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Contributions to 21st century projections of extreme sea-level change around the UK

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Abstract

We provide a synthesis of results of a recent government-funded initiative to make projections of 21st century change in extreme sea levels around the coast of the United Kingdom. We compare four factors that influence future coastal flood risk: (i) time-mean sea-level (MSL) rise; (ii) changes in storm surge activity; (iii) changes in the offshore wave climate; (iv) changes in tidal amplitude arising from the increase in MSL. Our projections are dominated by the effects of MSL rise, which is typically more than five times larger than any of the other contributions. MSL is projected to rise by about 53 to 115 centimetres at the mouth of the Thames and 30 to 90 centimetres at Edinburgh (5th to 95th percentiles at 2100 relative to 1981–2000 average). Surge model projections disagree on the sign of future changes. Typical simulated changes are around $+/-7$ centimetres. Because of the disagreement, our best estimate is of no change from this contribution, although we cannot rule out changes of either sign. Wave model projections suggest a decrease in significant wave height of the order of 7 centimetres over the 21st century. However, the limited sample size and uncertainty in projections of changes in atmospheric circulation means that we cannot be confident about the sign of future changes in wave climate. MSL rise may induce changes in tidal amplitude of more than 15 centimetres over the 21st century for the Bristol Channel. However, models disagree on the sign of change there. Elsewhere, our projected tidal amplitude changes are mostly less than 7 centimetres. Whilst changes in MSL dominate, we have shown the potential for all processes considered here to make non-negligible contributions over the 21st century.

1. Introduction

For the UK, major coastal flood events are typically associated with wintertime storm surges that coincide with a high tide and high wave conditions. For example, the storm surge of 6th December 2013, which affected sites all along the east coast of the UK and as far afield as Ullapool (in the north west of Scotland) and Portsmouth (on the south coast of England) caused seawater inundation of more than two thousand homes and businesses and three thousand hectares of farmland. Along the east coast, over ten thousand people were evacuated, and flood defences were breached in around fifty locations (Surgewatch database, Haigh *et al* (2015)). It is estimated that around £150 billion of assets are at risk from coastal flooding in the UK (Howard *et al* 2010), and estimated annual damages to the UK from coastal flooding are of the order of £500 million (Edwards 2017). Consequently, there is concern about the effects of climate change on the future frequency and intensity of extreme coastal sea-level events.

In general, climate change can give rise to increased coastal flood risk in two ways. The first is through an increase in time-mean sea level (often referred to simply as 'mean sea level', MSL), i.e. the baseline average water level on which tides, surges, waves and other sea-level extremes are superimposed. Menéndez and Woodworth

(2010) showed that this component has dominated the changes in observed extreme water levels in the historical record.

Observations show that MSL around the UK has risen by approximately 1.4 mm/year from the start of the 20th century, when corrected for land movement (Kendon *et al* 2018, Woodworth *et al* 2009). This is comparable to the global mean rate of 1.7 mm/year (1.5–1.9; 90% confidence interval) for the period 1901 to 2010 given by Church *et al* (2013). MSL around the UK is projected to continue to rise until at least the end of the 21st Century under all RCP climate change scenarios (Palmer *et al* 2018, see also section 3). The second way that climate change can influence future coastal flood risk is through changes in the drivers of sea-level extremes, particularly storm surge events and the wave climate. In this case there can be uncertainty in both the magnitude and sign of the change—with the potential for decreases as well as increases in extreme sea levels, relative to the mean. While MSL change has a direct influence on extreme sea-level events through simply raising the baseline water level, recent research has also highlighted a secondary, but important, effect on tidal amplitudes. The tide travels as a shallow water wave with phase speed determined by the water depth. This means that, since the frequency is fixed, the wavelength is controlled by the water depth. The wavelength in turn controls the locations of the amphidromic points, and changes in these locations affect the tidal range at the coast, which is of most practical relevance for flooding. The presence of near-resonant estuaries can introduce further sensitivities (Pickering 2017, Pelling *et al* 2013, Haigh *et al* 2019), with changes in MSL having the potential to move the tides in such estuaries either further from or closer to resonance depending on location. Thus, the four potential sources of changing coastal flood risk for the UK that we explore here are: (i) MSL rise, (ii) changes in the statistics of storm surge, (iii) changes in the statistics of offshore waves, and (iv) changes in the amplitude of the tide.

