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Advances in the Observation and Understanding of Changes in Sea Level and Tides

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Abstract

Climate change, of which sea level change is one component, is seldom out of the news. This paper reviews developments in the measurement and understanding of changes in sea level and tides, focusing on the changes during the past century. The main aim has been to demonstrate how sea level and tidal science are now connected intimately with the fields of climate change and geodesy.

1. Introduction

This paper reviews recent research findings on sea levels and tides. It starts with some history of the measurement of sea level on a global basis by different techniques (tide gauges and satellite altimetry). It then summarises what has been learned about changes in mean sea level (MSL) during the past century, and considers the use of 'proxy' sea level information to place those changes in the context of variations in sea level on timescales of several 100 years or longer. The discussion then turns to how changes in extreme sea levels have differed from those in MSL. The following sections discuss the measurement of land levels by modern geodetic techniques and how sea level is now to be thought of as an essential parameter in geodesy. The importance of sea level to understanding changes in the ocean circulation is then referred to. Next, consideration is given to research on ocean tides and their recent changes. Finally, the paper ends with a mention of research that should be pursued in the future.

It is hoped that this overview may be a guide to anyone planning to investigate one of these research aspects further. Otherwise, one can point to recent books which can be consulted for fuller treatments (e.g. Pugh & Woodworth, 2014; Shennan et al., 2015; Gerkema, 2019), while for topics in sea level science which are not covered here (e.g. projections of future sea level rise or their impacts) one can refer the reader to assessments by international study groups (e.g. Church et al., 2013; Oppenheimer et al., 2019; Fox-Kemper et al., 2021).

2. Sea Level Measurements during the Instrumental Era

2.1 The Tide Gauge Era

The oldest long-term sea level records during the instrumental era all stem from northwest Europe. They include Amsterdam (from 1700, van Veen, 1945), Brest (from 1711, Cartwright, 1972a,b; Wöppelmann et al., 2006), Stockholm (from 1774, Ekman, 1988) and Liverpool (from 1764, Woodworth, 1999). Most of these historical records were obtained by visual measurements of water level using graduated scales cut into the sea wall adjacent to a sluice or dock gate, the scales functioning as tide poles (or tide staffs).[1] The records were obtained for their different operational reasons, and not especially for 'science', although the fact of their preservation implies that they were recognised as having long-term importance. In addition, they were far from being time series of regular hourly or similar levels as obtained using modern tide gauges. For example, at Stockholm measurements were as infrequent as about once a week or so during the early years of the record, while at Liverpool only the heights and times of high (and not low) waters were recorded. The Brest record from 1711-1716 was of both high and low waters, measured using a tide pole with special attention given to timing, and conducted for what we would now call scientific research.

Although the documentation on how these historical records were collected is not as complete as one would like, it seems that most of the visual measurements were made without the benefit of the float and stilling well apparatus which had been first discussed during the mid-17th century. The basic idea, consisting of a float within a stilling well and with the level of the float noted by an observer, had been suggested by Sir Robert Moray in the very first volume of the Philosophical Transactions of the Royal Society (Moray, 1666). A similar suggestion for a float gauge was made at about the same time by the German polymath and eccentric Athanasius Kircher (Kircher, 1665) (Figure 1). Examples of their use in temporary installations can be found throughout the 18th century (e.g. during the second voyage of James Cook, Woodworth and Rowe, 2018). However, while it may have been familiar technology by the time of the NW European records mentioned above, there may have been financial or engineering reasons why the apparatus was not more widely adopted in permanent installations, especially at locations with a large tidal range.

It was not until the 1830s that the first 'self-registering' gauges were installed in the Thames estuary in England (Palmer, 1831; Matthäus, 1972; Woodworth, 2015). In these devices, the level of the float was recorded on the paper chart of a chart recorder driven by a clock. By the end of the 19th century, most major ports were equipped with such instruments, providing important information on both tidal and non-tidal variations in sea level, including the long-term changes in MSL now employed in climate change studies. Nowadays, there are many types of tide gauge (float, acoustic, pressure, radar), each of which has its advantages for use in different locations (Pugh & Woodworth, 2014; IOC, 2016). In addition to their local applications, many of these gauges contribute their data to global sea level and tsunami programmes.

This global network is our main source of information on sea level change during the past century. Gauges in this network were installed for many different purposes including port operations, hydrographic surveying and flood warning as well for oceanographic research. They make up an inhomogeneous set with a well-known bias in recording towards the northern hemisphere (Wooodworth & Player, 2003). In recent years, considerable efforts have been made to install new gauges through the Global Sea Level Observing System (GLOSS) (IOC, 2012). This has included a focus on areas such as Africa (Woodworth et al., 2007). However, there is a considerable amount of advice available to agencies in any part of the world intending to install new gauges. Regional tsunami

[1] Descriptions of tide staffs and tide gauges can be found in the manuals on sea level measurement and interpretation published by the Intergovernmental Oceanographic Commission (https://www.psmsl.org/train_and_info/training/manuals/). A tide staff or similarly-graduated scale to measure water level (in this case sea level) is essentially the same simple apparatus with which river levels have been measured through the years (e.g. using column Nilometers, Friedman, 2014).

programmes such as that in the Caribbean have also contributed many gauges (von Hillebrandt-Andrade, 2013). Data from many of these tide gauges are contributed to international data centres including the Permanent Service for Mean Sea Level (PSMSL; https://www.psmsl.org) and the University of Hawaii Sea Level Center (UHSLC; https://uhslc.soest.hawaii.edu), and to global sea level projects such as the Global Extreme Sea Level Analysis (GESLA) data set (Woodworth et al., 2017a; Haigh et al., 2022; https://www.gesla.org) (Figure 2). The PSMSL is the main data centre for long-term information on MSL from tide gauges (Holgate et al., 2013) and its data set containing almost 75000 station-years of information underpins research by many groups. It is also an important source of sea level information for research assessments such as those of the Intergovernmental Panel on Climate Change Sixth Assessment Report (IPCC AR6) (IPCC, 2021).

2.2 The Altimeter Era

The technique of satellite radar altimetry can be traced back to a first instrument on Skylab in 1973. Subsequently, the Seasat and Geosat missions provided the first useful altimeter data sets for oceanography and geophysics. However, it was not until the early 1990s that the first of what can be called 'precise altimeter missions' were launched. These were the ERS-1 mission of the European Space Agency (ESA) and the US/French TOPEX/Poseidon (T/P) mission. The latter was particularly important, by being a major contributor of sea level information for oceanographers involved in the World Ocean Circulation Experiment (WOCE) (TPSWT, 1991). T/P was to be the first of the 'reference series' of missions, followed by the Jason series and now Sentinel-6, all measuring sea surface heights within 66° N/S every 10 days. A list of these and later altimeter missions can be found in Vignudelli et al. (2019a,b) (Figure 3).

Altimeter sea surface heights require a number of important corrections before they can be used for research: these include the so-called dry and wet troposphere, ionosphere and sea-state bias corrections (for more details, see Chapter 9 of Pugh & Woodworth, 2014). However, thanks to the work of processing centres, the data are provided in a form that researchers can readily make use of. As a result, it is no exaggeration to say that the availability of altimeter data has revolutionised research into ocean circulation, tides and long-term changes in MSL. Many research papers have been based on altimeter data sets, and excellent reviews are available (e.g. Fu & Cazenave, 2001; Stammer & Cazenave, 2017). Altimetry was complemented by the launch of the US/German Gravity Recovery and Climate Experiment (GRACE) satellite gravity mission in 2002 which for many years provided time series of changes in mass in the hydrosphere, cryosphere and ocean (Tapley et al., 2019). That space gravity time series is now being continued by the GRACE Follow-On mission launched in 2018 (Kornfeld et al., 2019), which has provided further insight into temporal changes in terrestrial freshwater and the cryosphere (Velicogna et al., 2020) and in ocean mass (Chen et al., 2020).

In addition, the mid-2000s saw the completion of a global-ocean collection of approximately 3000 profiling Argo floats for near-global ocean measurement of temperature and salinity (and sometimes other water properties) every 10 days, primarily within the upper 2000 m layers of the ocean (Roemmich et al., 2017, 2019). Therefore, the research community now has available not only a near-global system for monitoring sea level change, but also ways of observing changes in ocean mass and steric (density) changes in sea level, by means of which one can understand more completely the reasons for sea level change.

However, while altimetry has proved to be an excellent monitoring system for the open ocean, there are problems in obtaining data of the quality one would like close to the coast. This is primarily owing to contamination of the radar return signals by the presence of land within the altimeter footprint. In addition, any land within the area illuminated by a radiometer carried on the same satellite can result in an incorrect determination of the wet troposphere correction, although this second aspect is not as great an issue as the first, given that the wet correction can also be estimated using atmospheric models. There are also difficulties in making maximum use of altimeter data near to the coast due to the limitations of presently available tide models in shelf areas (see further below).

The requirement to do better near the coast has resulted in a whole new field of research being created called 'coastal altimetry' (Vignudelli et al., 2011, 2019a,b). It has spurred more sophisticated processing (retracking) of radar returns from existing and past missions, in order to remove as much as possible of the land contamination from the data (Gommenginger et al., 2011; Gouzenes et al., 2020; Climate Change Initiative Coastal Sea Level Team, 2020; Cazenave et al., 2022). As a result, accurate altimeter heights from the earlier missions are now available within approximately 5 km of the coast. In addition, the altimeters of later missions such as CryoSat-2, launched in 2010, Sentinel-3, launched in 2016, or Sentinel-6 Michael Freilich (also called Jason-CS), launched in 2020, have been designed with an aim of measuring reliably closer to the coast (for the global coastline for the latter two missions, only certain sections of coast for the former) (Parrinello et al., 2018; Vignudelli et al., 2019b) (Figure 3).[2] An important aim now is to make use of coastal altimeter information together with sea level data from conventional tide gauges within an overall coastal sea level monitoring system (Benveniste et al., 2019; Woodworth et al., 2019; IAT, 2021).

3. Twentieth-Twenty First Century MSL Rise and Acceleration

The PSMSL can be said to have been established in 1933 and it produced its first compilation of MSL records from tide gauges in 1940 in the Publications Scientifiques of the International Association of Physical Oceanography (available on the PSMSL web site). Gutenburg (1941) was the first to make use of this data set in a research paper. His main interest was in the rates of Glacial Isostatic Adjustment (GIA, formerly called postglacial rebound) in Scandinavia. However, a section of his paper also discussed the likely change in global-average sea level over the century leading up to that time (what is sometimes called 'eustatic sea level change', a term now discouraged because of its having different meanings in different studies, Gregory et al., 2019). He concluded that global MSL, outside of areas significantly affected by GIA, had risen by approximately 11 cm during the previous century.

Since then, many researchers have made use of the PSMSL data set to estimate global-average MSL change. The problems have always centred on the fact that it is a sparse data set with inhomogenous spatial and temporal coverage, including the northern hemisphere bias referred to above (Holgate et al., 2013). Important issues have always been how to combine the individual PSMSL records to compute regional and global averages. In addition, the data set reflects, of course, only coastal sea level change and not that of the deep ocean, which is why the advent of satellite altimetry was so important.

Tide gauges measure relative sea level change, that is to say relative to the land on which they are located, unlike altimetry which measures sea level in a geocentric reference frame. Therefore, their records have to be adjusted for vertical land movements, using information either from geodynamic models or from measurements of land level (relative to the centre of the Earth) using modern geodetic techniques. (This topic is discussed in more detail in Section 6.) Unfortunately, the only models of

[2] The main aim of CryoSat-2 was to measure ice thickness and its changes with time. Consequently, while the high spatial resolution achieved by its SIRAL altimeter, particularly along-track, has proved very useful for monitoring sea level in coastal areas, it operates in the required Synthetic Aperture Radar (SAR) mode only in certain regions. In addition, the satellite has a very long orbital repeat period of 369 days which can complicate time series analysis of sea level. Sentinel-3 operates continuously in SAR mode and has a repeat cycle of about 27 days. Sentinel-6 has a similar altimeter. Such SAR-mode altimeters can provide accurate sea level measurements as close as 1 km from the coast in the along-track direction.

vertical land movements available on a global basis are those of GIA (e.g. that of Peltier, 2004), which omits many other reasons why land level might change.

Nevertheless, one can simply combine all available tide gauge records to compute a time series of global average sea level change by correcting each record for vertical land movement using a geodynamic model of GIA. However, that approach would clearly bias the resulting averaged time series towards those regions with most records (i.e. Europe, North America and Japan). A second approach could involve the combination of regional-average time series into a global-average one (e.g. Douglas, 1991; Holgate & Woodworth, 2004), or of averages in latitude bands into global averages (Merrifield et al., 2009), or basin-averages into global ones (Thompson & Merrifield, 2014).

More sophisticated methods have tried to account for the spatial variations between sea level changes in different parts of the world. These methods are sometimes known as 'reconstructions'. They have employed spherical harmonics; empirical orthogonal functions (EOFs) of modes of ocean variability derived from satellite altimetry or from ocean circulation models; and cyclo-stationary EOFs that represent progressive motions in sea-level variations instead of the standing waves of conventional EOFs (Nakiboglu & Lambeck, 1991; Church & White, 2011; Ray & Douglas, 2011; Llovel et al., 2009; Hamlington et al., 2011). Other methods have been designed to average individual records in a region, or globally, without consideration of particular modes of variability. These include the 'virtual station' technique of Jevrejeva et al. (2006, 2008) in which individual records, which may be quite short, are successively combined into regional and global time series. All of these techniques have their own drawbacks that are inevitable when using a sparse data set. The advantages and disadvantages of some of them have been discussed by Calafat et al. (2014a), while Pugh and Woodworth (2014) provides a summary of techniques.

Figure 4 provides a comparison of three of the reconstructions available at the time of the IPCC Fifth Assessment Report (AR5) (Church et al., 2014), together with more recent reconstructions discussed below. It can probably be said that there are far more similarities between these different findings than there are differences, although the details of the differences, which are inevitable when using a global tide gauge data set that is limited both spatially and temporally, have resulted in quite a copious set of literature. (An example of such discussion can be found in Foster, 2019).

