening of the tropical Pacific trade winds that cannot be explained by the changes of the PDO or IPO. Subsequent studies suggested that the strengthening easterly trade winds and the resultant intensification of SSH rise in the western tropical Pacific during this period was caused by teleconnections with the tropical Atlantic and Indian Oceans. The warming in the tropical Indian Ocean strengthened the easterly trade winds in the tropical Indian Ocean sector (Luo et al. 2012; Han et al. 2014; Hamlington et al. 2014). Likewise, the warming in the tropical Atlantic Ocean enhanced the easterly trade winds in the tropical Atlantic sector (McGregor et al. 2014). The resultant atmospheric teleconnections thus strengthened the tropical Pacific trade winds. The interplay of the coupled ocean–atmosphere climate systems in the Pacific sectors with those in the Indian and Atlantic Ocean sectors were found to be instrumental in causing the so-called global warming hiatus since the early 2000s—a period when global mean surface temperature does not increase. During this period, however, SSH data show persistent increase of global mean SSH (Church et al. 2013). Because SSH to a large extent reflects the integral heat content of the entire water column in the ocean, the continuing rise of global mean SSH suggests the excessive heat received by Earth’s surface has gone into the subsurface ocean despite no apparent increase in global mean surface temperature of the ocean was observed. However, a recent study argued that the lack of increase in global mean sea surface temperature during the “hiatus” period may be due to biases in the datasets used in previous analyses (Karl et al. 2015).

SSH show large decadal changes since the early 1990s in the western Pacific and eastern Indian Ocean that surround the Maritime Continent oceanic region (e.g., Lee and McPhaden 2008; Merrifield 2011). Estimation of global ocean heat content based on in situ observations primarily rely on the Argo array, which consists of approximately 4000 autonomous floats profiling the upper 2000 m of the ocean. However, there were very few Argo floats in the Maritime Continent region because of complicated geometry, narrow straits with shallow sills, and strong currents. The impact of not measuring the oceanic heat content in the Maritime Continent region was assessed using SSH as a proxy for oceanic heat content. Excluding SSH data in the Maritime Continent region reduced the estimated global mean SSH trend by 20% (7%) over the 2005–10 (1992–2010) period (von Schuckmann et al. 2014). Therefore, the lack of in situ observations in the Maritime Continent (especially from Argo) raises the question about its potential impact of the estimated decadal changes of global ocean heat content (von Schuckmann et al. 2014). Altimetry data thus provide a mean to fill in this in situ observing system capability gap.

Altimetry-derived SSH and scatterometry-derived ocean surface winds revealed large decadal fluctuations in SSH and winds in the Indo-Pacific sector since the early 1990s (Lee 2004; Lee and McPhaden 2008). Large-scale trends of the SSH and winds during 1993–2000 were generally opposite to those during 2000–06 over much of the Indo-Pacific domain (Lee and McPhaden 2008). The 1993–2000 and 2000–06 SSH trends for the tropical Indo-Pacific domain (Figs. 5-23c,d) were associated with opposite trends of zonal wind stress between these two periods both in the tropical Pacific and Indian Oceans (Figs. 5-23a,b).

The coherent decadal changes of zonal winds in the two tropical basins reflect decadal oscillation of the Walker circulation. These variations in zonal winds forced opposite changes in the meridional transports of warm surface waters in the two basins by Ekman currents (see schematic illustration by the red arrows in Figs. 5-23a,b), which are the upper limbs of the shallow
meridional overturning cells in the two basins. The SSH changes in the northwestern tropical Pacific Ocean are transmitted through the Indonesian Archipelago via coastal Kelvin waves to the southeast tropical Indian Ocean. Consequently, the SSH difference between the eastern and western boundaries are opposite in sign between the two basins.

According to the theory of geostrophy, the east–west SSH differences is proportional to the meridional transports of colder subsurface waters in the pycnocline, which are the lower limbs of the shallow meridional overturning cells in the two basins. Therefore, the Walker circulation and the wave transmission through the Indonesian Archipelago provided an atmospheric bridge and oceanic connection that linked the shallow overturning cells in the two basins, with the overturning cells in the Pacific and Indian Oceans playing opposite roles in regulating tropical Indo-Pacific upper-ocean heat content. Altimetry and scatterometry data are essential for inferring the changes of the Indo-Pacific shallow overturning circulation and their impacts on upper-ocean heat content that play an important role in the tropical coupled ocean–atmosphere system.

