Ocean circulation depends on momentum, heat and freshwater exchanges with the atmosphere, and on the temperature and salinity of the ocean, as these determine its density structure. From the density structure it is possible to calculate the ocean flows using geostrophy (in much the same way as for the atmosphere). The concept of geostrophy is familiar to anyone who has watched a weather forecast on television. It essentially represents the balance between pressure gradients and the Earth’s rotation (Coriolis effect) that leads to flows at right angles to the pressure difference. Knowing the density of the ocean gives information on the pressure differences from which the flow (circulation) can be calculated.\(^1\) In the atmosphere this means that the wind on synoptic scales does not blow from high pressure to low pressure but perpendicular to the pressure difference along the isobars (ignoring the added complication of friction). The greater the pressure difference, over a fixed distance, the faster the flow.

Salinity represents the amount of salt dissolved in sea water and is expressed traditionally as a ratio ‰ (parts per thousands, a dimensionless number). More technically, in terms of how salinity is actually measured, practical salinity is defined on the Practical Salinity Scale of 1978 in terms of a conductivity ratio, which is the electrical conductivity of a seawater sample at a temperature of 15°C and a pressure equal to one standard atmosphere, divided by the conductivity of a standard potassium chloride solution at the same temperature and pressure (see IOC, SCOR and IAPSO, 2010; Talley et al., 2011). Although it is dimensionless, salinity is sometimes referred to using the terms pss (practical salinity scale) or psu (practical salinity units). The open ocean surface salinity varies globally between 33 and 37, roughly equivalent to 33–37g of salt dissolved in a litre (kg) of water. With a salinity of around 40, the Red Sea holds some of the saltiest seawater on Earth, while the English Channel, with a salinity of around 35, is closer to the global ocean average of 34.6 (75% of the total volume of the ocean has a salinity between 34 and 35; Talley et al., 2011). At first glance, these variations appear small, but they have a significant impact on the density of sea water, especially in cold water regions.

Traditionally oceanographers have made measurements of temperature and salinity from ships and buoys, but to get a global picture requires a different approach. Two approaches have been developed, one being the global array of Argo floats (Riser et al., 2016), the other making measurements from space. Routine measurements of sea surface temperature (SST) from space, using infrared and passive microwave radiometers, have been available since the 1970s. In contrast, corresponding sea surface salinity (SSS) measurements have only been available since the launch of the European Space Agency’s (ESA) Soil Moisture and Ocean Salinity (SMOS) mission in 2009 (Kerr et al., 2010; see Figure 1). This article discusses why measurements of salinity are needed, why from space, how such measurements are made, and what these measurements can tell us about the ocean.

**Why measure ocean salinity and why from space?**

On global and ocean basin scales atmosphere–ocean heat and freshwater fluxes drive the thermohaline circulation, also known as the ‘great ocean conveyor’, which is important for the ocean’s role in climate change. One part of the ‘conveyor’ – known as the Atlantic Meridional Overturning Circulation (AMOC) – involves the formation of deep water at high latitudes in the North Atlantic, which depends on the salinity of the surface waters there. As the warm waters flow north in the Atlantic, they give up heat to the weather systems crossing the ocean towards northwest Europe, which leads to a more temperate winter climate than might otherwise be expected. As the northward-

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\(^1\)Unlike for the atmosphere, pressure is not used directly, as obtaining sufficiently accurate measurements of pressure in the ocean is problematic (see Talley et al., 2011).
flowing waters cool, they become denser and eventually sink and return as a deep flow. Changes in the surface salinity due to climate change, such as increased rainfall or enhanced melting of the Greenland ice sheet, could disrupt the AMOC by adding fresh water, making the sea water less dense and unable to sink. Such changes would affect both the climate and the sea levels of the regions surrounding the North Atlantic (for more details of the AMOC and its impacts see Srokosz and Bryden, 2015).

On interannual timescales the tropical Pacific Ocean plays an important role in the El Niño-Southern Oscillation phenomenon. Naively it might appear that exchange of heat between the atmosphere and ocean are responsible for tropical Pacific atmosphere–ocean interactions, but many studies have emphasised the part played by salinity and its link to tropical precipitation (particularly in the Inter Tropical Convergence Zone – ITZC; see, for example, Yu, 2014). More recently a link has also been found between Atlantic salinity and Sahel rainfall on intra-annual timescales (Li et al., 2016), which may allow improved prediction of precipitation over the Sahel.