Most studies up until the AR5 produced estimates for global-average sea level change during the 20th century of between 1.4 and 2 mm/year. Later publications have included those of Hay et al. (2015) and Dangendorf et al. (2017) who reassessed the previous methods, particularly in the light of the possible contributing components of sea level change, including ice sheet melting and ocean warming, being unable to fully account for the earlier, higher estimates. Hay et al. (2015) obtained an average rate of 1.2 ± 0.2 mm/year during the 20th century up to 1990 using a probabilistic technique employing a Kalman Smoother (KS), while Dangendorf et al. (2017) obtained 1.1 ± 0.3 mm/year for essentially the same period using a virtual station technique. Dangendorf et al. (2019) then attempted to make use of a hybrid technique to use the best aspects of the KS method in combination with other approaches arriving at a value of 1.6 ± 0.4 mm/year for the period 1900-2015. The spread of these recent estimates obtained by modern methods shows how difficult it is to arrive at one agreed value for the 20th century trend in global average MSL.[3]

A detailed appreciation of the time-dependent spatial 'fingerprints' of all processes contributing to regional sea level change (ocean, glaciological, hydrological and solid earth) is essential in such studies

[3] A secular trend is conventionally expressed as a rate ± its standard error (i.e. 68% confidence level). However, a reader should be aware that 90 or 95% confidence level uncertainties are also often used. In this paper, I have simply reported the values given by the individual authors and anyone particularly interested in uncertainties should always check with the original publications.

(e.g. see Tamisiea & Mitrovica, 2011 for discussion of the regional fingerprints expected from the viscoelastic changes in the solid earth due to GIA and the elastic changes due to recent exchanges of water between continents and the ocean). The probabilistic and other methods employed by those later authors who suggest somewhat smaller values around 1.2 mm/year are more consistent with 20th century data on earth rotation than the larger, earlier estimates, and they go some way to reconciling the 'sea level enigma' of Walter Munk (Mitrovica et al., 2015). On the other hand, Adhikari et al. (2018) pointed out that this picture was not complete without consideration of the convective flows which drive the plate tectonics that also contribute to polar wander. Hamlington and Thompson (2015) and Thompson et al. (2016) suggested that the lower values obtained in more recent analyses could have been a consequence of station selection. However, Hay et al. (2017) maintained that their Bayesian fingerprint method was robust to different station selection. Meyssignac et al. (2017) and Hamlington et al. (2016, 2018, 2020) provided further recent insight into trends and variability in MSL on a regional rather than global level.

If one focuses now only on the last three decades for which altimeter data have been available, and so one might imagine that the spatial and temporal inhomogeneities in the measurement of global sea level are less important, an important thing to mention is that it remains essential to have as full as possible an understanding of the various corrections applied to the altimeter data, as mentioned above, and of the orbit of the satellite. There are also potential instrumental drifts in the altimeter itself to be aware of.[4] As a consequence, because the data sets are not perfect, they have to be revised by the processing centres from time to time as better knowledge is acquired on each term involved. With the most recent insights into these technical factors, one obtains a rate of global-average MSL rise from altimetry of 3.16 mm/year during 1993-2015 (SROCC, 2019) updated to 3.25 mm/year during 1993-2018 in the AR6.[5]

Most recently, the various estimates of sea level change from both tide gauges and altimetry have been discussed by the IPCC Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC, 2019) and by the IPCC AR6 (Fox-Kemper et al., 2021). The SROCC report concluded that global MSL increased from approximately 1.4 mm/year during 1901-1990, to 2.1 mm/year during 1970-2015, to 3.2 mm/year during 1993-2015 as already mentioned, and 3.6 mm/year during 2006-2015, these conclusions being based on the findings of the above papers. The latter were considered to be the highest rates during the instrumental era. The AR6 confirmed that global MSL rose faster in the 20th century than in any prior century over the last three millennia, with a 20 cm rise over the period 1901-2018. The rate of rise increased since the late 1960s, with an average rate of 2.3 mm/year over 1971–2018 increasing to 3.7 mm/year over the period 2006–2018.

[4] The only significant drifts in the main missions (the T/P and Jason series) have been in the radiometers which have since been carefully calibrated against independent data, and in the altimeter point target response in the early days of TOPEX (Beckley et al., 2017). The accuracy of these data sets is improved continually as more is learned about them. Tide gauges are employed at special sites for the ongoing monitoring of altimeter biases (see Section 6). In addition, the global tide gauge network is used to provide further checks on the altimeter data (Mitchum, 2000). On several occasions, this has flagged up problems with the altimeter information that have subsequently been corrected. Otherwise, it is important to note that measurements of, for example, trends in regional or global MSL by the two techniques are independent of each other.

[5] The latest estimates of rates of change in global and regional sea level from altimetry can be found at https://sealevel.nasa.gov/understanding-sea-level/key-indicators/global-mean-sea-level/ (Beckley et al., 2017). These rates are usually expressed after correction for the change in the geocentric position of the sea surface that one would expect from GIA. This correction depends on location but is approximately 0.3 mm/year averaged over the ocean, see Tamisiea (2011) or Tamisiea and Mitrovica (2011). In principle, there should also be small corrections for ocean bottom deformation due to present-day elastic processes (Frederikse et al., 2017) but these are not applied routinely at present.

The SROCC and AR6 findings show that there is never a linear rate of sea level rise but at times the rate is accelerating or decelerating. For example, most long tide gauge records demonstrate a slow acceleration in MSL between the 19th and 20th centuries of the order of approximately 1 mm/year per century (Woodworth, 1999; Woodworth et al., 2011; Hogarth, 2014), findings which are supported by evidence from proxy (i.e. salt marsh) data (e.g. Gehrels & Woodworth, 2013; Gehrels et al., 2020), as discussed in Section 7. However, the significance of that acceleration (in the tide gauge data at least) depends on the analysis method, some analyses finding that the evidence for acceleration is less persuasive than others (Watson, 2016a,b, 2017, and see discussion of methods for determining acceleration in Haigh et al., 2014, Visser et al., 2015 and Watson, 2021). The SROCC concluded that four out of five recent global sea level reconstructions extending back to at least 1902 indicated an acceleration of between -0.2 and +1.9 mm/year per century.

However, it is important to realise that on timescales of decades, rather than centuries, the rate of change of global-average sea level can vary considerably, just as global-average temperature can vary (e.g. Hamlington et al., 2013). Woodworth et al. (2009) presented evidence for positive acceleration, or 'inflexion', around 1920-30 and a negative acceleration around 1960 based primarily on inspection of individual long tide gauge records; this can be seen also in a number of the reconstructions of Figure 4. Some of this low-frequency variation has led to the suggestion of a ~60-year cycle in MSL (Chambers et al., 2012; Ding et al., 2021). Dangendorf et al. (2019) considered that there had been a persistent acceleration since the 1960s, mostly in the Indo-Pacific and South Atlantic Oceans, connected with changing wind fields leading to increased oceanic heat uptake and thereby sea level. Frederikse et al. (2018) showed that the observed trends and accelerations observed in each ocean basin in this period were consistent with the sum of possible contributing factors (the so-called 'budget') with the possible exception of the South Atlantic.

Studies of sea levels during the altimeter era since the 1990s have quantified the acceleration in global-average sea level during the last 30 years. This has been estimated (using the same units) as approximately 8 mm/year per century (Watson et al., 2015; Nerem et al., 2018) and is thought to be due primarily to an increase in Greenland mass loss since the 2000s (Chen et al., 2017; SROCC, 2019; Fox-Kemper et al., 2021). This evidence for the recent importance of Greenland is supported by data from the ICESAT and ICESAT-2 laser altimetry missions, which showed that melting grounded ice in Greenland and Antarctica had contributed a net 14 mm of sea level rise between 2003 and 2019 (Smith et al., 2020).

The various factors which contribute to the budget of sea level change include thermal expansion of the water column due to increased heat uptake by the ocean, and meltwater runoff from glaciers and the Greenland and Antarctic ice sheets. Many attempts have been made to quantify these various components (e.g. WCRP, 2018). For example, Table 4.1 of the SROCC indicates that for the period 1993-2015, thermal expansion was the largest contributor to global sea level rise (1.36 mm/year) followed by the other three contributors (0.56, 0.46 and 0.29 mm/year respectively). Confidence in estimating the budget accurately has been made possible in recent years due to the available of in situ thermal expansion information from the Argo network of floats (Roemmich et al., 2017, 2019) together with measurements of ocean mass from GRACE space gravity (Chambers et al., 2017). Changes in both of these factors are considered to be largely related to large-scale climate change. Consequently, the SROCC concluded that the dominant cause of global MSL rise since 1970 has been anthropogenic (see also Dangendorf et al., 2015; Slangen et al., 2016; Marcos et al., 2017).

In probably one of the most complete recent studies, Frederikse et al. (2020) reconciled, within the various uncertainties, the observed rates of global and regional MSL rise since 1900 and the factors responsible for them (i.e. the budget of thermal expansion, ice-mass loss and terrestrial water storage changes). The average rate of rise during 1900-2018 was estimated to be 1.56 ± 0.33 mm/year, in between the higher rates obtained prior to the AR5 and the lower values published since that time (Figure 4). The rate was found to vary over time, as does the relative importance of each contributing factor (Figure 5). Overall, however, glacier-dominated cryospheric mass loss caused twice as much rise in MSL as thermal expansion since 1900, once one takes into account 'uncharted glaciers' (Parkes & Marzeion, 2018). Glacier and Greenland Ice Sheet mass loss can explain the high rates of global MSL change during the 1940s, while a sharp increase in water impoundment behind dams was the main cause of lower rates during the 1970s. An acceleration in MSL since the 1970s was claimed to be caused by a combination of thermal expansion and increased Greenland mass loss. The fact that the importance of different contributing factors varies over time (especially the relative importance of dam impoundment versus climate-related processes) calls into question the validity of 'semi-empirical' parameterisations of sea level in terms of temperature alone, that seemed to offer a way forward in understanding and prediction some years ago (e.g. Rahmstorf, 2007; Vermeer and Rahmstorf, 2009), although some semi-empirical methods have attempted to take into account temporal changes in the partitioning of contributions to sea level rise (e.g. Mengel et al., 2016). Work of this type includes the 'transient sea level sensitivity' study of Grinsted and Christensen (2021) who as a result suggest that the upper level of sea level projections in IPCC reports is too low.

In summarising all these recent findings, the AR6 concluded that the measurement of MSL rise during the 20th-21st centuries, and the reasons for the rise, are now understood better than ever. The new observation-based estimates published since SROCC lead to an assessed sea level rise over the period 1901 to 2018 that is consistent with the sum of all the individual components and that role of each component varies. In particular, while ocean thermal expansion (38%) and mass loss from glaciers (41%) were considered to have dominated the total change from 1901 to 2018, ice sheet mass loss has increased and accounts for about 35% of the sea level increase during the period 2006–2018.

Where does this lead to for the remainder of the 21st century? The AR6 modelling suggested that global-average sea level could rise by 0.28–0.55 m by 2100 (relative to 1995–2014), assuming a very low greenhouse gas emission scenario, or 0.63–1.01 m, assuming a very high scenario. This rise is caused primarily by thermal expansion and mass loss from glaciers and ice sheets, with minor contributions from changes in land-water storage. (These likely range projections do not include those ice-sheet-related processes that are characterized by deep uncertainty.) Such a rise in sea level, combined with possible changes in the pattern of frequency and magnitude of storms and their associated storm surges, will require major expenditure in raising coastal defences, with sometimes consequent undesirable modifications to coastal environments. Moreover, in some cases, such as small island states, sea level rise will represent a major threat to the viability of the state itself.

4. Proxy Sea Level Measurements

The previous sections have shown that the growth of the global tide gauge network, the apparent increase in the rate of sea level change, and the onset of recent climate change due to industrialisation all occurred within a few decades around the end of the 19th century. Therefore, much longer records are needed in order to establish a baseline of long-term sea level change against which more recent climate-related changes can be compared. Unfortunately, we have seen that there are few tide gauge records that extend back to the early 19th or 18th centuries, and they are anyway confined to Europe. Consequently, it is important to investigate other sources of information by what are sometimes called 'proxy' sea level techniques that can provide useful data on sea level change on timescales of several 100 years or more, and which ideally have a geological coverage wider than Europe. Two such sources (salt marshes and archaeology) will be mentioned below. Discussion of the use of corals and other proxy techniques may be found in the 'handbook' of Shennan et al. (2015) and more recent publications (e.g. Sisma-Ventura et al., 2020).

Salt marshes are intertidal areas in which layers of sediments have accreted and now support a wide variety of rushes, sedges and grasses. When sea level rises over many years, sediments accumulate and the marsh surface builds up vertically. As a result, a history of sea level rise can be reconstructed from cores of sediment obtained from the salt marsh. Most reconstructions are from single cores, usually obtained from high marsh settings. The method works less well in low marshes where sedimentation rates can be too rapid to be resolved by dating (as mentioned below) and where the vertical ranges of different types of microfossils (foraminifera and diatoms) are much larger. These microfossils and plants within the different layers of sediment allow a relationship with former sea level (or 'indicative meaning') to be established by comparison to the distribution of microfauna and plants on the modern coast obtained from present-day surface samples. The approach clearly depends critically on an assumption that the vertical distribution (relative to a chosen tide level) of fossil remains in the historical and modern marshes are the same. The need for continuous sedimentation means that marsh areas prone to erosion and highly fluctuating sedimentation rates, such as the lower salt marsh and areas near tidal creeks, are to be avoided. There are many technical aspects involved in analysis of sediment cores before useful data can be obtained (e.g. how to handle problems of sediment compaction), although considerable experience in best practice has been acquired over many years (e.g. since van de Plassche, 1986). Modern insight into such technical aspects are discussed in detail in Edwards and Wright (2015) and several other chapters in Shennan et al. (2015), and also by Barlow et al. (2013) and Gehrels and Kemp (2021).