Altimetry data have also provided insights into the zonal structure of the pycnocline transports associated with the lower limbs of the shallow overturning circulation in the Pacific Ocean. Such zonal structure has significant implications to the regulation of the upper-ocean heat content and the associated ocean–atmosphere coupling. Altimeter-derived zonal slope of the SSH anomalies across the low-latitude western boundary currents (LLWBCs) and across the interior provide a good proxy for the variations of the geostrophic transports of the LLWBCs and the interior circulation. In contrast to the time-mean flow field where the interior and LLWBC transports reinforce each other (both being equatorward),...
the interannual and decadal anomalies of the interior and LLWBC transports were found to counteract each other (Lee and Fukumori 2003). Therefore, they play opposite roles in regulating the tropical Pacific upper-ocean heat content. However, the variations of the interior transports overcompensate those by the LLWBCs. These observed features were used to benchmark the performance of ocean models and data assimilation products in characterizing the processes controlling upper-ocean heat content (e.g., Hazeleger et al. 2004; Capotondi et al. 2005; Schott et al. 2007). Scatterometer and altimeter data together showed that the anticorrelated variability of LLWBC and interior transports was a result of the oscillation of the tropical gyres in the western Pacific caused by off-equatorial wind stress curl associated with the trade wind varia-
tions (Lee and Fukumori 2003).

The countering LLWBC and interior transports have important implications to the monitoring of the lower branch of the Pacific shallow overturning circulation and the related meridional heat transport. Previous studies of the decadal variation of the lower branch of the Pacific STC were mostly based on interior XBT and CTD observations and the assumption that LLWBC transports do not change significantly (e.g., McPhaden and Zhang 2002). While broadscale in situ observations in the interior ocean provide observations to estimate the interior transport, they generally do not have sufficient coverage near the LLWBCs. Sustained in situ observations of the LLWBCs is a significant technological challenge. Therefore, altimeter-derived SSH provide a means to fill this observational gap, especially with sustained multiple altimeter missions operating simultaneously and with the upcoming Surface Waters Ocean Topography (SWOT) mission (Fu and Ubelmann 2014) that will enhance the temporal and spatial samplings needed to adequately resolve the mesoscale and submesoscale variability associated with the LLWBCs. The contribution of altimetry to fill this observational gap is important to closing the volume, heat, and freshwater budget of the tropical Pacific Ocean.

5. The polar oceans

With the remoteness of the Arctic and Southern Oceans and the logistical difficulties of conducting field programs in these polar environments, observations from various satellites since the mid-1970s have provided the staple of essential basin-scale records that allowed understanding and monitoring of changes in the sea ice cover (extent and thickness), and more recently the dynamic topography of the ice-covered oceans. Here, we highlight some of the significant developments and scientific results from satellite observations of the polar oceans over the past four decades.

a. Sea ice coverage

Sea ice, formed by freezing of seawater, is an important component of the global climate system and a sensitive indicator of climate change. The interest in sea ice coverage pertains to its central role in the powerful ice albedo feedback mechanism that enhances climate response at high latitudes; the presence of sea ice over the polar oceans modifies the exchange of heat, gases and momentum between the atmosphere and polar oceans, and it redistributes freshwater via the transport and subsequent melt of relatively fresh sea ice and hence alters ocean buoyancy forcing. For nearly 40 years, at this writing, retrievals from a series of satellite passive microwave systems have provided near-continuous mapping of the Arctic and Antarctic ice covers (e.g., Markus and Cavalieri 2000; Comiso and Nishio 2008), offering composites that depict the time-varying changes of sea ice extent (defined as areas with at least 15% ice coverage) on a daily basis. Although still relatively short as a climate record, this remarkable dataset has allowed us to document the seasonal and interannual behavior, and multidecadal trends in sea ice coverage at both poles. This record has served as a bellwether of changes in the climate of the Arctic (Vaughan et al. 2013) and to a certain degree that of the global climate system, and also contrasted the recent behavior of sea ice coverage in the Northern and Southern Hemispheres.