Here we consider the relative sizes of these factors, using simulations developed as part of a recent UK government-funded initiative, the United Kingdom Climate Projections 2018 (hereafter referred to as UKCP18) and documented in the UKCP18 Marine Report (UKCP18-Marine, Palmer *et al* (2018)). Following the approach of previous studies (e.g. Lowe *et al* 2009, Vousdoukas *et al* 2018), we treat the factors independently (except in the sense that the changes in tidal amplitude are driven by MSL change). For a discussion of this approach see Vousdoukas *et al* (2018) and references therein. We do not comprehensively examine the strength of the factor interactions, although we note the recent work of Arns *et al* (2017) indicating that such interactions may be important at some locations. UKCP18 considers many physical aspects of climate change for the UK, and is the latest in a series of similar initiatives (Hulme *et al* 2002, Lowe *et al* 2009). UKCP18 and associated publications such as Palmer *et al* (2018) were comprehensively reviewed by an independent panel of experts.

Our simulations of factors (i), (ii) and (iii) above are self-consistent in the sense that they are all predicated on data from climate model simulations produced for the Coupled Model Intercomparison Project phase 5 (CMIP5, Taylor *et al* 2012) under representative concentration pathway 8.5 (RCP8.5, van Vuuren *et al* 2011). We base our assessment on RCP8.5 because it represents the strongest radiative forcing of the four representative concentration pathways adopted by the Intergovernmental Panel on Climate Change (IPCC) for its Fifth Assessment Report (AR5, IPCC 2013), and thus provides the largest signal-to-noise ratio in terms of forced climate response versus unforced climate variability. Compared to IPCC AR5 (Church *et al*, 2013), the UKCP18 MSL projections use an updated estimate for the contribution from Antarctica and a more traceable and comprehensive treatment of the regional uncertainties. Both surge and wave simulations make use of CMIP5-downscaled simulations from Euro-CORDEX. This is a step towards addressing the structural uncertainty, in contrast with some previous studies, for example Lowe *et al* (2009), who used a perturbed parameter ensemble of a single climate model, or Sterl *et al* (2009), who used a perturbed initial condition ensemble of a single climate model.

In section 2 we summarize the methods used to produce the simulations of each source of change. Results are presented and discussed in section 3

2. Methods

As discussed above, all of our assessments are based on RCP8.5, and we focus on the century-scale change. We note that shorter-term changes in the extremes may be dominated by natural variability, for example the 18.6-year nodal tidal cycle, or the North Atlantic Oscillation. We give an indication of the size of inter-annual variability in the storm surge extremes in section 3.

2.1. Time-mean sea level (MSL)

The projections of MSL change presented here are taken directly from UKCP18-Marine. In this section, we present a brief synopsis of the UKCP18-Marine methods and refer the reader to Palmer *et al* (2018) for the full details.

The UKCP18 projections are rooted in the sea-level methods of the IPCC AR5 (Church *et al* 2013), based on CMIP5 climate model simulations (Taylor *et al* 2012) under the RCP climate change scenarios (Meinshausen *et al* 2011). Recent assessment of CMIP5 model simulations demonstrates their ability to reproduce the main components of global and regional rise over the historical record (Slangen *et al* 2017, Meyssignac *et al* 2017) and promotes confidence in their ability to provide useful projections of future change. The main innovations to the UKCP18 sea-level projections relative to IPCC AR5 are: (i) inclusion of the scenario-dependent estimates of Antarctic dynamic ice input from Levermann *et al* 2014; (ii) use of a regression approach to estimate the regional oceanographic changes that better isolates the climate change signal from CMIP5 simulations (e.g. Perrette *et al* 2013, Bilbao *et al* 2015); (iii) more comprehensive treatment of regional uncertainties and direct traceability to the CMIP5-based projections of global sea level. The inclusion of the Levermann *et al* (2014) estimates results in a change in the 5th to 95th percentile ranges of global sea-level rise under RCP8.5 from 0.53–0.98 m to 0.56–1.12 m compared to AR5 (after adjusting values to the 1981–2000 baseline used in UKCP18).