Salt marsh records are obtained primarily (but not exclusively) from mid-latitude locations with some records spanning many hundreds or a thousand years or more, with the temporal resolution of a core depending on the rate of sedimentation. [6] For marshes with a high rate, shorter timescale changes can be resolved, and the resulting sea level record spanning recent decades can be compared to the instrumental information from a nearby tide gauge (e.g. Gehrels & Woodworth, 2013). The dating of various layers of a given core can be aided by different methods depending upon where the core was obtained. They include radiometric analyses (^{14}C, ^{210}Pb) and stratigraphic marker techniques (e.g. ^{137}Cs, pollen, charcoal, Pb isotopes, metal concentrations). For example, the ^{137}Cs method can be used to detect the time in the mid-1960s when nuclear bomb testing was at its peak. In some regions (e.g. the southwest Pacific and the Atlantic coast of North America), pollen in the cores can provide time markers by indicating changes in vegetation following settlement by Europeans during the 18th and 19th centuries. The availability of as many such markers in a core as possible is highly desirable when its information appears to suggest rapid changes in sea level. For example, the traditional method of radiocarbon dating does not work well in sediments younger than 300 years, and ^{210}Pb dating, which is commonly applied to young sediments, also has its limitations (Marshall et al., 2007).

The large and relatively pristine estuaries of the North American Atlantic coast have always been important locations for salt marsh measurements (e.g. Donnelly et al., 2004). However, recent cores discussed in a series of papers by Kemp et al. (2011, 2013, 2015) have improved significantly on the spatial extent of measurements along that coastline from Florida to Connecticut, temporal coverage (~2000 years) and temporal resolution. These records have been used to demonstrate the temporal variations in the local rates of sea level rise, how contrasting rates are to be found on the American and European sides of the North Atlantic (Long et al., 2014), and how different rates either side of Cape Hatteras can be related to ocean dynamics (Woodworth et al., 2017b). Kopp et al. (2016) and Kemp et al. (2018) used these data together with other salt marsh information from around the world, and also with tide gauge data, in a study of temperature-driven global sea level variability in the last 2000 years (Common Era).

[6] Salt marsh records are obtained primarily from locations where sea level is rising. However, they can also be used where sea level is falling as long as sediment is accumulating where the cores are obtained. That is usually the case for cores in the middle or high marsh but below highest astronomical tide. Gehrels and Kemp (2021) provide some examples.

Gehrels et al. (2020) used some of the previously published information for this coastline, together with their own recent measurements at Nova Scotia, Maine and Connecticut spanning the past 500-1000 years, showing the now unquestionable acceleration in 20th century sea level rise above the longer-term baseline, qualitatively consistent with the accelerations in sea level between the 19th and 20th centuries observed in tide gauge data (Figure 6a). However, they also pointed to evidence for a rapid rise in the 18th century for Nova Scotia, Maine and Connecticut, almost as rapid as in the 20th century (Figure 6b), suggesting a similar earlier experience in this potential 'hot spot' area of sea level rise (Yin et al., 2009; Sallenger et al., 2012; Wise et al., 2020). The authors suggested possible connections to the North Atlantic Oscillation and to changes in the cryosphere and ocean circulation. This hotspot interpretation has since been questioned by other analysts (Walker et al., 2021). The same group have suggested that modern sea level rise commenced several decades before the turn of the 19/20th centuries, consistent with early global warming and glacier melt (Walker et al., 2022).

An advantage of the salt marsh technique, additional to its temporal coverage, is that it can be applied to parts of the world without any historical tide gauge information, notably in the southern hemisphere, thereby extending the geographical coverage of our knowledge of sea level change. Aside from the North American studies mentioned above, examples include measurements as far apart as New Zealand and Australia (Gehrels et al., 2008, 2012; King et al., 2020), Greenland (Long et al., 2012), Iceland (Saher et al., 2015), SW California (Avnaim-Katav et al., 2017, although this paper does not include an actual sea level reconstruction), Norway (Barnett et al., 2015), NW England and Scotland (Mills et al., 2013; Barlow et al., 2014), northern Spain (Leorri et al., 2008) and Croatia (Shaw et al., 2018). And for many other examples, see the review of Gehrels and Kemp (2021). Although some of the cores in these studies lack the temporal coverage and resolution of the North American ones, they can still be important in establishing baselines of change to which those of the instrumental era can be compared. In addition to climate-related sea level changes, microfossils have for some time also been employed in research into the rapid changes in land level due to great earthquakes (e.g. Zong et al., 2003; Engelhart et al., 2013).

A second 'proxy' technique (although again largely confined to Europe) involves the use of Roman, Greek and medieval archaeological markers (e.g. fish ponds, quaysides and wells) to infer historical sea levels, or at least to provide upper or lower limits on change. This has been an active of research for many years in the Mediterranean where there are both many such markers and where the tidal range is relatively small, enabling more reliable determination of historical MSL (e.g. Flemming, 1978). For example, Sivan et al. (2004) reported on variation in the water level in Roman-Byzantine-Crusader wells close to the coast of Israel, with the conclusion that ocean volume has been essentially the same for the past 2000 years. In addition, one of most important archaeological sources of sea level information comes from Roman fish ponds. Lambeck et al. (2004) used such data from the Tyrrhenian coast in combination with nearby tide gauge information, concluding that modern sea level rise commenced between 1850 and 1950. More recently, Lambeck et al. (2018) used the fish pond information from the same coastline, concluding that local sea level during the Roman Period between about 2100–1900 BP was about 1.20 ± 0.20 m below present. Archaeological sea level research obviously has its limits in other areas where there are fewer markers and tidal ranges are larger. Nevertheless, such attempts have been made in the past (e.g. in southern Britain by Akeroyd, 1972). Novel ideas, such as study of paintings that contain historical sea level information can also be considered a useful sub-branch of archaeological sea level research (Camuffo & Sturaro, 2003; Camuffo et al., 2017).

5. Changes in Extreme Sea Levels

An important question with regard to studies of impacts of sea level rise concerns whether extreme sea levels are changing at rates similar to those in MSL. Sea level extremes tend to occur mostly during tropical cyclones at lower latitudes and winter storms at mid-latitudes, although, of course, the tide can also be a major contributor to an extreme (Pugh & Woodworth, 2014, Chapter 7). The question is important because predicting future MSL change is uncertain enough (e.g. Church et al., 2014), and if extremes were to changes quite differently to MSL, then it would make the prediction of future extremes even more difficult.

Reviews of findings on extremes can be found in Lowe et al. (2010), Seneviratne et al. (2012), Church et al. (2014), SROCC (2019) and Fox-Kemper et al. (2021).As far as one can tell, extreme sea levels in most regions are indeed increasing in line with MSL. Menéndez and Woodworth (2010) were the first to investigate this topic, since largely confirmed by other authors using tide gauge records from particular locations or in different regions (e.g. Calafat et al., 2022). Menéndez and Woodworth (2010) also demonstrated the dependence of extremes on MSL on interannual timescales (e.g. due to the El Niño Southern Oscillation, ENSO), as well as on long-term MSL change. However, while changes in extreme sea levels and MSL are clearly related, there are other factors at work. For example, Marcos and Woodworth (2017, 2018) showed that for about a quarter of stations around the North Atlantic coast, significant changes in extremes still remained even after accounting for MSL changes (Figure 7, see also Marcos et al., 2015).

However, considerations of MSL change aside, the tide itself remains an important factor in the variation of extremes sea level on interannual timescales. Menéndez and Woodworth (2010) showed how extreme sea levels can also occur on nodal and perigean timescales (i.e. 18.6 and approximately 4.4 years respectively), the latter being explainable in harmonic terms for semidiurnal regimes largely as a result of a combination of the M2, N2, S2 and K2 constituents, perigean variations in diurnal regimes being somewhat less (Haigh et al., 2011; Ray & Foster, 2016; Ray & Merrifield, 2019; Baranes et al., 2020; Li et al., 2021).

Another aspect of this area of research concerns the provision of reliable extreme sea level distributions for the entire global coastline. These distributions provide information on the frequency of given levels being exceeded during any one year, in other words the Return Period (or Average Recurrence Interval) for a given level at any given location. Such distributions are often described adequately by Gumbel distributions (Gumbel, 1941) or by more general Generalised Extreme Value (GEV) and Generalised Pareto Distribution (GPD) parameterisations (Arns et al., 2013; Wahl et al., 2017). Several decades of annual maximum sea level values obtained from a tide gauge record are usually adequate to determine the Gumbel parameters at each gauge location; the question is then how to obtain reliable estimates of parameters between locations. The solution would appear to be provided by numerical modelling of tides and surges on a global basis (e.g. Muis et al., 2016), even though Calafat et al. (2014b) indicated that models have limited capability to capture the magnitude of extremes in some cases. Such future modelling would include consideration of interannual ocean variability, and be validated as far as possible by the observational evidence from tide gauges (Hunter et al., 2017). Alternatively, on a regional basis one can consider approaches that couple sparse tide gauge observations with numerical model outputs and apply bias correction schemes (e.g. Arns et al., 2015). In addition, Calafat and Marcos (2020) have presented a Bayesian spatiotemporal model that leverages spatial dependences in extremes to compensate for data sparseness along coastlines, allowing estimation of extreme event probabilities at ungauged locations with a precision comparable to that of traditional approaches at observed stations. However, the method will clearly not work in areas without any data at all.

Extreme level distributions have immediate applications in present-day coastal engineering (Pugh & Woodworth, 2014, Chapter 12). However, another application is in determining the likely increases in the frequency of given levels being exceeded following a future rise in MSL, with an assumption that the statistics of the meteorological forcings that result in storm surges do not change, and also in the calculation of ‘allowances’ for sea level rise, which are the amounts by which defences need to be raised in order to provide the same likelihood of coastal flooding following a rise in sea level (Hunter, 2012; Slangen et al., 2017; Woodworth et al., 2021a). Unfortunately, the distributions used in some previous studies of impacts of MSL change on a global basis were seriously in error resulting in incorrect estimates of impacts, as pointed out by Hunter et al. (2017) and Muis et al. (2017). The new sets of global modelling referred to above represent considerable progress on previous catalogues of extreme sea level distributions for the world coastline.

Other particular areas of interest in studies of extremes concern the way that surge and tide interact in shallow waters so as to reduce an overall extreme (Bernier & Thompson, 2007; Horsburgh & Wilson, 2007; Arns et al., 2020), the way that surge, tide and river flow interact in rivers and estuaries (e.g. Familkhalili & Talke, 2016), and the spatial scales of storm surges and extremes sea levels along a coastline (Enríquez et al., 2020; Rashid & Wahl, 2020). Modifications to the seasonal cycle of MSL can also impact on sea level extremes (Wahl et al., 2014). Also of importance to the estimation of extreme sea levels is consideration of the higher-frequency variability that is often missed by the hourly and similar sampling of some tide gauge data. This sub-hourly variability includes seiches generated by tides, winds or waves which can be several decimetres in magnitude (e.g. Vilibić & Sěpić, 2017; Woodworth, 2017; Pugh et al., 2020; Zemunik et al., 2022). The role of wave setup as well as the usually modelled wind setup is also an important consideration within the overall description of a storm surge (Melet et al., 2018; Woodworth et al., 2019; Woodworth, 2020), even though such an estimation may be a difficult undertaking at the large scale.

Sometimes one wishes to study extreme non-tidal variability (or ‘extreme surge’), rather than the extremes in total sea level, and for this one can choose to work with the residuals of a tidal analysis of the sea level record, or instead to compute skew surges at each high astronomical tide (Pugh & Woodworth, 2014). One has to be careful when using residuals as older data especially can be affected significantly by bad tide gauge timing. On the other hand, skew surges rely only on reliable measurement of high water level and not on its exact timing. Therefore, they tend to be a more robust quantity. Examples of their use include Williams et al. (2016) and Marcos and Woodworth (2017).

6. Sea and Land Level Changes

As mentioned above, tide gauges measure sea level relative to the height of the land on which they are situated, unlike satellite altimeters which measure sea level relative to the centre of the Earth. Therefore, if the two data types are to be compared in the same analysis, then the tide gauge information must be expressed in the same geocentric reference frame as the altimeter data. This is done using Global Navigation Satellite System (GNSS) measurements to find the geocentric height of the tide gauge benchmarks (reference levels of the land on which the gauges are located). In this way, one can make use of the tide gauge data at special sites, where measurements by the two techniques are obtained from the same or nearby locations, to calibrate any biases in the altimeter sea surface heights (e.g. Bonnefond et al., 2010; Haines et al., 2010).

In addition, if one has access to extended time series of geocentric vertical land movement using GNSS at each tide gauge, then one can combine the GNSS-derived time series with that of relative sea level (i.e. relative to the land) from the tide gauge itself, thereby obtaining time series of GNSS-adjusted sea levels at each location. These time series would then, in principle, be free from contributions from geological processes which otherwise could lead to a misinterpretation of land level change as real sea level change. The geological processes that result in land level changes include the low-frequency, global-scale one of GIA, and abrupt, but more localised, changes due to earthquakes or to anthropogenic extraction of subterranean oil or water. There are already readily available global models of GIA (Tamisiea & Mitrovica, 2011). However, while the use of a GIA model to correct tide gauge data for vertical land movement is a reasonable first approximation, it cannot substitute completely for a measurement of the total land movement at a gauge site.

Many thousands of GNSS receivers have been deployed around the world, primarily for providing positioning information for many practical applications, but also for scientific research. The latter include receivers which function as seismometers for earthquake and tsunami monitoring (Blewitt et al., 2006). In addition, many are located alongside tide gauges for the purposes explained above, and there is now a requirement of the GLOSS programme that all of the gauges in its network be equipped with GNSS receivers (IOC, 2012).

A particular technical problem at some tide gauge stations is that the GNSS receivers are not located exactly at the gauges (as shown in Figure 8) but sometimes at a considerable distance away (e.g. several km). In all cases, it is essential that the datums of the gauge and the receiver (in effect the benchmarks at the two locations) are related by what is called a 'geodetic tie'. This tie is usually made by means of conventional spirit levelling between the two, although nowadays it is just as likely to be made using a GNSS connection (plus a small local adjustment for the geoid difference). The tie must be repeated at regular intervals (e.g. every two years) as differential land level changes between the two can occur. Unfortunately, these sorts of routine measurements are often the hardest to ensure as the responsibility for them is split between tide gauge and geodetic agencies (Woodworth et al., 2017c).