For the Northern Hemisphere (NH), the decline in summer sea ice coverage in the warming Arctic has been particularly striking over the past decade (Fig. 5.24, left panels). Record or near record lows in Arctic ice extents have occurred in the years 2005–18. In September 2012, the summer ice extent reached the current record minimum of $3.6 \times 10^6 \text{km}^2$, which was $2.2 \times 10^6 \text{km}^2$ or 30% less than the record set five years earlier in September 2007. For the NH sea ice coverage, which includes ice in the subpolar seas, the trend in average September ice extent has been $-13.0\%$ per decade (1979–2017, relative to the 1981–2010 average; Fetterer et al. 2018); this large decadal trend suggests that ice-free periods during summers (defined as less than a million square kilometer of coverage rather than entirely ice free) may occur around midcentury. This can be compared to a more modest annual decline (all months instead of just September) in the NH sea ice extent of $-4.7\%$ per decade (1979–2017). Spatially, ice cover changes are relatively large in the eastern Arctic basin and most peripheral seas in winter and spring, while changes in the summer
are pronounced almost everywhere in the Arctic basin, except above 82°N.

As less ice survives the summer melt in a warmer climate, the Arctic has lost a significant area of the thicker ice that constituted the perennial ice cover (ice that survives the summer), and seasonal ice now covers more than half of the Arctic Ocean. This represents a marked transition from an ice cover, with a different character, that was once dominated by older multiyear ice (Jeffries et al. 2013). This can be seen in both the passive microwave and scatterometers records of multiyear sea ice coverage of the NH, where the older lower-salinity sea ice can be identified because of their distinct microwave signatures (e.g., Kwok 2004; Comiso 2012). A large volume of literature has been dedicated to the understanding the observed changes, and more recently to the short-term forecast of summer ice behavior (Eicken 2013) and longer-term projection of the fate of the Arctic ice cover (Stroeve et al. 2012).

Between 1979 and 2015, the total Antarctic sea ice extent from satellites has increased at a small but significant rate (Fig. 5-24, right panels) although this masked large regional variations with large opposing trends. A number of mechanisms have been put forward to explain this increase in ice extent (National Academies of Sciences, Engineering and Medicine 2017), but with no consensus on the underlying mechanisms. The increasing trend in the Antarctic sea ice coverage, though modest, is surprising given the warming in the Antarctic region. In fact, climate models generally simulate a decrease in Antarctic ice extent (National Academies of Sciences, Engineering and Medicine 2017). Over the same period, in contrast, global climate model simulations have produced

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**Fig. 5-24.** September trends in sea ice concentration and ice extent anomalies in the (left) Arctic and (right) Southern Ocean. Anomalies (%) are relative to the mean September ice extent between 1980 and 2010. [Data provided by Fetterer et al. (2018).]
negative trends in Arctic sea ice coverage that are consistent with observations. In 2016, satellite observations captured a large and significant decline in Antarctic sea ice extent, which modified the overall trend that was attributed to the anomalous atmosphere circulation that year (Turner et al. 2017). Including the recent declines in 2016 and 2017, the current trend in Antarctic ice extent stands at \(-1.1\) % per decade (Fetterer et al. 2018). For these reasons, the contrasting behavior of the Arctic and Antarctic sea ice has presented a conundrum for global climate change science. This underscores the crucial need to continue monitoring the changing ice cover of both hemispheres.

b. Sea ice drift and circulation

Sea ice moves (drifts) in response to wind and ocean forcing over a broad range of time and length scales. At the large scale, the circulation of sea ice determines the advective part of the ice mass balance while its spatial gradients determine the deformation of the ice cover (Fig. 5-25). Divergent ice drift controls the abundance of thin ice and the many surface processes dependent on thin ice, such as turbulent heat flux to the atmosphere and salt flux into the ocean. Convergent motion changes the roughness (local drag coefficient) and the distribution of the thickness of the ice cover by rafting and ridging.