Global sea-level projections, which include estimates of global thermal expansion and future mass addition from glaciers, ice sheets and changes in land water storage, form the basis of the regional sea-level projections for the UK mainland. The regional projections take account of the spatial patterns that arise from the different ice mass and land water changes as a result of the response of Earth's gravity field, rotation and vertical land motions (e.g. Tamisiea and Mitrovica 2011). Some representation of the uncertainty associated with these effects is included by the use of two sets of these 'mass fingerprints' (Slangen *et al* 2014, Spada and Stocchi 2007). The effects of local changes in ocean circulation and seawater density are accounted for using simulations from 21 CMIP5 models, which introduces substantial additional uncertainty at regional scales. Finally, estimates of the effects of glacial isostatic adjustment (GIA) on regional MSL, which are dominated by the effects of vertical land motion, come from a 15-member ensemble produced as part of the NERC BRITICE-CHRONO project (Bradley *et al* 2018). The uncertainties of the various contributions to regional sea-level change are combined using a 100,000 member Monte Carlo simulation, following the same approach as used for the global MSL projections presented in IPCC AR5. The 50th percentile of the resulting projections is used as our central estimate of the MSL change. Following IPCC AR5, we use the 5th and 95th percentiles of the Monte Carlo distribution to illustrate the uncertainty in the MSL projections.

2.2. Surge

Both wave and surge changes depend on atmospheric changes, and there is consistency in our approach to these two components: we show results from models driven by atmospheric projections from CMIP5 in both cases. In both cases we also use atmospheric data from regional model simulations performed by the Swedish Meteorological and Hydrological Institute (SMHI) using their regional atmospheric model RCA4 as part of the Euro-CORDEX experiment (Jacob *et al* 2014). The consistency is incomplete because different CMIP5 models were selected for the two different strands of work. The selection was based on data availability and suitability, and is discussed in more detail by Palmer *et al* (2018).

The potential for change in the statistics of storm surge associated with projected changes in atmospheric storminess is assessed here using atmospheric projections from two CMIP5 models: HadGEM2-ES (Jones *et al* 2011) and GFDL-ESM2M (Dunne *et al* 2012). We choose here to present HadGEM2-ES out of the set of five RCA4-downscaled simulations used in UKCP18-Marine because, out of these five, HadGEM2-ES has the most negative 21st-century trend in extreme surge. GFDL-ESM2M, on the other hand, was selected as a CMIP5 model which might be expected to exhibit a large positive trend according to two metrics of storminess change (see section 3.1). This expectation is borne out by the results shown in section 3.

For HadGEM2-ES, we use surface winds and pressure downscaled through RCA4 as part of the Euro-CORDEX experiment, to drive the CS3 storm surge model (Flather 1994, 2000). CS3 is a depth-averaged two-dimensional model covering the NW European shelf from 12 degrees west to 13 degrees east and 48 to 63 degrees north, with a horizontal resolution of 1/6 degree longitude $\times$ 1/9 degree latitude. HadGEM2-ES produces a very credible simulation of the latitudinal variation of time-averaged storm density (count per unit area) for storms crossing north west Europe, with the three observed average density peaks being particularly well represented (Palmer *et al* 2018). When HadGEM2-ES is used to drive CS3 via the RCA4 regional atmospheric model, the inter-annual variability in the annual maximum surge is well simulated at most UK locations, giving confidence in the surge modelling system.

For GFDL-ESM2M, corresponding downscaled data was not available and so we chose to drive CS3 directly with surface winds and pressure from the global model simulation. For a further discussion of this choice and its implications see Palmer *et al* (2018). This choice does not have any important implications for the results shown here. The storm density is less well simulated by GFDL-ESM2M, with too little latitudinal variation, although the average values are quite realistic. The inter-annual variability in the annual maximum surge is also less well simulated when CS3 is driven by GFDL-ESM2M, but the spatial pattern of the variation in this metric appears similar to the pattern in the observations. Further details of model evaluation are given in Palmer *et al* (2018).

In contrast to the tidal simulations discussed in section 2.4, our surge simulations are based on present-day bathymetry. This approach has been used in many previous studies (for example Lowe *et al* 2009, Sterl *et al* 2009, Debernard and Røed 2008) and is supported by sensitivity tests shown by Palmer *et al* (2018), which indicate that the effect of MSL change on the skew surge is small (less than 2 cm for 50 cm of MSL rise at all of the sites considered in the case studies) compared to the effect on the tides. (Lowe *et al* 2001, Howard *et al* 2010) find similar insensitivity of skew surge to MLS increase. Skew surge (de Vries *et al* 1995) is now widely preferred as a metric of surge (e.g. Cannaby *et al* 2016, Howard *et al* 2010), owing to both its impact-relevance and its independence of tidal level (Williams *et al* 2016). To characterize the extremes, the concept of a return level (e.g. Coles 2001) is routinely used. For example, the ten-year return level is the level which we expect to be exceeded on average once every ten years. Here we characterize the change in the extremes of skew surge using the change in the one-year return level. Palmer *et al* (2018) show that there is insufficient evidence within the RCA4-downscaled simulations for considering independent change in other return levels, so that the change in the one-year return level applies to all return levels for those simulations. In contrast, there *is* sufficient evidence for such independent change in the GFDL-ESM2M simulation. For this simulation, some results for the 200-year return level are shown in UKCP18-Marine. For consistency, we consider only results for the one-year return level here.