The state of the art in geodetic monitoring of land levels for sea level studies has been reviewed by Pugh and Woodworth (2014, Chapter 8) and Wöppelmann and Marcos (2016). Other techniques have been considered for this application and have been tried in the past including absolute gravity (Teferle et al., 2006) and DORIS (Doppler Orbitography and Radiopositioning Integrated by Satellite, Cazenave et al., 1999). However, neither technique has provided the benefits of GNSS in terms of cost, global coverage and accuracy. The geodetic technique with apparently the most complementarity to GNSS is InSAR (Interferometric Synthetic Aperture Radar) which provides regional maps of rates of vertical land movement at high spatial resolution (e.g. Filmer et al., 2020). Zerbini et al. (2018) provide an example of how networks of GNSS stations, together with information from InSAR and historical levelling campaigns, can provide detailed information on regional patterns of vertical land movements, including those at the coast. Other recent studies using InSAR include those of Mahapatra et al. (2018), Blackwell et al. (2020), Buzzanga et al. (2020) and Shirzaei et al. (2021).

Considerable progress has been with GNSS, and a global databank called SONEL (Système D'Observation du Niveau des Eaux Littorales, https://www.sonel.org) now exists from which one can obtain time series of GNSS data near to a large number of tide gauges. Quasi-global maps of vertical land movements are available from the global network of GNSS receivers (Hammond et al., 2021). However, many problems remain to do with network coverage and measurement accuracy (Marcos et al., 2019). In particular, the accuracy and stability of the global reference frame in which measurements are made (called the International Terrestrial Reference Frame, ITRF) continue to be a major concern (Blewitt et al., 2010). In addition, although receivers are now more affordable than they were some years ago, there is still a need to have them installed at every tide gauge.

The latter deficiency may be rectified partially should the technique of GNSS reflectometry, in which a single receiver provides both sea (i.e. tide gauge) and land (i.e. vertical land movement) information, was to be employed more widely alongside, or instead of, conventional tide gauges (Larson et al., 2013, 2017, 2021; Williams et al., 2020). Figure 9 compares monthly MSL time series at 6 sites obtained from GNSS reflectometry to those from nearby conventional tide gauges, showing that the technique is capable of providing valuable new sets of sea level data.

7. Sea Level and Geodesy

Sea level has been considered as a geodetic parameter for many years. For example, in the 19th century many countries adopted MSL, measured over several years by tide gauges at one or more sites, as national land survey datums (e.g. Bradshaw et al., 2016; Wöppelmann et al., 2014). However, it was not until the advent of new, largely space-based technologies in recent years that the importance of MSL at particular locations, and of the Mean Sea Surface (MSS) generally, began to be fully appreciated within a geodetic consideration of the planet as a whole.

There are said to be 'Three Pillars' of geodetic research (Rummel et al., 2002). They comprise the geometric shape of the Earth, the gravity field and its temporal variations, and the orientation of the Earth in space as a function of time. Measurements associated with each pillar are nowadays the responsibility of the Global Geodetic Observing System (GGOS) (Plag & Pearlman, 2009).

As regards the first pillar in the context of sea level research, GNSS now enables the position of any point on land or sea (or even in flight) to be expressed within an ITRF (Altamimi et al., 2016). The previous section explained how rates of vertical movement of the land on which tide gauges are located can now be monitored with GNSS, and that tide gauge MSL can be measured within the same reference frame as satellite altimeter data, enabling tide gauge and altimetric sea surface heights to be studied together. As a result, the availability of GNSS means that a combined MSS product (i.e. a global map of MSL at all points in the ocean, measured relative to the centre of the Earth, and calculated for a certain period of time) can now be derived with centimetric accuracy (e.g. Andersen & Knudsen, 2009).

The second pillar concerns the gravity field which can be considered from two perspectives: a time-averaged field which provides a model of the geoid with high spatial resolution that can be used in combination with the above MSS; and a time-dependent field which largely reflects temporal changes in the cryosphere, terrestrial hydrosphere and ocean.[7] A review of how the gravity field has been measured through the years by various types of satellite tracking in combination with marine and terrestrial gravity data, and how measurements have been revolutionised following the launch of several remarkable gravity missions (CHAMP, GRACE and its Follow-On, and GOCE), is given by Pugh and Woodworth (2014, Chapter 9). For example, the GOCE mission had the aim of enabling the geoid to be known to 1-2 cm accuracy for wavelengths larger than about 100 km. The importance of improved knowledge of the time-averaged geoid is that it can be subtracted from the MSS to give the Mean Dynamic Topography (MDT), which is the component of the MSS with a magnitude of ±1 m due to the ocean circulation. Figure 10(a) demonstrates the large-scale MDT obtained from knowledge of the MSS and geoid supplemented by information from various ocean data sets (Mulet et al., 2021). The geostrophic ocean currents corresponding to that MDT are shown in Figure 10(b).

A second important application of an accurate global geoid model is that it leads towards World Height System Unification, a universal datum (i.e. the geoid) capable of replacing the various national datums mentioned above (Rummel, 2012; Woodworth et al., 2012; Sideris, 2014). Research in this area has led to an explanation of why MSL, measured with respect to a national levelling system or to a model of the geoid, appears to vary along coastlines as a response to the nearby ocean circulation (Featherstone and Filmer, 2012; Woodworth et al., 2012; Higginson et al., 2015; Lin et al., 2015, 2021).

[7] As a consequence of changes in the cryosphere etc., the 'time-averaged' marine geoid will itself change slowly through time. This should be taken into account in ocean circulation studies based on satellite altimeter data referenced to a geoid field (Siegismund et al., 2020).

Meanwhile, the importance of GRACE for the better understanding of temporal changes in sea level has been discussed above.

The third pillar relates to sea level studies insofar as the budget of sea level change must be consistent with the observed changes in Earth rotation. As an example, Mitrovica et al. (2015) argued that a combination of the more recent lower estimates of the 20th century MSL rise by Hay et al. (2015), together with improved modelling of GIA, and correction of the eclipse record for a signal due to angular momentum exchange between the fluid outer core and the mantle, means that three quantities: (i) Earth rotation changes over the last three millennia which demonstrate a long-term slowing of the Earth's rate of rotation, (ii) amplitude and (iii) orientation of the Earth's rotation vector over the past century, can now be reconciled with each other. This reconciliation thereby resolves the 'sea level enigma' discussed by Munk (2002).

8. Sea Level and Ocean Circulation

The availability of satellite altimetry data sets since WOCE has resulted in a vast improvement in the understanding of the temporal and spatial scales of variability in the ocean circulation. Furthermore, analysis of altimeter data in combination with meteorological information and other ocean data, including various types of in situ information (Roemmich et al., 2017, 2019), has resulted in a greater understanding of what forcings are responsible for that variability. At first, most papers in this field were concerned with the largest sea surface height signals, such as to do with ENSO, but now virtually all analyses of deep ocean and coastal circulation make some use of altimeter data, often in combination with data from other ocean instruments (e.g. in the OSNAP monitoring of the North Atlantic, Lozier et al., 2019). A summary of findings can be found in the many chapters of Fu and Cazenave (2001) and Stammer and Cazenave (2017).

Some tide gauges have been shown to record sea level variability that differs from that measured in the nearby deep ocean by altimetry (Woodworth et al., 2019). Nevertheless, it would be incorrect to conclude that tide gauges cannot continue to play an important role in providing useful insights into many aspects of wider ocean dynamics, more so than altimetry in some aspects. Examples include the relationships between Antarctic sea levels and variability in the Antarctic Circumpolar Current (ACC) (Hughes et al., 2003; Meredith et al., 2004) and between European coast sea levels and eastern boundary currents (Calafat et al., 2012). Many other publications have featured data from the North American Atlantic coast, partly because of the interesting oceanography in the region, and also thanks to the quantity and quality of regional tide gauge data. Examples include links between variability in coastal sea levels and off-shore flows (Florida Current, Gulf Stream and Atlantic Meridional Overturning Circulation, AMOC) including major differences in variability due to flows north and south of Cape Hatteras (Bingham & Hughes, 2009; Thompson & Mitchum, 2014; Woodworth et al., 2014; Ezer, 2015; Little et al., 2019; Piecuch, 2020); links between variability and the North Atlantic Oscillation (Andres et al., 2013; McCarthy et al., 2015); and modulations of annual sea levels by deep ocean Rossby waves (Calafat et al., 2018). In addition, the superior temporal sampling of tide gauges has obvious advantages over altimetry for monitoring flows through straits such as Gibraltar (Garrett et al., 1999; Gomis et al., 2006) or within the Florida Current (Meinen et al., 2021), while bottom pressure recorders (BPRS, or deep-sea tide gauges) have proved capable of monitoring the variability of transports through the Drake Passage (Meredith et al., 2011).

BPRs were invented in the 1970s and first deployed for the measurement of deep ocean tides (Cartwright, 1999), although they were since employed in studies of non-tidal ocean processes from low (e.g. Cartwright et al., 1987) to high (e.g. Vassie et al., 1994) latitudes, including the monitoring of the Drake Passage and other 'choke points' of the ACC during WOCE. More recently, it has been noted that ocean dynamical signals in the deep tropical ocean are so small that bottom pressure measurements there can be used to determine the annual cycle of mass exchange between ocean and land (Hughes et al., 2012, 2018; Williams et al., 2014; Hsu & Velicogna, 2017). In addition, bottom pressure measurements along the continental slope have been shown to be capable of monitoring basin-scale dynamics (e.g. the AMOC), uncontaminated by the mesoscale variability that dominates most other ocean measurements (Hughes et al., 2018).

9. Ocean Tides

The availability of precise satellite altimeter data in the early 1990s revolutionised the development of regional and global tide models (Andersen et al., 1995; Shum et al., 1997). Some of these models were obtained by means of harmonic analysis of the altimeter sea surface heights at each point in the ocean, essentially as one would analyse a tide gauge record. In other words, they provided a mapping of derived tidal amplitudes and phase lags for each constituent but without any dynamical constraints. Other models have been based on assimilation of altimeter (and sometimes tide gauge) tidal information into barotropic tide models. As a result, they provide information on depth-averaged tidal currents as well as tidal elevations (e.g. Egbert et al., 1994; Taguchi et al., 2014).

It is important to realise that an altimeter measures the geocentric tide which is the combination of the ocean tide, earth load tide and solid earth body tide. The latter can be computed straightforwardly from the tidal potential and subtracted from the geocentric tide. The sum of the first two comprise what is called the elastic ocean tide. The load tide can be computed to good accuracy if one has a good a priori tide model (e.g. Francis & Mazzega, 1990). The ocean tide can thereby be determined by subtraction of the load tide from the altimetry-derived elastic ocean tide.

Stammer et al. (2014) undertook a more recent assessment of the available set of ocean tide models. One of its tests was to calculate differences in tidal heights from the eight major constituents in the models and the heights derived using the same constituents obtained from analyses of in situ (tide gauge) data. The best models showed root-sum-square agreement of approximately 0.9, 5.0, and 6.5 cm for pelagic (deep ocean), shelf and coastal regions, respectively. This demonstrated the centimetric accuracy for the best models over most of the deep ocean; Figures 11(a,b,c) present maps of the M2 component of the ocean tide from one of the most recent models (Lyard et al., 2021). However, the modelled tides were found to be considerably less accurate close to the coast, where problems remain due to the inherent limitations of spatial and temporal sampling by altimeters, technical issues to do with land contamination in altimeter and radiometer footprints, and the fact that coastal tides are larger and more complex than those of the deep ocean with a multiplicity of shallow-water constituents (Ray et al., 2011; Woodworth et al., 2019). Tidal models are also likely to be less accurate in high latitudes, including under ice shelves, where there is a limited amount of suitable information from altimetry (Ray et al., 2020).

Many aspects of tidal research are included in papers in a special issue of Ocean Science (Woodworth et al., 2021b). Otherwise, a summary of progress in measuring tides with altimeter data was provided by Ray and Egbert (2017). This includes an updated calculation of the energy budget of the barotropic tides in the global ocean discussed previously by Egbert and Ray (2000, 2001). Third-degree tides have also attracted recent interest, being mapped on a global basis using tide gauge data and models (Woodworth, 2019) and also, remarkably given how small they are (typically several millimetres in amplitude), using satellite altimetry (Ray, 2020). Research is now turning to methods for exploitation of the next generation of altimeters for tidal research (e.g. SWOT now due for launch in 2022, Morrow et al., 2019). These will provide considerably greater information on tidal variations on short spatial scales. In addition, much research is now being focused on internal tides. As examples, a series of papers by Zhao and colleagues (e.g. Zhao et al. 2016; Zhao, 2019) provide demonstrations of how the transmission of internal tides across ocean basins can be measured using altimeter data (Figure 11d), while the empirical model of Zaron (2019) was found to perform particularly well in a review of available internal tide models (Carrere et al., 2021). Modelling of internal tides in the global ocean is now being undertaken with increasing reliability (e.g. Arbic et al., 2012; Li & von Storch, 2020; Jithin et al., 2020; Arbic, 2022), although the exact mechanisms by which the energy in the barotropic tide sustains mixing via conversion to internal tides remains uncertain (Vic et al., 2019).

The evolution of the barotropic tide since the last glacial maximum has been modelled for many years (e.g. Arbic et al., 2004, 2008; Egbert et al., 2004; Uehara et al., 2006; Griffiths & Peltier, 2008). During this time, the tide had important impacts on ice sheets and shelves including playing a role in Heinrich events (Arbic et al., 2004; Doake et al., 2002; Bindschadler et al., 2003). However, more progress has been made recently on understanding the tides of previous, and potentially future, geological epochs; this field is called 'deep-time tide modelling' (e.g. Davies et al., 2020; Green et al., 2017; Green et al., 2020; Daher et al., 2021). Such studies are important for a more complete understanding of topics such as species evolution (Balbus, 2014). The same modelling techniques are now being applied to the study of tides on other planets and even exoplanets (Green et al., 2019; Blackledge et al., 2020).