Sea ice drift is readily observable in time-sequential satellite imagery. The quality of these derived drift estimates depends more on the geometric fidelity (i.e., undistorted representation of distance and angle) and time–space resolutions of the imagery, rather than on a thorough physical understanding of the ice signatures themselves, and hence less sensitive to sensor calibrations. Sea ice drift estimates have been derived from a variety of imaging sensors: lower-resolution passive microwave (PW) brightness temperature and scatterometer backscatter fields (e.g., Agnew et al. 1997; Emery et al. 1997; Girard-Ardhuin and Ezraty 2012), moderate-resolution imagery in the visible/near-IR bands (Ninnis et al. 1986), and high-resolution SAR imagery (Kwok et al. 1990). Despite the fairly coarse resolution of ice drift estimates from PW/scatterometers (~order of kilometers), these fields have been useful for understanding large-scale phenomena like ice advection, ice export, synoptic and longer-term drift patterns, and for assessment and assimilation into ice models rather than for the detailed characteristics (divergence/convergence) of daily drift field (Fig. 5-26). The great strengths of the PW ice drift record are its spatial coverage and the length of the data record, which is close to 40 years for the combination of SMMR, SSM/I, and AMSR-E. Although AVHRR imagery provides wide-swath coverage (~1000 km) at a resolution of ~1 km for motion analysis, gaps in coverage due to clouds make it difficult to maintain routine sampling of the drift fields.

From the record of sparsely sampled buoy drift archived by the International Arctic Buoy Programme, Rampal et al. (2009) found an increase in average drift speed between 1978 and 2007 of 17% ± 4.5% per decade in winter and 8.5% ± 2.0% per decade in summer. A more basin-scale picture of the pattern of the drift trend
can be seen in daily satellite ice drift: Spreen et al. (2011) found that between 1992 and 2008 the area-averaged winter ice drift speed increased by 10.6% ± 0.9% per decade and demonstrated that the observed trend is unlikely due to trends in wind speed. The large increases, during the second half of the period (2001–09), are associated with the years of rapid ice thinning and the strengthening of circulation of the Beaufort Gyre. Increases in drift speed are seen over much of the Arctic except in areas with thicker ice (e.g., north of Greenland and the Canadian Archipelago). In the NH, the circulation patterns (Rigor et al. 2002) and ice export (Kwok and Rothrock 1999) can be linked to the Arctic Oscillation. In the Southern Ocean, the circulation patterns were shown to be linked to the Southern Oscillation and the southern annular mode.

At smaller length scales (meters to kilometers), the response of the ice cover to large-scale atmospheric and oceanic forcing is concentrated along fractures up to kilometers wide and lengths that can span thousands of kilometers (Kwok 2001). As sea ice is a brittle solid, it does not deform continuously throughout the ice cover (like that of a fluid); instead, it moves and deforms when cracks and fractures are introduced by material failure. The geophysical interest in detailed ice drift is the need to develop improved representations of sea ice behavior in models (currently modeled with viscous-plastic...
rheology) with resolutions that approaches these length scales of kilometers. This nature and details of ice drift at this length scale places significant requirements on the spatial resolution and coverage of the imaging sensors for ice motion analysis.

Of particular interest, therefore, has been the small-scale ice drift from high-resolution SAR acquisitions (Fig. 5-26). Spaceborne SAR imaging of the ice cover has the advantage of all-weather, day and night operational capability, fine ground resolution (~10–100 m), and good geometric accuracy. Ice motion retrievals from SAR data were first demonstrated with Seasat imagery, followed by ERS-1, and then imagery acquired by subsequent SAR missions [e.g., RadarSat, Advanced Land Observing Satellite (ALOS), Envisat, Sentinel-1, etc.]. The resolution of these drift estimates approaches the length scales and time scales needed for observing the expressions of the small-scale sea ice processes. The need has been to understand the behavior of small-scale ice drift, for improving ice dynamics in sea ice models (e.g., Coon et al. 2007; Sulsky and Peterson 2011; Bouillon and Rampal 2015), for documenting changes in sea ice deformation (Spreen et al. 2017), and for assimilation into higher-resolution ice–ocean models (e.g., Nguyen et al. 2011).

These finescale drift fields from SAR imagery have been used to quantify the various measures of the opening, closing and shear of the ice field, estimate ice production and thickness, and for assessment of ice models. At even shorter time scales, ice deformation at subdaily ice drift associated with tidal forcing or inertial effects are becoming more pertinent as the ice cover thins (Kwok et al. 2003). Subdaily sampling, however, remains a challenge because of the lack of sufficiently frequent repeat coverage from orbiting satellites. New ice production due to the recurrent openings and closings at subdaily time scales, has been shown to be significant within the winter pack in both the Arctic and Antarctic. This is a challenge to satellite remote sensing of the sea ice cover.