The mesoscale, the scale of fronts and eddies in the ocean (typically about 100km), is the scale at which the most energetic motions in the ocean occur. Eddies have different water properties (either temperature or salinity, or both) from the surrounding waters (Melinchenko et al., 2017). Fronts are characterised by ‘abrupt’ changes in water properties, usually temperature and/or salinity (Spall, 1995). For both eddies and fronts, salinity can be important in determining the density structure, which, in turn, affects the dynamical processes. Additionally, eddies are responsible in part for transferring energy and water mass properties and for ocean mixing, all of which impact the ocean circulation.

The processes through which salinity affects the ocean circulation are linked to the global hydrological cycle. The ocean surface salinity depends on the atmosphere–ocean exchange of fresh water through precipitation (P) and evaporation (E) (Yu, 2011; Reul et al., 2014). Early studies found a simple linear relationship between salinity at the ocean surface and the balance of E minus P (E–P). More generally, the sea surface salinity is determined by E–P, advection, river run off, glacial melt and by turbulent mixing of the surface with underlying waters. As E–P is difficult to estimate for the oceans – in particular, measuring precipitation is problematic, whereas evaporation can at least be estimated (Yu, 2007) – using ocean surface salinity measurements it might be possible, by combining with surface current information and mixed layer depth data, to estimate E–P (Vinogradova and Ponte, 2013). By comparing the annual mean global distributions of E–P and sea surface salinity given in Figure 2, the link between the two can be seen clearly.

As noted above, routine SST measurements from space have been available since the 1970s and have revealed much about the complex structure of the ocean circulation in terms of currents such as the Gulf Stream and the eddies that are ubiquitous in the ocean. Similarly, satellite RADAR altimetry has contributed hugely to the areas of sea level rise (Cazenave and Llovel, 2010), the study of Rossby waves and eddies (see, for example, Chelton and Schlax, 1996; Chelton et al., 2007) and many other aspects of ocean circulation. More recently, particularly since the mid-2000s, the array of Argo profiling floats that measure temperature and salinity over the top 2000m of the ocean every 10 days, and report the data back via satellite links, has greatly increased the number of observations of salinity available to oceanographers (Riser et al., 2016). However, even with the 3000 or so Argo floats currently active, the data are available at a spacing of approximately 300km, so their coverage of the global ocean is limited to larger scales. In addition, a faster sampling than every 10 days might be advantageous for the study of some ocean phenomena, such as fronts and eddies. For a long time global observations of ocean salinity to complement those of SST from space have been seen as desirable, but why did it take from the 1970s SST era to the end of the 2000s before such measurements were achieved?

How is salinity measured from space?

The ocean emits radiation across the electromagnetic spectrum. For example, its infrared emissions have long been used to measure SST from space. In the microwave part of the spectrum, the ocean’s emissions (as measured by the brightness temperature) depend on the dielectric properties of sea water (essentially a function of conductivity), which, in turn, are known to depend on temperature and salinity. Therefore, the measurement of sea surface salinity from space relies on changes in the sea water’s dielectric constant (which affects the emissivity and brightness temperature), at microwave frequencies, due to variations in salinity. It has long been known that such changes can be measured using a passive microwave radiometer to determine the brightness temperature of the sea surface, in principle. However, such measurements need to be made at a relatively low microwave frequency (L-band), typically 1.4GHz. This is because, at lower microwave frequencies, the sensitivity to sea surface salinity variations is greatest and the effects of other factors are smaller, with the exception of temperature. Therefore, to determine sea surface salinity requires knowledge of the sea surface temperature (as this too impacts the brightness temperature). Note that L-band measurements are largely unaffected by atmospheric phenomena, with the exception of very heavy rain, in contrast to the longstanding infrared measurements of SST, which cannot ‘see’ through clouds; passive microwave SST measurements, however, are unaffected by clouds but can be affected by rain.

The major constraining factor on making such a measurement from space, and the reason for the several decades of delay in measuring SSS as compared with SST, is the size of the antenna required to obtain a given spatial resolution on the ocean surface. For a given orbital height h for a spaceborne sensor, the spatial resolution d depends on the ratio of the wavelength l of the microwave radiation to the antenna size D; in fact, \( d \approx \frac{l}{h} \). At 1.4GHz, l ≈ 21cm, and for a typical satellite polar orbit height

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2The depth of the ocean’s homogeneous surface layer, which is well mixed by turbulent processes.