It is clear that climate affects the barotropic tides directly through changes in water depth and the shape of coastlines (Haigh et al., 2020). Similarly, it is clear that tides are important in maintaining our present climate, given that they (together with winds) play a major role in mixing the ocean and thereby maintaining the global meridional overturning circulation (Munk, 1966; Munk & Wunsch, 1998; Johnson et al., 2019). In addition, there are many examples of variation in the tides being reflected in the regional climate of shelf seas where they are an important factor in influencing variations in various coastal ocean processes. For example, the spring-neap oscillations in tides result in a variation of the mixing and in the position of the shelf sea fronts that have great economic importance to the fishing industry on the NW European continental shelf (Sharples & Simpson, 2012). Similarly, spring-neap variations in tides have been found to affect mixing and circulation in other shelf areas such as around Indonesia (Ray et al., 2005; Ray & Susanto, 2016) where resulting changes in sea surface temperature (SST) are strong enough to induce a local atmospheric 'tide' (Ray & Susanto, 2019). On much longer timescales and on a global rather than regional shelf sea basis, modelling has suggested that tidal dissipation during the Last Glacial Maximum would have been much larger than today, impacting on the overturning circulation and therefore climate (Wilmes & Green, 2014; Wilmes et al., 2019). Munk et al. (2002) and Munk and Bills (2007) may also be consulted for discussion of the case for similar processes occurring on millennial timescales.

Therefore, it is somewhat surprising that there are not plausible (if weak) examples of tides affecting present-day global climate variability through modulations in tidally-induced vertical mixing. Such connections on decadal timescales were investigated by Ray (2007). For example, the amount of mixing in the global ocean could be modulated by the 18.6 year variation in tidal currents, and especially diurnal tidal currents. Such mixing, with obvious connection to the overturning circulation, would manifest itself as a variation in SST on nodal timescales. People have looked for such connections before. Yasuda et al. (2006) studied SST data from the North Pacific, where the diurnal currents are large so there is a better chance of observing nodal variations there than at other locations. Their findings were generally supportive of such a connection, see also later related ocean modelling studies discussed in Osafune et al. (2020). Ray (2007) extended the earlier analyses of SST time series from the Pacific and Atlantic coasts of Canada by Loder and Garrett (1978) which had pointed to possible evidence for this process, but with the conclusion that much longer time series are needed for a more convincing case. Therefore, evidence for any association of the tides and the ocean circulation on nodal and longer timescales remains for further research (e.g. see the recent modelling of Joshi et al., 2022).

One intriguing aspect of recent tidal research concerns the reasons for the apparent changes in tides in the last few decades along many coastlines, with these changes being much larger than can be explained by secular changes in the tidal potential (Woodworth, 2010). To some extent these changes are related to changes in MSL (i.e. ocean depth) and so can be studied using barotropic modelling techniques similar to those described above. However, changes in depth cannot be the whole story. This topic has been reviewed in detail by Haigh et al. (2020). Other published papers on this topic include Talke and Jay (2020) and Jänicke et al. (2021), while an interesting paper by Tyler (2021) provides support for recent tidal changes at Honolulu with the use of magnetic field data. In most cases, the recently observed changes in the tide are not large enough to impact significantly on the discussion of changes in extreme coastal sea levels which remain dominated by changes in MSL and storm surges.

10. Tsunamis and Meteotsunamis

Until 2004, there was little coordination between international 'sea level' measurement programmes (meaning those interested in sea level changes due to the ocean circulation and climate change) and those operated for tsunami warning, although some individual tide gauges, especially in the Pacific, provided information for both purposes. That situation ended after the Sumatra tsunami of December 2004, followed a few years later by the Tōhoku (or Sendai) tsunami in March 2011. Since then, there has been much improved cooperation between programmes, and considerable investment in new tide gauges and related equipment in the Pacific and Indian Oceans, Caribbean and Mediterranean, together with the establishment of regional tsunami warning systems. There has also been scientific recognition that tsunami studies will benefit from a fuller appreciation of sea level change (e.g. Li et al., 2018).

From a tsunami perspective, tide gauges are needed for two main purposes. The immediate objective, following an earthquake alert, is to provide confirmation to a warning centre that a tsunami exists, together with an estimate of its magnitude, so that the centre can pass reliable warnings on to more distant locations. Such confirmation takes place in combination with information from deep-sea tsunameters (bottom pressure tides gauges connected to surface buoys for data transmission) and other instrumentation (Pugh & Woodworth, 2014, Chapter 8). A second objective is to obtain sufficient data from a wide area on the characteristics of tsunami propagation in order to improve existing tsunami modelling (e.g. Titov et al., 2005). This introduces a requirement to monitor sea level at intervals of about 1 minute, most tsunami periods spanning a range of minutes to under an hour. By contrast, up until 2004 sea level programmes, with their primary interests in tides, surges, ocean circulation and changes in MSL, were largely content with their tide gauge having 6- or 15-minute or hourly sampling.

However, the higher rates now possible using modern sensors have provided one of the overall improvements that satisfies both communities. For example, the improved sampling enables higher-frequency processes such as seiches to be included in sea level studies, with a particular benefit to investigations of sea level extremes, as mentioned above. A second major development has come from improvements in data transmission on a broad front (increase in transmission slots for public geostationary meteorological satellites; use of other satellites such as provided by Inmarsat; and general improvement in land-based broad band transmission access, see IOC, 2016). There are mutual benefits in the real-time data reporting now possible, by enabling any faults to be identified and fixed much faster. As a consequence, the GLOSS programme of IOC now recognises fully the complementarity with tsunami applications (IOC, 2012). In addition, the IOC operates a Sea Level Station Monitoring Facility at the Flanders Marine Institute (http://www.ioc-sealevelmonitoring.org), by means of which the status of each gauge can be monitored continuously for the benefit of both communities. The value of this facility was demonstrated following the recent Tonga eruption and its associated tsunami and the response of the ocean to the accompanying air pressure wave (Carvajal et al., 2022; Amores et al, 2022). Meanwhile, there is an aspiration for eventual reporting at 1 Hz or similar, to aid post-event analysis of tsunami wave forms, and for the routine measurement of coastal waves.

Meteotsunamis are long waves in the ocean resulting from meteorological disturbances such as cyclones, frontal squalls, air pressure jumps, thunderstorms etc. The waves propagate until they reach a coastline, where they are amplified by shoaling and local resonance, often resulting in coastal flooding. Sea levels can rise by several decimetres or even metres, but considerable damage can result, even where the sea level signal is small, due to the associated oscillations in currents. Meteotsunamis are particularly well-known phenomena in locations such as the Mediterranean or Japan where there have been very damaging historical events (Monserrat et al., 2006; Jansa et al., 2007; Vilibić, and Šepić, 2009). However, it is becoming clearer that they can occur just about anywhere; Pugh and Woodworth (2014) provides a partial list of where meteotsunamis have been observed around the world. The reason that they form a relatively under-researched branch of sea level science is partly due to the same lack of high-frequency tide gauge data referred to above for tsunamis, a situation now being much improved, and also in this case to a lack of high-frequency meteorological data in order to understand their forcings better. Individual events, however, are now being modelled much more reliably than previously (e.g. Vilibić et al., 2016; Williams et al., 2019).

11. Some Challenges

There are many challenges which remain if one is to make progress on the topics discussed above. Lists of requirements for sea level research can be found in several recent publications (e.g. Horton et al., 2018), including those following major sea level conferences (e.g. Church et al., 2010; Stammer et al., 2018; Woodworth et al., 2019). Only a few of these requirements that are particularly important to me are mentioned here:

(1) The case for both space-based and in situ observing systems in general has been made for many years (e.g. Wilson et al., 2010) and needs to be updated at regular intervals (e.g. Le Traon et al., 2015; Marcos et al., 2019) in order to maintain the justification of the cases for funding.

(2) In particular, the global tide gauge network (GLOSS) must be completed to a common high technical standard, with all sites providing data every minute or faster in near-real time, enabling them to also function as tsunami gauges (IOC, 2012). All GLOSS gauges should be equipped with GNSS receivers for the geodetic reasons given in Section 6. Gauges remain of particular importance in highly populated coastal areas and for monitoring ocean circulation in high latitudes not accessible to altimetry (e.g. Hughes et al., 2003). The latter can be complemented by judiciously-located BPRs.

(3) And information from tide gauges needs to be combined in the most effective ways with data from a new generation of altimeter missions capable of monitoring sea levels closer to the coast (Hughes et al., 2019; Ponte et al., 2019; Woodworth et al., 2019). Such missions are described by Vignudelli et al. (2019a,b), Morrow et al. (2019) and IAT (2021).

(4) Expanded campaigns of data archaeology (or data rescue) are required in order to provide historical sea level information from the 18th-20th centuries that are not so far in databanks, enabling the best possible assessment of 'baseline' sea level change before the industrial period (Talke & Jay, 2013; Bradshaw et al., 2015; Hogarth et al., 2021). Such historical instrumental (i.e. tide gauge) information will be complemented by the 'proxy' sea level data from salt marshes. In addition, the quality of historical records needs to be re-assessed using modern data analysis techniques leading to improved information on trends and variability (Hogarth et al., 2020).

(5) Where long continuous tide gauge records are not available, but the benchmarks of earlier (e.g. 19th century) tide gauge installations still survive, then new temporary campaigns of tide gauge measurements can provide useful information on long-term sea level change (e.g. Hunter et al., 2003; Woodworth et al., 2010).

(6) From a modelling perspective, global numerical ocean models of 1/12° resolution or higher, optimised for sea level studies need to be developed (e.g. free of problems to do with mass conversation and representation of heat and freshwater fluxes and without a requirement for drifts to be constrained by climatology, and capable of representing both extreme sea level and long-term MSL variability). Such models could then be used to provide essential quality control to tide gauge and altimeter data, as well providing insight into the many deep ocean and coastal processes responsible for sea level variability.

(7) From a longer-term modelling perspective, more reliable projections are required of future regional (as well as global) MSL change which, together with modelling of tide and surge, will enable the changes in future extreme sea levels to be assessed more reliably (e.g. Vousdoukas et al., 2017, 2018). Within surge modelling, wave setup and run-up need to be considered to a greater extent alongside the much-studied wind setup (Dean & Walton, 2009; Dodet et al., 2019; Fieldler et al., 2020) as does the contribution of river runoff to sea level records (e.g. Piecuch et al., 2018).

(8) Much remains to be learned about the ocean tide and its variability over seasonal and longer timescales including the tidal changes which might occur due to a rise in sea level (Haigh et al., 2020). More information needs to be acquired globally on internal tides and short-spatial scale variations of the tide which occur near the coast. New missions should enable considerable progress to be made in these areas. In addition, more esoteric aspects such as modelling of third-degree tides offer interesting research topics for tidalists (Woodworth, 2019; Ray, 2020).

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Figure Captions

1. A diagram of a float and stilling well tide gauge by Athanasius Kircher (Kircher, 1665). Image courtesy of the Herzog August Bibliothek, Wolfenbüttel, Germany.

2. Locations of tide gauge stations with sea level information in the GESLA-3 dataset (Haigh et al., 2022). There are 5119 records containing 91021 station-years of information.

3. Satellite altimetry missions: past and future. A fill pattern with dots indicates a mission with an altimeter operating in SAR mode. Adapted from Vignudelli et al. (2019b).

4. 'Reconstructions' of global sea level change from Church and White (2011), Jevrejeva et al. (2006) updated, Ray and Douglas (2011), Hay et al. (2015), Dangendorf et al. (2017, 2019) and Frederikse et al. (2020). Each record is shown with an offset for presentation purposes. Different authors express uncertainties in different ways so an attempt has been made here to standardise on 68% confidence level uncertainties (1σ). The errors for Hay et al. (2015) and Dangendorf et al. (2019) are apparently larger due to the KS technique taking the full covariance structure of the data into account.

5. (a) Observed changes in the global-average height of MSL, and its estimated barystatic and thermosteric contributions (i.e. the contributions resulting from changes in ocean mass and from changes in density related to ocean temperature change, respectively) and their sum. (b) The barystatic contribution and its individual components. The terrestrial water storage (TWS) term is the sum of groundwater depletion, water impoundment in artificial reservoirs and the natural TWS term. (c) 30-year-average rates of observed global MSL change and the different contributing processes. (d) 30-year-average rates of global MSL change due to the barystatic contribution and its individual components. The shaded regions denote 90% confidence intervals. The values in (a) and (b) are relative to the 2002–2018 mean. From Frederikse et al. (2020).

6(a). Relative 'proxy' sea level reconstructions for the North American Atlantic coast and Iceland together with nearby tide gauge records. (a,c) Temporal sea level evolution since 900 CE for the proxy records. Yellow crosses are proxy sea level estimates from cores with 2 sigma vertical and age uncertainties. The red crosses are basal sea level index points, immune to compaction, and show that records are not significantly affected by compaction. Also shown are tide gauge records from nearby stations (black lines) and a Gaussian Process fitted to the individual index points (blue) with 1 and 2 sigma uncertainties from 1000 Monte Carlo simulations as shading. (b,d) Map with locations of proxy (yellow stars) and tide gauge records (black dots). SS—Scotian Shelf, GoM—Gulf of Maine, MAB—Mid Atlantic Bight, CH—Cape Hatteras. From Gehrels et al. (2020).

(b). Sea level changes from proxy records along the North American Atlantic (blue) and Icelandic (red) coast. Shown are the nonlinear trends calculated by a Gaussian Process Regression including their 1 and 2 sigma uncertainties (dark and light blue/red bands, respectively) for the six salt marsh reconstructions corrected for site-specific GIA effects. The black dotted line marks a rate of 0 mm/year in each panel. From Gehrels et al. (2020).