c. Sea ice freeboard and thickness

Observations of changes in both the ice coverage and thickness are needed to understand the behavior of sea ice cover in a warming climate. The importance of ice thickness in sea ice mass balance and in the heat and energy budget at the surface (air–sea–ice interface) has long been recognized; hence remote determination of ice thickness at almost any spatial scale is desired. Satellite microwave radiometers or SAR, however, can see only radiation emitted or scattered from the surface or the volume within the top few centimeters of the ice and do not see the ice-water interface, making it difficult to obtain direct observations of ice thickness. The current approach has been to use altimeter-derived freeboard (i.e., the height of the ice surface above the local sea surface) together with the assumption of hydrostatic equilibrium to determine ice thickness. Owing to the relatively small differences between ice and water densities (ice density is about one-tenth that of water density), only a small fraction of the ice thickness is visible as freeboard above the sea surface: the nearly tenfold multiplication of freeboard uncertainties in the estimation of ice thickness is daunting and places stringent demands on the accuracy of freeboard retrievals.

The first geophysical results of Arctic ice freeboard/thickness estimates from satellite radar altimetry (ERS-1) were demonstrated by Laxon et al. (2003). However, the orbit inclination of the ERS-1 platform limited the coverage of the Arctic Ocean to 82°N, and hence was not able to monitor basin-scale thickness changes. Although the acquisitions from the ICESat lidar mission (2003–09) (Zwally et al. 2002) provided coverage to 86°N, restricted lidar operations allowed only surveys of the fall and winter sea ice cover of the Arctic and Southern Oceans. Even so, the combined ice thickness records from U.S. Navy submarine and ICESat showed a dramatic thinning of the Arctic ice cover of almost 50%, from 3.64 m in 1980 to 1.75 m in the short span of ~30 years (Kwok and Rothrock 2009) (Fig. 5-27).

ESA’s CryoSat-2 (launched in 2010) operations at an orbit inclination that allowed near-complete coverage of the Arctic Ocean—to a latitude of 88° (Wingham et al. 2006). While ICESat lidar operations were restricted to only two to three 34-day mapping campaigns annually, the monthly CryoSat-2 thickness maps of the Arctic ice cover provided more complete and frequent temporal surveys for resolving seasonal and basin-scale processes, as well as responses of the ice cover to thermodynamic and especially shorter-time-scale dynamic forcing. A first assessment of Arctic ice volume based on two years (2011–12) of CryoSat-2 acquisitions (Laxon et al. 2013) showed further declines in Arctic sea ice volume following the ICESat mission. At this writing, more than 9 years of CryoSat-2 data are available to the science community. The combined records of ice extent, thickness and volume (Fig. 5-28) from ICESat and CryoSat-2, have provided a more comprehensive picture of the behavior of the Arctic ice cover in a changing climate (summarized in Kwok 2018). The instrument on the recently launched ICESat-2 mission (September 2018) (Markus et al. 2017) is a photon-counting lidar that provides even higher spatial resolution and denser sampling of surface height.

In the retrieval of sea ice thickness from freeboard, estimates of snow loading (snow depth and density)—a
source of uncertainty—are required (Kwok 2011). Currently, routine observations of snow depth and density over the Arctic Ocean are not available. Retrievals from satellite altimetry provide only sea ice freeboard (e.g., radar altimeters on CryoSat-2, AltiKa, Sentinel-3) or the combined snow and ice freeboard (lidars on ICESat, ICESat-2), and snow loading—in the calculation of thickness—is either modeled or empirically derived. Hence, there is extensive interest in the climatology, seasonal and interannual variability, and spatial distribution of snow depth for ice thickness estimation, climate analyses and modeling, and forecast of sea ice behavior. Presently, uncertainties in Arctic ice thickness estimates vary regionally and seasonally, and assessments show that they are typically ±0.5 m (e.g., Laxon et al. 2013; Kwok and Cunningham 2015).

Retrieval of Antarctic sea ice thickness from satellite altimetry remains a challenge (Kwok and Maksym 2014) largely because of uncertainties in snow depth and potential penetration issues at radar wavelengths (Giles et al. 2008). Discrepancies between ice thickness estimates stem from different approaches used to determine snow depth (Zwally et al. 2008; Kurtz and Markus 2012; Xie et al. 2013). These results point to the need for better sampling of the Antarctic sea ice cover, so as to understand the sampling biases in terms of the field observations, which are critically needed for the assessment of the satellite retrievals and for informing methodologies for conversion of freeboard to thickness. However, the ice areas within the interior Antarctic sea ice cover are largely impenetrable to most icebreakers, and the accessibility issues remain a challenge to provide adequate field-based sampling of the ice cover.