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Figure 2. (a) Annual average SSS from SMOS for 2010–2013 (note that salinity is dimensionless); (b) annual average SSS from the Met Office FOAM model for 2015; (c) global SSS from the data available in the World Ocean Atlas (2005–2012); (d) annual average E–P (evaporation minus precipitation (myr⁻¹)); NCEP/NCAR reanalysis.)
of $h \approx 700$km, to obtain a spatial resolution of $d \approx 1$km (similar to that obtained from infrared for SST) requires an antenna size of $D \approx 147$m. Such an antenna size is clearly impractical to launch into space! Even reducing the spatial resolution requirement to $d \approx 10$km still requires an antenna of $D \approx 14.7$m.

The measurement requirements of high spatial resolution, which necessitates a large antenna, and a wide swath, to provide good spatial coverage and higher temporal resolution, led ESA to develop an interferometric synthetic aperture passive microwave sensor. This was based on an aperturization technique used in radio astronomy, where a large number of smaller antennae are used to synthesise a larger antenna (for a detailed description see Martin-Neira et al., 2014). In November 2009 ESA launched the SMOS mission with an interferometric synthetic aperture microwave radiometer consisting of 69 small antennae on three Y-shaped arms (see Figure 1). This design allowed the arms to be folded and so fit inside the nose cone of a launch vehicle. As a result of this design SMOS is able to measure SSS with a spatial resolution of 35–50km across a swath of about 1000km and provide complete coverage of the global ocean in 3 days. At the time of writing (mid-2017) SMOS continues to provide SSS data (and terrestrial soil moisture measurements, which it was also designed for).

Since the launch of SMOS, two further L-band passive microwave missions have been launched. The joint United States–Argentinian Aquarius mission for SSS was launched in June 2011 and unfortunately failed in June 2015, a year short of its design lifetime of 5 years. The United States SMAP (Soil Moisture Active Passive) mission was launched in January 2015 primarily to measure soil moisture, but it also provides data on SSS. These missions used simpler smaller single antennae that folded up for launch. Unlike SMOS, Aquarius made three ‘spot’ measurements rather than measuring across a swath, by having three receivers. Due to the antenna size each spot measurement was about 100km across, and global coverage of SSS could only be achieved every 7 days.

However, launching a suitable instrument into space to measure SSS is only the first step in the process of obtaining useful data. The measurements of brightness temperature depend not just on SSS, but also on SST and the wave conditions at the sea surface, including the effects of foam from breaking waves (Font et al., 2010). Furthermore, the sensitivity at L-band to SSS decreases with decreasing SST, making measurements of SSS more difficult in the colder ocean waters at higher latitudes (Köhler et al., 2015). One reason that the microwave frequency of 1.4GHz was chosen for making the measurements is that this is a protected frequency band used by radio astronomers to study the background cosmic and galactic radiation, and the reflection of that background L-band radiation from the ocean surface has to be accounted for when SSS is estimated from the measured brightness temperature. What rapidly became apparent from the first data obtained by SMOS is that there was considerable radio frequency interference (RFI) in the protected band from sources on the Earth’s surface (e.g. from some military RADARs). These signals are much stronger than the natural L-band emissions for the ocean and so must be removed from the data. In addition, land emits more L-band radiation than the ocean, making it a brighter target and so affecting SSS measurements near the coast. All these factors lead to errors making the measurement accuracy aim of the mission of 0.1 in salinity over 10–30 days and 100–200km very challenging. A recent review of SMOS performance concluded that the accuracy aim has not been met globally but mostly achieved in regions within 45°S–45°N with SSSs higher than ~10°C (Mecklenburg et al., 2016) for higher-latitude colder waters see Köhler et al., 2015; and for approaches to dealing with biases see Banks et al., 2016; Kolodziejczyk et al., 2016). Despite these challenges good SSS data are being obtained from SMOS, giving a global picture of the changing ocean salinity over the last 7 years.

One factor that needs to be considered concerning SSS measured from space is that the microwave emissions at L-band come from the top 0.5–1cm of the ocean surface (a similar ‘skin effect’ has long been known for infrared SST, where the infrared emissions come from the top micron (10−5m) of the ocean; the depth of this influence depends on the wavelength of the radiation). Therefore, care needs to be taken when comparing satellite SSS data with more traditional measurements from ships or buoys, or even the newer Argo floats, which typically measure salinity a few metres below the surface. For example, in heavy rain conditions the salinity at the sea surface may differ significantly from that a few metres below the surface. Nevertheless, comparisons with in situ data and with ocean forecast models, such as that run at the UK Met Office, have shown that the SMOS SSS data are of good quality and capture the known behaviour of ocean salinity.