7. Linear trends (mm/year) of annual 99th percentiles of a) total sea level, b) with median removed, and c) for non-tidal residuals with median removed. Linear trends of annual 99th percentiles of d) skew surges, e) with median removed, and f) with low-pass filter MSL removed. Tide gauge data from 1960 to present are used. Black dots indicate where the trends are not significant. From Marcos and Woodworth (2018).

8. An acoustic tide gauge at Burnie in northern Tasmania, Australia, and, to its right, a special pillar with a GNSS receiver on top. In this example, the tie between the tide gauge and GNSS receiver is very short. Photograph courtesy of Geoscience Australia.

9. Monthly mean sea levels (in metres) from Ny-Alesund (Svalbard, Norway), Friday Harbor (Washington, USA), Andenes (Norway), Tregde (Norway), Spring Bay (Tasmania, Australia) and Crescent City (California, USA). MSL values from the PSMSL are shown in blue while red values are from GNSS reflectometry data offset by 0.2 m in each case for presentation. Figure courtesy of Simon Williams.

10. (a) The CNES-CLS18 model of the mean dynamic topography (MDT) of the ocean, determined largely by computing the difference between the mean sea surface (MSS) and the geoid, supplemented by information from various ocean data sets. (b) The geostrophic surface currents corresponding to the MDT in (a). From Mulet et al. (2021).

11. (a,b,c) Co-tidal charts of the M2 ocean tide from the FES2014 model (Lyard et al., 2021). The white lines indicate Greenwich phase lag every 30°, a lag of zero degrees being shown by the bold line. The colours show amplitudes in metres. Images courtesy of Florent Lyard, GRGS, Toulouse. (d) Amplitude of the mode-1 M2 internal tide from multi-satellite altimetry. The 3000 m isobath contours are in black while regions with high mesoscale variability are masked in light blue. From Zhao et al. (2016).

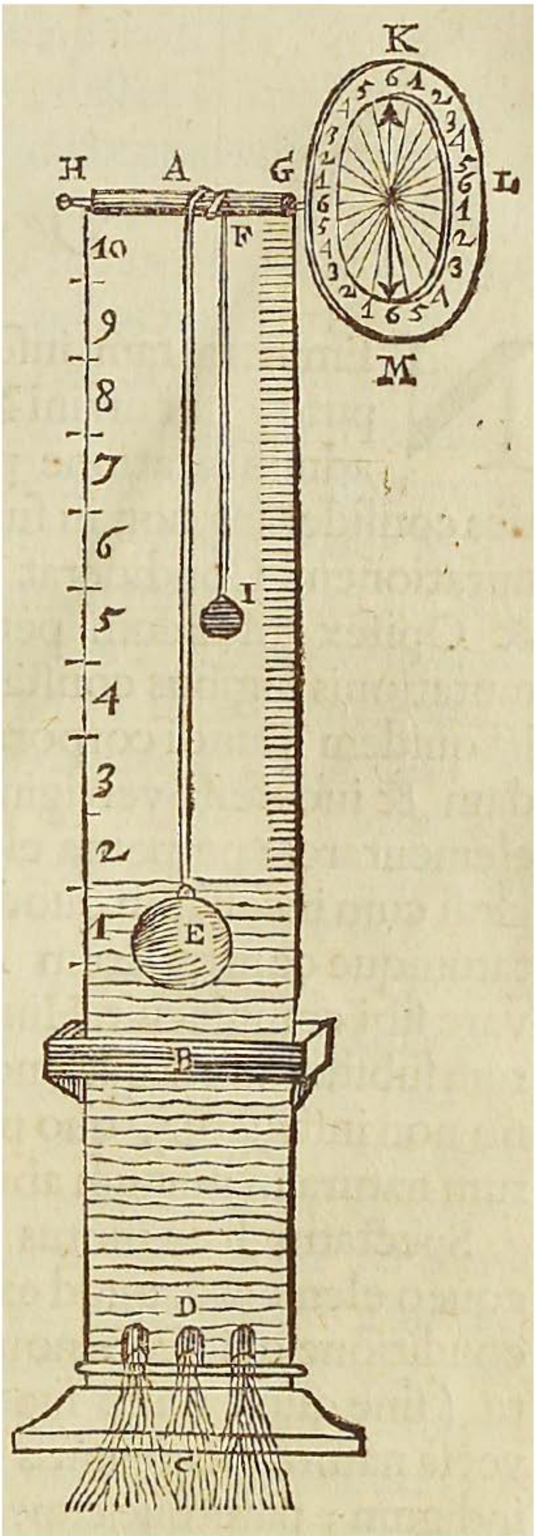


Figure 1

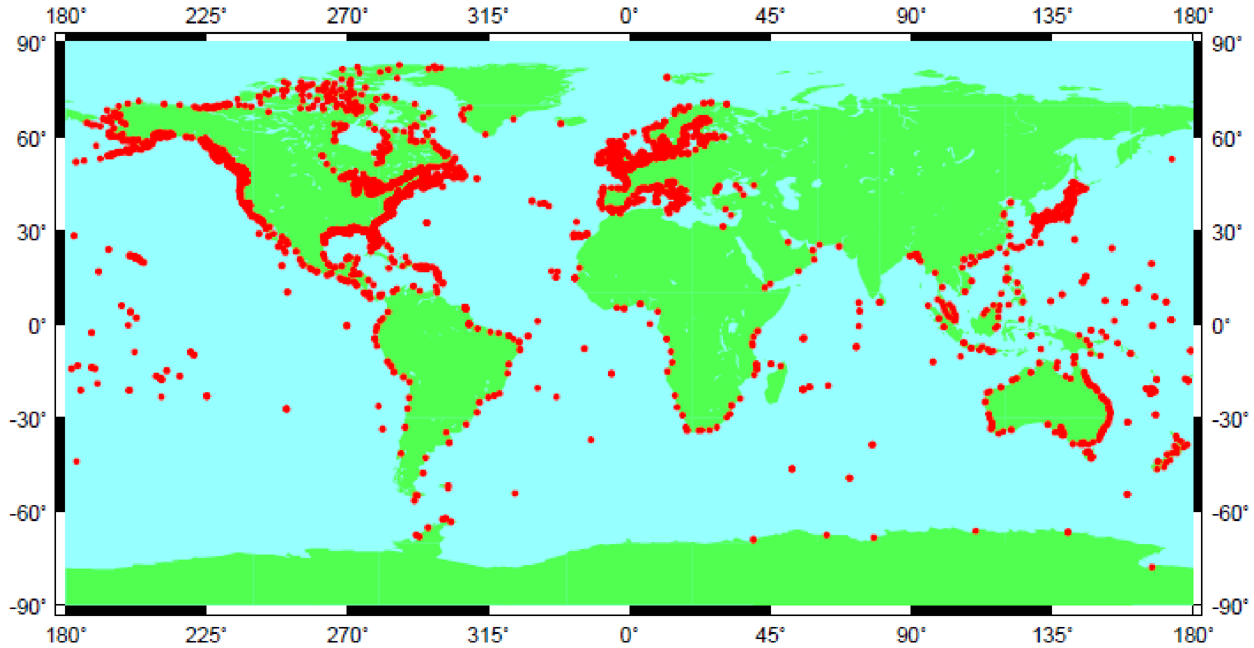


Figure 2

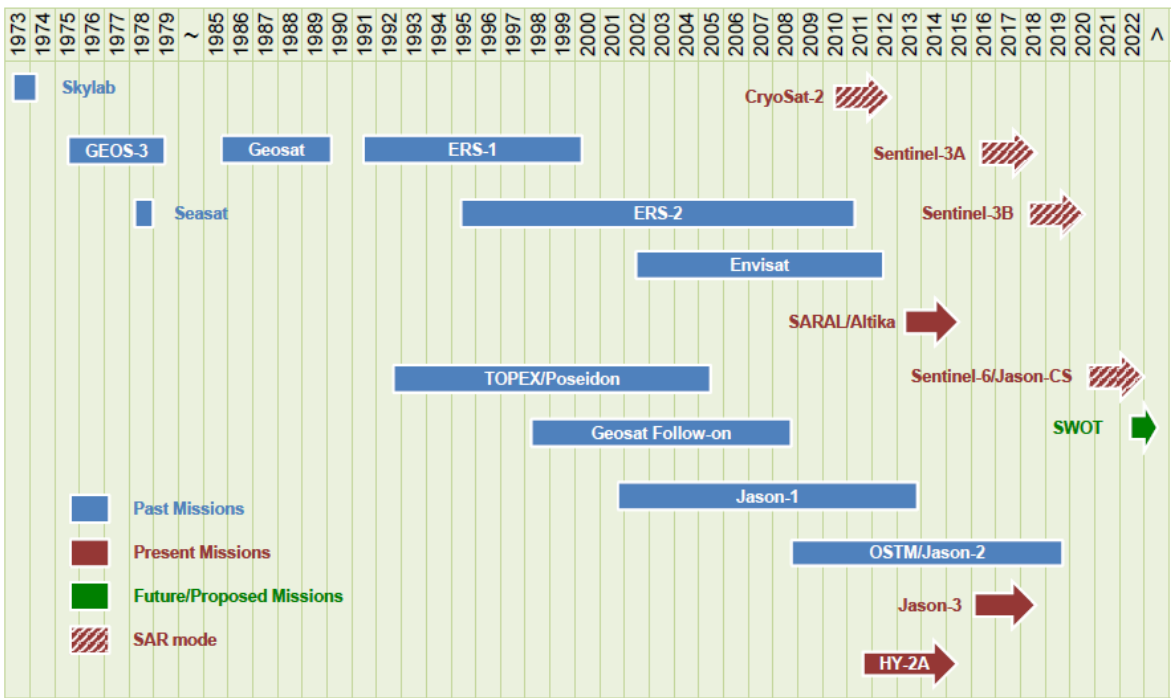


Figure 3

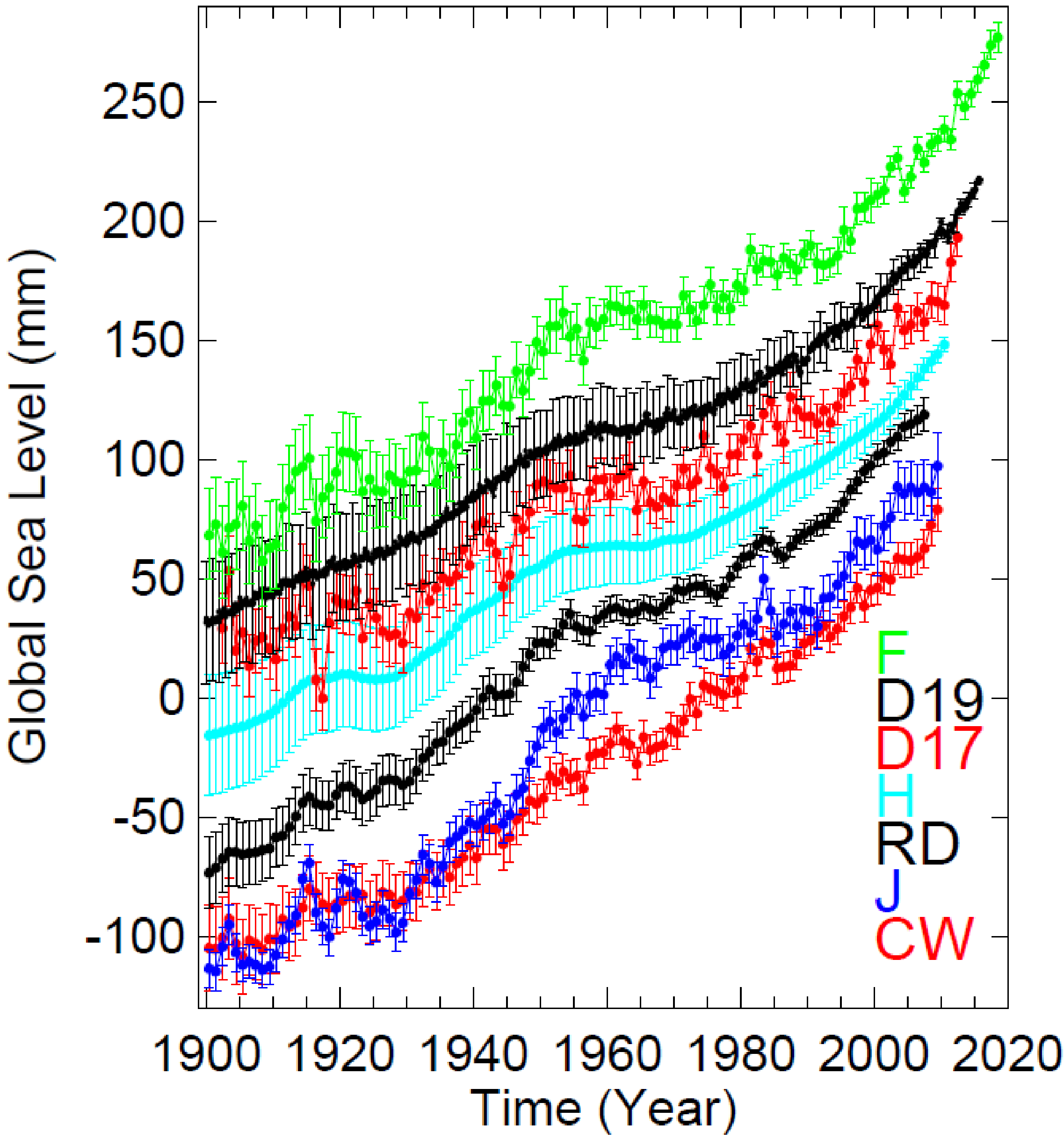

Global Sea Level (mm)
Time (Year)
250
200
150
100
50
0
-50
-100
1900
1920
1940
1960
1980
2000
2020
F
D19
D17
H
RD
J
CW