d. Dynamic topography of the ice-covered oceans

Interest in the dynamic topography of the ice-covered oceans pertains to its importance in controlling the circulation of sea ice, mixing and exporting freshwater, and providing heat to melt ice. For several decades, the TOPEX/Poseidon and follow-on missions have provided synoptic-scale measurements of sea surface topography of lower-latitude oceans (discussed in earlier section). Sampling of sea surface of the ice-covered oceans of the Northern and Southern Hemispheres, however, has been restricted to only a few percent of openings (exposed ocean surface) within bounds of a satellite’s inclination. Furthermore, specially designed approaches are needed to separate the returns from the ice and sea surface.

Though not optimized for remote sensing of the sea ice covers, radar altimeters on early ESA missions [ERS (1990–2011) and Envisat (2002–12)] have provided valuable observations of the polar ice covers Peacock.
and Laxon (2004) using ERS-2 radar echoes from open leads, offered a first glimpse of SSH of the Arctic Ocean. With a 15-yr record from these altimeters, Giles et al. (2012) noted that increases in height of the dome of the Beaufort Gyre, reported by Proshutinsky et al. (2009), can be observed in the sea surface height anomalies. Dedicated ice missions (e.g., ICESat, CryoSat-2), since 2003, with higher orbit inclinations have provided more routine sampling and broader coverage of the sea surface of the polar oceans. The ICESat mission (2003–09) was a higher-resolution (footprint: 50–70 m) and high-precision lidar instrument (shot-to-shot repeatability of ~2–3 cm) that, allowed unambiguous identification of water or thin ice in narrow openings. Though coverage was limited by cloud cover, a number of investigators have produced and assessed fields of dynamic ocean topography (DOT) of the Arctic Ocean derived from ICESat (e.g., Forsberg and Skourup 2005; Kwok and Morison 2011; Farrell et al. 2012). The significant correlations (0.92) between ICESat-derived ocean topography and dynamic heights from in situ hydrography, noted by Kwok and Morison (2011), highlight the large baroclinic signals in the Arctic Ocean. In a basin-scale study, Morison et al. (2012) using spatial changes in ICESat DOT, GRACE ocean bottom pressure, and hydrography showed that recent increases in Arctic freshwater content were associated with a cyclonic shift in the ocean pathway of Eurasian runoff. These changes were forced by the strengthening of the west-to-east Northern Hemisphere atmospheric circulation associated with the Arctic Oscillation. As in the low-latitude oceans, these results highlight the geophysical utility of these time-varying sea surface heights from altimetry.

Compared to the campaign mode operation (2–3 times per year) of the ICESat lidar, the mapping of the polar oceans with the CryoSat-2 radar altimeter (launched in 2010) has become more routine. The instrument includes a synthetic aperture altimeter, a radar design that provides higher-resolution profiling of the surface compared to traditional altimeters (Wingham et al. 2006). This continuous data acquisition of CryoSat-2 allows for a closer look at the time-varying component of the ocean’s topography. For example, Kwok and Morison (2015) produced and assessed SSH fields of the Arctic and Southern Oceans, and the work by Armitage et al. (2017) summarized the geostrophic circulation of the Arctic Ocean since 2003 (Fig. 5-29). By making use of recently available DOT and ice drift, a recent study by Dewey et al. (2018) utilized observations of geostrophic ocean currents driven by ice–ocean stress to examine the relative role of ocean current and wind speeds in the stabilization of the Beaufort Gyre. They showed the shift of the Beaufort Gyre from a system in which the wind drives the ice and the ice then drives a passive ocean, to one in which the ocean often, in the absence of high winds, drives the ice. The stress exerted on the ocean by the ice and the resultant Ekman pumping is reversed (because of the relative drift speed of the ice

![Figure 5-29. Dynamic ocean topography from CryoSat-2. (a) Arctic Ocean [after Kwok and Morison (2015)]; (b) Southern Ocean [after Armitage et al. (2018)] (contour interval: 40 cm).](image-url)
and ocean), without any change in average wind stress curl. Through these curl reversals, the ice–ocean stress provides a key feedback in Beaufort Gyre stabilization.