**What can salinity measured from space tell us?**

The primary scientific benefit of SMOS is that for the first time oceanographers have global maps of SSS on timescales of days to a month and on space scales of 35–100km. Figure 2 shows global multi-annual averages of SSS from SMOS (2010–2013), the World Ocean Atlas (WOA; Zweng et al., 2013) using data for the years 2005–2012, and FOAM output for 2015 (the Met Office Forecast Ocean Assimilation Model, which assimilates Argo data; Storkey et al., 2010), together with global E–P from NCEP/NCAR reanalysis. It is clear that even with 8 years of in situ data the WOA SSS still has regions where SSS has not been measured. SMOS too has regions near the coast where it does not provide good SSS data due to the problems noted earlier (contamination from land and RFI). Some of the differences seen in Figure 2 between FOAM and SMOS near coasts may be due to these problems. Nevertheless, SMOS provides unprecedented coverage of the global ocean. The patterns of SSS seen in SMOS, FOAM and WOA are similar, with higher salinities visible in the evaporation-dominated subtropical gyres, where E–P is positive, and lower salinities in the high rainfall regions, where E–P is negative. Particularly noticeable is the effect of the high precipitation in the Inter-Tropical Convergence Zone (ITCZ) in the tropical Pacific.

Figure 3 shows a single month (September 2015) of global SSS from SMOS together with that derived from Argo floats. Similar patterns of high and low salinities are visible in both plots. However, the key point to notice is that the Argo data are much sparser, even though all the 1° squares in which an Argo measurement occurs at some point during the month are plotted. It is important to note that the spaceborne SSS and Argo salinity data are complementary, in that Argo provides subsurface salinity observations that cannot be obtained from space.

Rossby waves are ubiquitous in both the atmosphere and the ocean. In the ocean, they have previously been studied using spaceborne observations of changes in the sea surface height, from RADAR altimetry, and in SST. In Figure 4 anomalies of SSS, SST and sea surface height are shown as Hovmöller (longitude–time) plots at 15°S in the Indian Ocean, for longitudes 65°–100°E over 4 years (February 2010 to February 2014). Clearly visible are the diagonal patterns in the data indicating propagation of signals from east to west in the ocean. These have been previously identified as Rossby waves (or possibly ocean eddies) based on SST and sea surface height observations from space and, as expected, they are also visible in the SSS anomalies. The propagation speeds derived from these data are consistent with theoretical estimates of Rossby wave speeds, typically ~5–10cms−1 depending on latitude and Rossby wave mode (see figure 6 of Banks
et al., 2016). Detecting these signals in SSS would not have been possible using in situ measurements. Indeed, it is only with the advent of satellite observations that Rossby waves have been clearly detected in the ocean and found to be ubiquitous due to the global coverage that satellites provide. Another propagating signal that has been seen now in SSS, as well as SST and sea surface height, is that of Tropical Instability Waves that occur in the equatorial region of the Pacific Ocean (Lee et al., 2012).

The global SSS observations have also provided insight as to how major rivers, such as the Amazon, flowing into the ocean can have a major effect on ocean salinity (Tzortzi et al., 2013). For example, Figure 5(a) shows the annual range (maximum minus minimum) of SSS found in the equatorial Atlantic in 2010. Over most of the Atlantic from 20°N to 20°S the range is small (<1) but on the western side the salinity is influenced by the strong freshwater outflow from the Amazon and Orinoco rivers, while on the eastern side the Niger and the Congo rivers have a similar effect. The results show large variations in salinity during the year over large areas of the tropical Atlantic. This occurs every year over the SMOS period (not shown), just as it did in 2010 (Figure 5(a)). An interesting aspect that the SMOS SSS data combined with river flow and E and P data revealed is that the behaviour of the changes in SSS on the western and eastern sides of the tropical Atlantic is out of phase by 6 months (Figures 5(b) and (c)). The salinity is highest (lowest) on the western (eastern) side in February and lowest (highest) in August. However, in terms of the overall SSS for the tropical Atlantic region, these out of phase variations largely cancel out (Figure 5(c)).