Figure 4

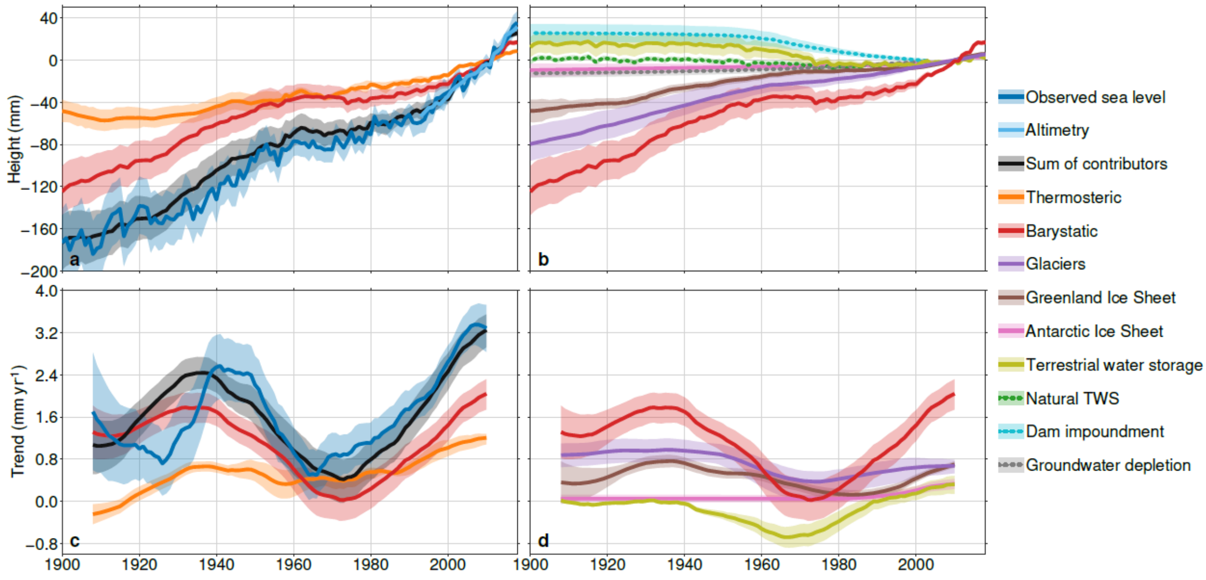
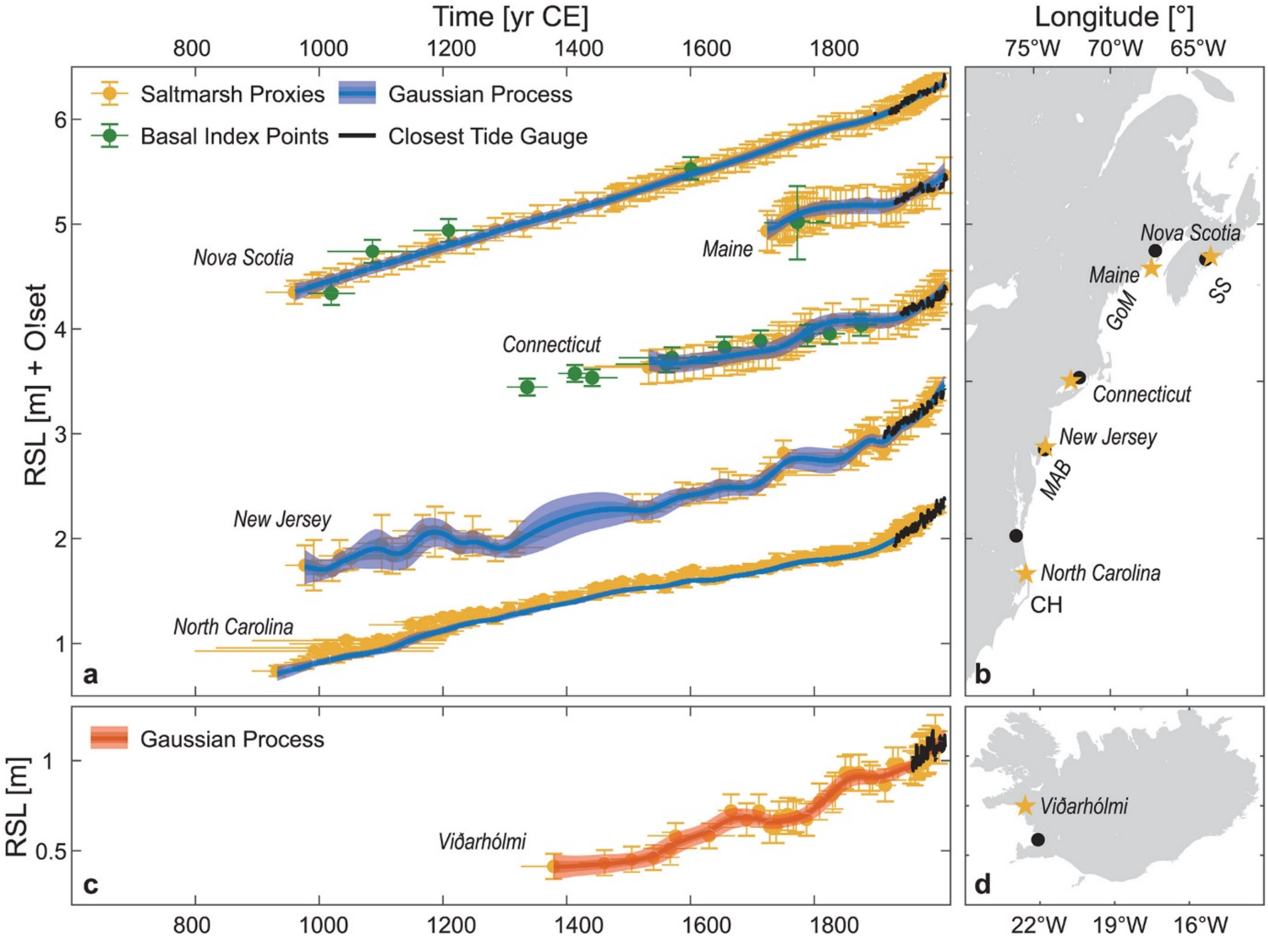
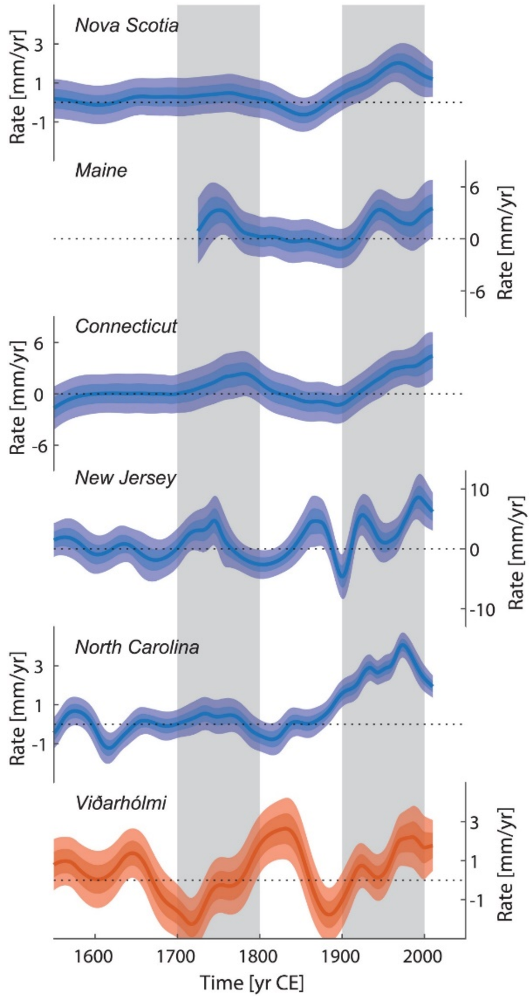

Height (mm)
Trend (mm yr⁻¹)
a
b
c
d
Observed sea level
Altimetry
Sum of contributors
Thermosteric
Barystatic
Glaciers
Greenland Ice Sheet
Antarctic Ice Sheet
Terrestrial water storage
Natural TWS
Dam impoundment
Groundwater depletion

Figure 5



Time [yr CE]
800
1000
1200
1400
1600
1800
Saltmarsh Proxies
Gaussian Process
Basal Index Points
Closest Tide Gauge
RSL [m] + O!set
Nova Scotia
Maine
Connecticut
New Jersey
North Carolina
a
Longitude [°]
75°W
70°W
65°W
GoM
SS
MAB
CH
b
Gaussian Process
RSL [m]
Viðarhólmi
c
d
22°W
19°W
16°W

Figure 6(a)



Nova Scotia
Maine
Connecticut
New Jersey
North Carolina
Viðarhólmi
Rate [mm/yr]
Time [yr CE]
1600
1700
1800
1900
2000

Figure 6(b)

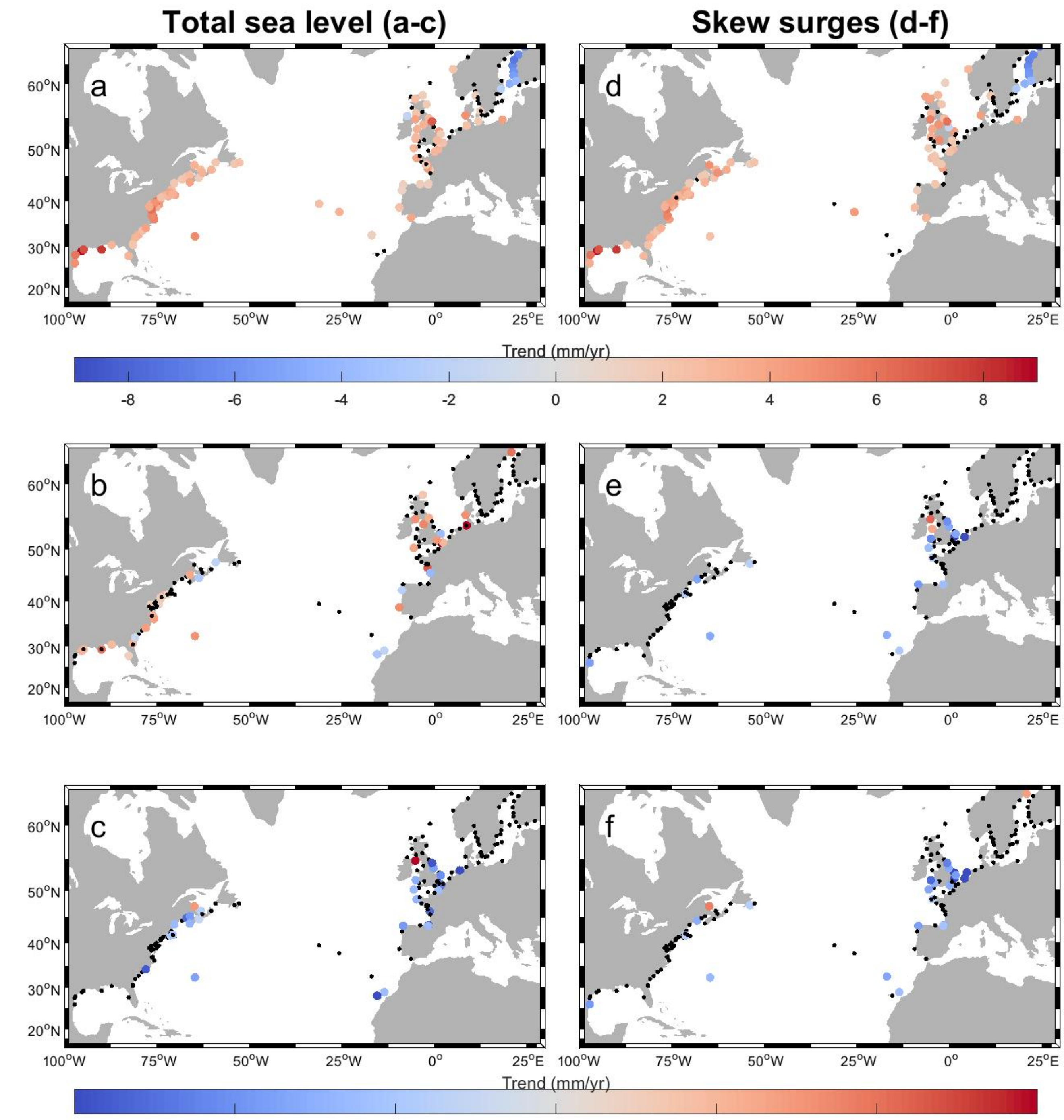

Total sea level (a-c)
Skew surges (d-f)
a
b
c
d
e
f
60°N
50°N
40°N
30°N
20°N
100°W
75°W
50°W
25°W
0°
25°E
Trend (mm/yr)
-8
-6
-4
-2
0
2
4
6
8
Trend (mm/yr)
-3
-2
-1
0
1
2
3