In the first large-scale mapping of the sea surface of the ice-covered Southern Ocean, Armitage et al. (2018) combined altimetric SSH estimates of the ice-covered Southern Ocean (Kwok and Morison 2015) with conventional open-ocean SSH estimates from CryoSat-2 to produce monthly composites of dynamic ocean topography and sea level anomaly spanning 2011–16. These composites allowed examination of the seasonal cycle of the Southern Ocean SSH and revealed an antiphase relationship between sea level on the Antarctic continental shelf and the deeper basins, with coastal SSH highest in autumn and lowest in spring. The pattern of the seasonal SSH variability showed the barotropic component of the Antarctic Slope Current (ASC) has speeds that are regionally up to twice as fast in the fall. The variability of SSH was also shown to be linked to two dominant climate modes: the Southern Oscillation and the southern annular mode. During the strong 2015–16 El Niño, a sustained negative coastal SLA of up to 26 cm, implying a weakening of the ASC, was observed in the Pacific sector of the Southern Ocean. The ability to examine sea level variability in the seasonally ice-covered regions of the Southern Ocean—climatically important regions with an acute sparseness of data—makes this new merged sea level record of particular interest to the oceanography and glaciology communities.

6. Global mean sea level change

The rise of the global mean sea level has been documented by the observations from tide gauges over the past 100 years (Douglas et al. 2000). The sparse and uneven distribution of the tide gauges has created poorly determined sampling errors in achieving the estimate of a global mean. The advent of precision altimetry missions beginning with T/P has led to the first directly observed global mean sea level, well sampled from 66°S to 66°N. However, the small change of the global mean sea level had not been considered an objective of satellite altimetry until considerable progress was made in improving the estimate of a global mean. The advent of precision altimetry missions beginning with T/P has led to the first directly observed global mean sea level, well sampled from 66°S to 66°N. However, the small change of the global mean sea level had not been considered an objective of satellite altimetry until considerable progress was made in improving the measurement performance of T/P and other altimetry missions.

Over the two decades since the launch of T/P in 1992, the global mean sea level observed from altimetry shows a trend of $3.4 \pm 0.4 \text{ mm yr}^{-1}$ (Nerem et al. 2010). A recent study by Nerem et al. (2018) demonstrated that the global mean sea level has been accelerating at a rate of $0.084 \pm 0.025 \text{ mm yr}^{-2}$. The exclusion of the high-latitude oceans from the series of precision altimetry missions of T/P and the Jason series has caused concerns for the representation of their results for the global mean. More recent analysis (Ablain et al. 2017) compared the rate of sea level rise from these missions with that from high-inclination missions, arriving at 2.98 mm yr$^{-1}$ (the former) versus 2.36 mm yr$^{-1}$ (the latter) for the period of 1993–2011. Nevertheless, the observations from altimetry have been used for improving the estimates from the past tide gauges by fitting the contemporary global spatial patterns of the sea level trends estimated from altimetry to the data from the past tide gauges. Displayed in Fig. 5-30 are two versions of the “reconstructed” global mean sea level superimposed on the altimetry record (Church and White 2006, 2011). The evolution of the estimated uncertainty reflects the impact of altimetry and the increasing coverage of tide gauges with time.

As noted earlier, the desire of measuring Earth’s gravity field from space in parallel to satellite altimetry had been conceived at the beginning of the space age. Although the original rationale was to provide the geoid for estimating ocean surface circulation from sea surface height, it was realized later that the temporal variability of the gravity field might have a wide range of important applications to Earth system science (National Research Council 1997). One of them is the attribution of sea level change to the effects of melting ice as shown in Fig. 5-31 (from Ablain et al. 2017). When integrated globally, the

![Fig. 5-30. Global average sea level from 1860 to 2009 as estimated from the coastal and island sea level data (blue). The one standard deviation uncertainty estimates plotted about the low passed sea level are indicated by the shading. The Church and White (2006) estimates for 1870–2001 are shown by the red solid line and dashed magenta lines for the one standard deviation errors. The series are set to have the same average value over 1960–90 and the new reconstruction is set to zero in 1990. The satellite altimeter data since 1993 is also shown in black. [From Church and White (2011); reprinted by permission from Springer Nature.]](image-url)
results of GRACE, as previously shown in Fig. 5-6, are able to account for the contributions of ice melt to the change of the global mean sea level. The combination of the steric sea level change (caused by the change of ocean density), estimated from the Argo data, with the barystatic sea level change (caused by the change of ocean mass) matches the altimetry measurement of the total sea level change fairly well (see Fig 5-31). The synergy of the three measurement systems creates an opportunity of cross validation of this difficult and important global measurement, which is an indicator of the extent of climate change as well as its impact on the world’s coastal zones.