A number of studies has investigated the freshening of the ocean surface due to precipitation through the use of both satellite and in situ salinity data, and these are summarised in Boutin et al. (2016). For freshwater to remain on the surface (within the ~1cm mentioned above) there must be little or no wind, as this would result in turbulent mixing of the surface freshwater with the more saline ocean waters. Several of the reported studies provide evidence for reductions in SSS related to rainfall in such conditions that correspond to studies using in situ measurements.

![SSS SMOS Sept 2015](image1)

**Figure 3.** (a) Global SSS from SMOS for September 2015 (note that lack of data south of 60°S is due to the presence of sea ice); (b) global SSS from Argo floats for September 2015 (note that every 1° square in which there is an Argo measurement is filled in).

![SSS Argo Sept 2015](image2)

![Hovmöller plots](image3)

**Figure 4.** Hovmöller plots at a latitude of 15°S in the Indian Ocean, showing anomalies in (a) SSS from SMOS; (b) SST provided with SMOS data; and (c) sea surface height from RADAR altimetry. The plots cover longitudes 65–100°E and the period February 2010 to February 2014. Clearly visible is the propagation of Rossby waves from east to west (diagonal striations in the plots). Dashed lines represent estimates of the Rossby wave propagation speeds from each of the observables (after Banks et al., 2016).
Salinity from space

To summarise, since the launch of SMOS in 2009 and the advent of Argo profiling floats in the mid-2000s a new era of ocean observations has commenced, where oceanographers are able to measure not only temperature but also salinity on a global scale. As temperature and salinity together determine the density of the sea water, this provides new opportunities for understanding the ocean circulation, from the mesoscale through to the gyre, basin and global scales. Some issues remain with regard to improving the quality, and utility, of SSS observations from space, including: RFI; biases in SMOS data (Banks et al., 2016; Kolodziejczyk et al., 2016); and difficulties in accurate retrievals close to land and at high latitudes (Köhler et al., 2015). However, ongoing research is making improvements in all of these areas, and future salinity missions are under consideration; the prospects are therefore good for continued measurements of SSS from space.

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On a larger scale the global water cycle and the SSS have been investigated using SMOS data, and a review is presented in Reul et al. (2014) that covers many of the topics discussed above. In addition, Reul et al. present materials on a variety of other topics, including monitoring eastern Tropical Atlantic freshwater pools and large-scale SSS interannual variability in the tropical Indian and Pacific Oceans, as well as fresh pool interactions with wind-driven processes (including tropical cyclones).

Figure 5. (a) SMOS SSS range for 2010 (max – min over the year) in the Tropical Atlantic, 20°N–20°S. Solid black lines show the boundaries of the western and eastern subregions defined such that the range in SSS is greater than 1.5; (b) mean SMOS SSS for January (top) and July (bottom) for 2010, with mean rainfall contours (myr⁻¹) overlaid. P contours for January are plotted every 0.5myr⁻¹, and those for July every 1myr⁻¹ (for clarity); (c) Area-weighted mean of SMOS SSS over the whole region 20°N–20°S (red), the western subregion (blue), and the eastern subregion (green) in 2010. (after Tzortzi et al., 2013).
Exploiting wind profiler information

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Introduction

Wind profilers are essentially Doppler RADAR instruments which sample the wind speed and direction at selected heights in the column of air vertically above the installation. Measurements repeated at regular time intervals can then be used to generate a continuous time–height profile of the wind vectors above that point.

In 1996 the European Cooperation in Science and Technology framework (COST) initiated project CWINDE (COST Wind Initiative Network Demonstration in Europe) to illustrate the benefits of a network of wind profilers providing near-real-time data. Figure 1 shows the location of a number of profilers which contributed data to CWINDE. Near-real-time data from this network was obtained from the Met Office (at http://www.metoffice.gov.uk/science/specialist/cwinde/profiler/) and used for this article. However, the CWINDE network has been superseded by the EUMETNET collaboration E-WINPROF, and European wind profiler data are now available from http://eumetnet.eu/activities/observations-programme/current-activities/e-profile/radar-wind-profilers/. More extensive data are also available through Met Office internal sources (personal communication from David Edwards of the Met Office).

In May 2005 the Met Office installed a wind profiler on the Isle of Man. The objective was to compliment the profilers already installed at Stornoway, Aberystwyth and

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