Figure 7

Figure 8

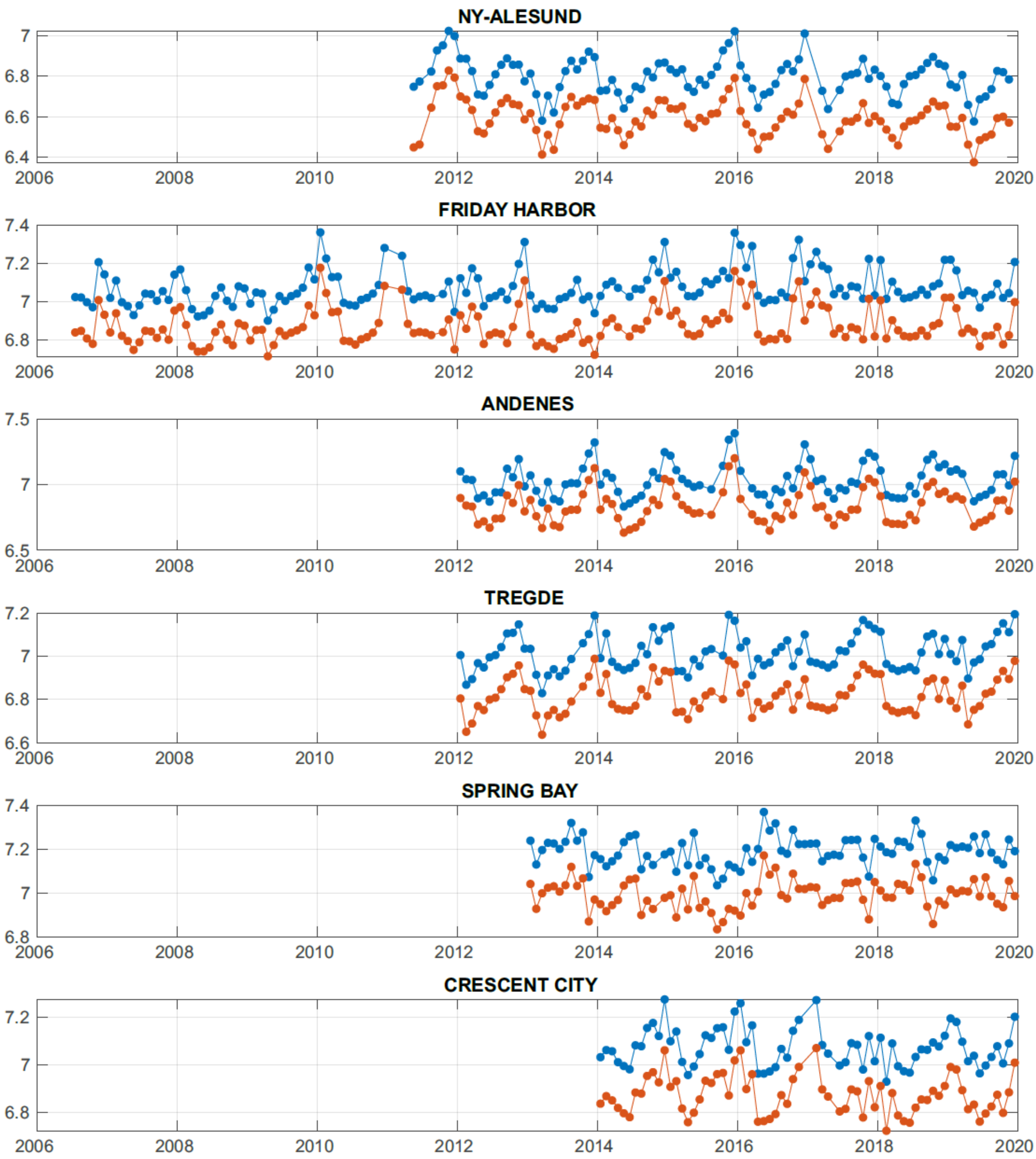
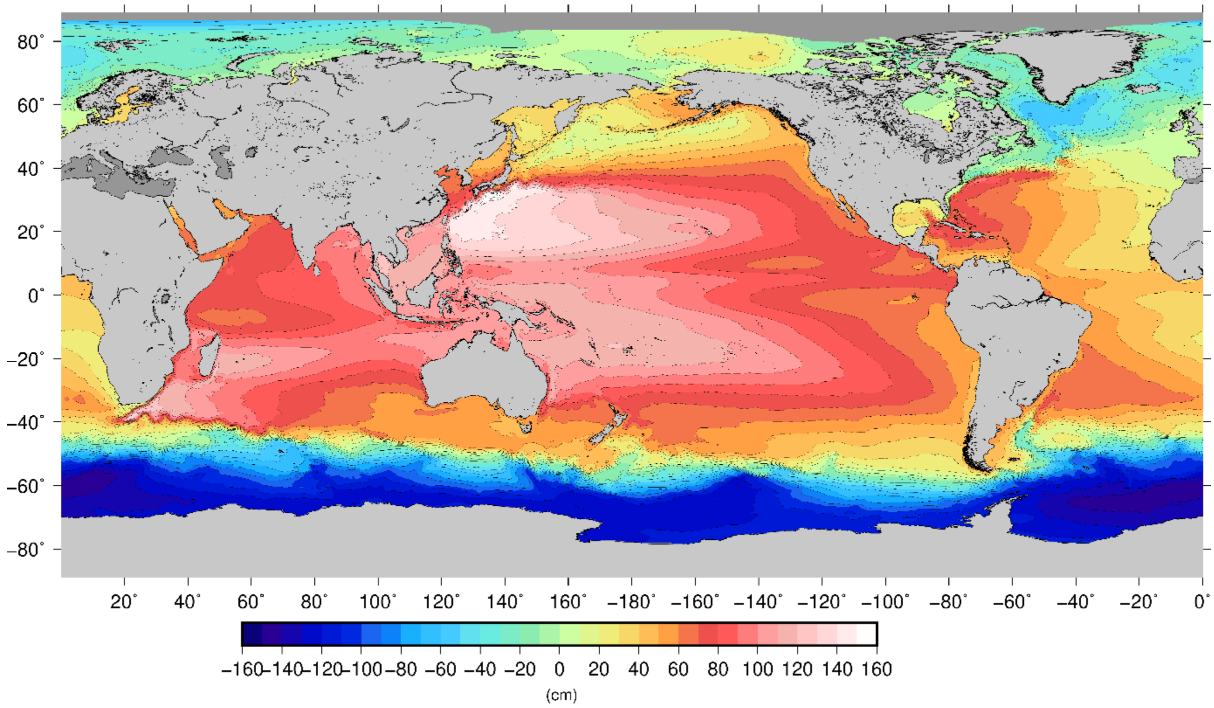
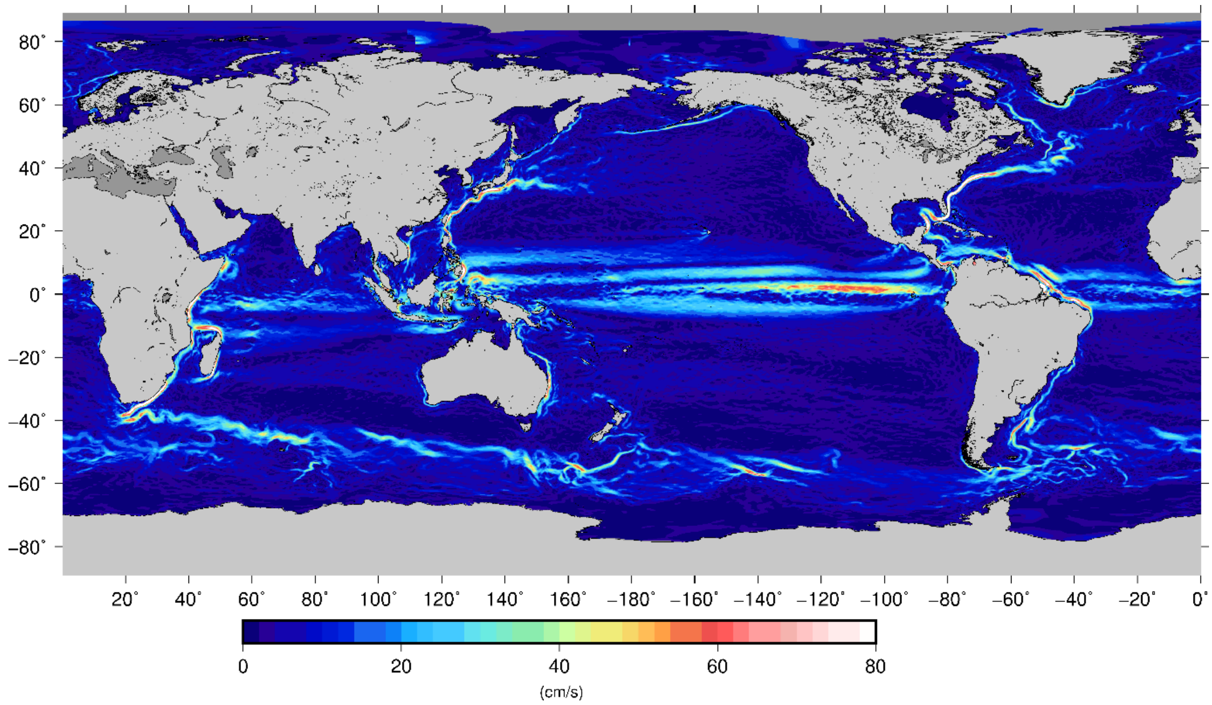

NY-ALESUND
7
6.8
6.6
6.4
FRIDAY HARBOR
7.4
7.2
7
6.8
ANDENES
7.5
7
6.5
TREGDE
7.2
7
6.8
6.6
SPRING BAY
7.4
7.2
7
6.8
CRESCENT CITY
7.2
7
6.8
2006
2008
2010
2012
2014
2016
2018
2020

Figure 9



80˚
60˚
40˚
20˚
0˚
−20˚
−40˚
−60˚
−80˚
20˚ 40˚ 60˚ 80˚ 100˚ 120˚ 140˚ 160˚ −180˚ −160˚ −140˚ −120˚ −100˚ −80˚ −60˚ −40˚ −20˚ 0˚
−160 −140 −120 −100 −80 −60 −40 −20 0 20 40 60 80 100 120 140 160
(cm)

(a)



80˚
60˚
40˚
20˚
0˚
−20˚
−40˚
−60˚
−80˚
20˚ 40˚ 60˚ 80˚ 100˚ 120˚ 140˚ 160˚ −180˚ −160˚ −140˚ −120˚ −100˚ −80˚ −60˚ −40˚ −20˚ 0˚
0 20 40 60 80
(cm/s)

(b)

Figure 10

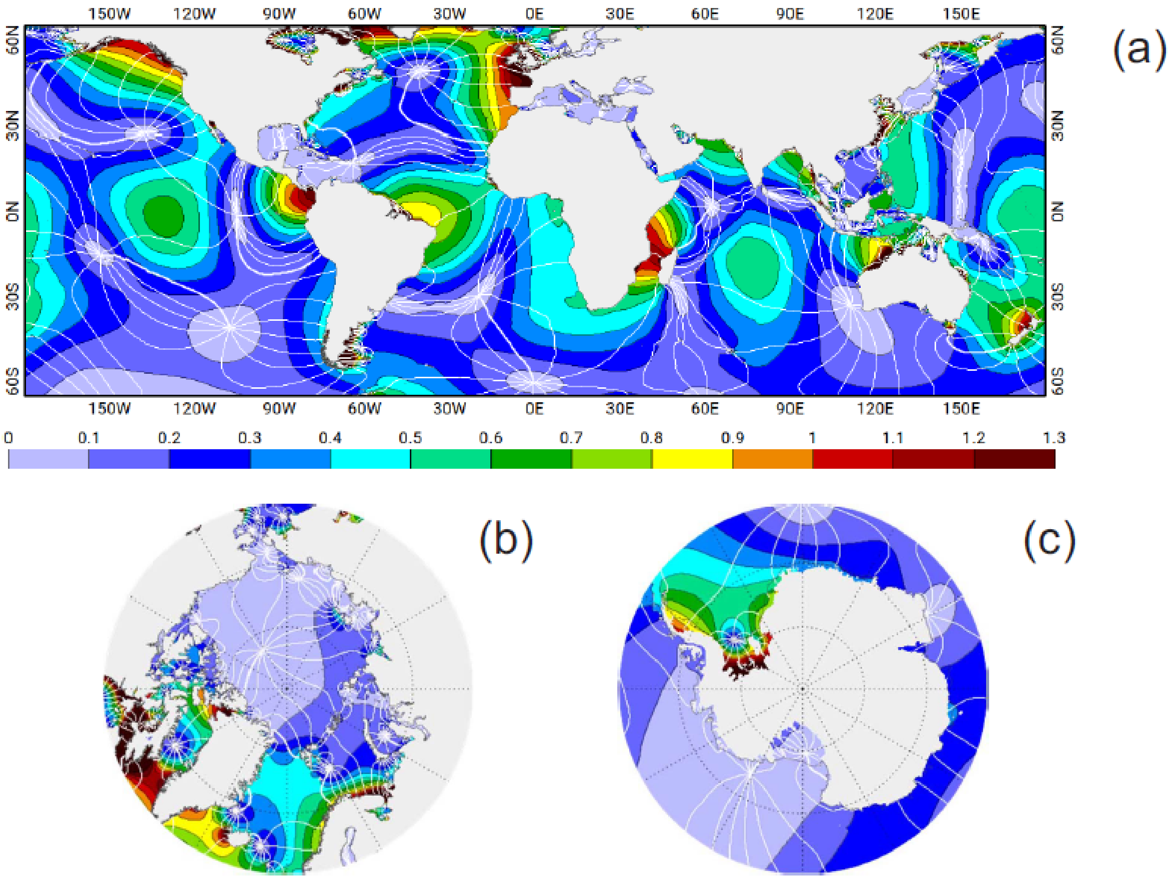


Figure 11(a,b,c)

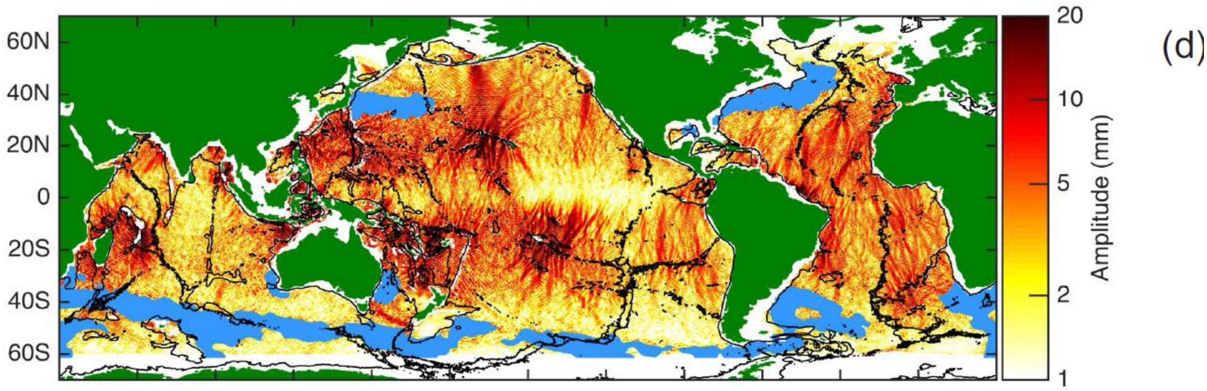


Figure 11(d)

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