7. Tides

To study ocean circulation from SSH, one must remove the signals of ocean tides that dominate the altimetry observations of SSH. The orbit of T/P was specifically selected to allow the separation of the signals of ocean tides from those of ocean circulation. This aspect of mission design and the ensuing efforts of research have led to an unexpected revival of the age-old tidal science. Shown in Fig. 5-32 is the amplitude of the principal semidiurnal M2 tides derived from an inverse model of tides fitting to the T/P observations (Ray and Egbert 2018). Independent bottom-pressure measurements in the deep ocean agree with this chart with a root-mean-square difference of 0.5 cm. The state-of-the-art tidal models are now able to predict the tides in the open ocean with accuracy better than 2 cm, often approaching 1 cm.

New estimates of tidal dissipation estimated from the tide models derived from altimetry have indicated significant dissipation of tidal energy in the open ocean, in sharp contrast to the well-established notion of dominant dissipation over the sallow seas. Displayed in Fig. 5-33 (from Egbert and Ray 2000) is the geographic distribution of the tidal dissipation. It shows that approximately 10^{12} watts, or 1 terawatt (TW), accounting for 25%–30% of the total tidal dissipation, takes place in the deep ocean, mostly over areas of rough topography. This result suggests that roughly half of the total 2 TW energy required for ocean mixing to maintain the large-scale thermohaline circulation of the ocean (Munk and Wunsch 1998) is provided by the tides, with the other half provided by atmospheric forcing near the surface of the ocean.

The primary mechanism of the tidal dissipation in the open ocean is via the generation of internal tides. Given their small signals of less than 5 cm in SSH, the finding of internal tides in the T/P altimetry data was among the most surprising results of the mission (Ray and Mitchum 1996). Shown in Fig. 5-34 is an example of the SSH wave fronts associated with the M2 internal tides in the tropic Atlantic, showing distinct patterns of beam of waves from the source regions (from Ray and Egbert 2018). In sharp contrast to the picture conveyed in Wunsch (1975) that internal tides were incoherent in space and time, the findings from altimetry showed that internal tides were primarily phase locked to the astronomical potential and were spatially coherent over distances more than 1000 km.

The global observations of internal tides from altimetry have literally revived the field of internal tides, leading to dedicated field experiments like the Hawaii Ocean Mixing Experiment (HOME; Pinkel and Rudnick 2006), as well as modeling of internal tides with ocean general circulation models (Arbic et al. 2012). These new findings have profound impact on our understanding of tides and their effects on ocean mixing, an important element of the ocean circulation energy budget.

8. Concluding remarks

Over the past 50 years we have witnessed a revolution in oceanography with the advent of the capability of observing the global oceans from space. The notion of a sluggish laminar flow of the global ocean circulation has been transformed to one of an energetic turbulent fluid system with variability on all scales. This new finding has modernized the practice of oceanography, changing it from a curiosity-driven discipline to one that is essential to the well-being of society, particularly in dealing with
the ongoing change of climate. The large international investment in satellite oceanography as well as the global ocean in situ observing systems has produced well-earned return, having equipped society with an observing system that is sufficient for meeting certain important objectives in detecting climate change and consequences.

An example is the system of satellite altimetry, gravimetry, and in situ drifting floats in addressing the problem of global sea level change and oceanic heat uptake, which is a grand challenge facing society. The current uncertainty (1 standard deviation) of detection of the global mean sea level change from the system is 0.5 mm yr\(^{-1}\) estimated over the course of a decade. Given the estimated acceleration of the global mean sea level rise at a rate \(\sim 1\) mm yr\(^{-1}\) decade\(^{-1}\) (Nerem et al. 2018), the present capability is able to detect the acceleration at a confidence level of 95%.

A challenge facing society in the future is the decision in appropriating available funds in making global observations of the ocean. How to balance the sustenance of the existing system vis-à-vis the investment in
developing new technologies? How to develop the methodology that can provide quantified analysis of the trade-off of various options leading to a strategy for sound decision-making (National Research Council 2015)? We believe this is a priority in the future development of satellite remote sensing of the ocean.

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