To quantify the extremes we use a statistical model based on the five largest independent skew surge events each year (Coles 2001). The approach, which is described in detail in Palmer *et al* (2018), identifies a linear rate of change (or 'trend') in the one-year return level of skew surge at each model grid point based on the five largest skew surges for each year of 93 years of simulation (2007–2100). Here we multiply this rate of change by 100 years to characterize the 21st-century change. This metric is labelled S_H and S_G in figures 2 and 3. S_H refers to the change exhibited by the HadGEM2-ES-RCA4 simulation, and S_G refers to the change exhibited by the GFDL-ESM2M simulation.

Although we consider different atmospheric driving models, we downscale them to surge with a single shelf model. There are reasons to anticipate that the sensitivity to a change of surge model would be small. First, Flowerdew *et al* (2010) showed that the uncertainty in an ensemble surge forecast was dominated by the atmospheric rather than the ocean conditions. Secondly, Jordà *et al* (2012), studying the contribution of atmospheric pressure and wind to the 21st-century sea-level variability in Southern Europe, find that uncertainties in sea-level results are mostly induced by the uncertainties in the atmospheric fields used to force the ocean model. Finally, and perhaps most importantly, we anticipate that any robust century-scale change in the statistics of local extreme surges will be a reflection of large-scale changes in atmospheric storminess, rather than a result of the model used to downscale the atmospheric storms to surges. Our finding of no clear signal of change is consistent with the overall disagreement between CMIP5 models on the projected changes in storminess over the region considered. This is discussed further in the UKCP18 Marine Report. A similar argument applies to the wave model used (see next section): any meaningful change must come from the driving atmosphere, not the wave model used to downscale it. Consistent with this, Morim *et al* (2019) find that uncertainty in projections of changing wave climate is dominated by climate model-driven uncertainty, rather than uncertainties in wave modelling. The wave model used here was included in the Morim *et al* (2019) study, where the sensitivity to wave model used is discussed in more detail.

2.3. Waves

We use results from regional wave model projections for the 21st century based on the CMIP5 model EC-EARTH (Hazeleger *et al* 2012). Surface winds from this model are used to drive the spectral wave model WaveWatch III version 3.14 (Tolman 2009) as part of the EU RISES-AM project. Improved regional detail around the UK is provided by a nested regional version of the same wave model. This in turn is driven by surface winds from EC-EARTH downscaled through RCA4 as part of the Euro-CORDEX experiment.

Bricheno and Wolf (2018) present a detailed statistical evaluation of the wave model performance, using fourteen UK tide gauges with between five and fifteen years of observations at each. They show that the model successfully simulates temporal and spatial variability in a full range of wave conditions from calm to stormy, for both sheltered and exposed sites, experiencing swells, windsea and bimodal conditions. Wave direction is well-simulated in the regional model. They find no consistent bias or drift in the model when driven with atmospheric reanalysis data. Biases are seen in the results when driven with climate model data, but the spatial pattern of these biases is not correlated with the spatial pattern in the signal of change, and the EC-EARTH simulation gives a realistic spatial pattern of extreme waves for the historical period. Further details and further results are reported in Bricheno and Wolf (2018) and Palmer *et al* (2018).

Although results shown in section 3 are based on this single model (EC-EARTH) taken from the CMIP5 'ensemble of opportunity', our discussion is also informed by results—shown by Palmer *et al* (2018)—of global wave model simulations forced by other CMIP5 models.

In an effort to identify the century-scale change and minimize the influence of short-period variability, we take means of SWH over two twenty-year periods of the simulation (1981–2000 and 2081–2100) and the difference between the two is taken to represent the 21st-century change.

The effect of MSL change on the waves is not modelled in these experiments (see section 3).

2.4. Tide

A growing body of evidence shows that tidal properties have changed, and continue to evolve, due to non-astronomical factors including sea-level change, and that changes in tidal properties are likely to occur over the next centuries (Pickering *et al* 2012, Haigh *et al* 2019). To make a projection of the contribution to extreme sea-level change from this source, the CS3 model is deployed again without atmospheric forcing but with an adjusted bathymetry to provide insight into potential tidal changes. The exclusion of atmospheric forcing is supported by sensitivity tests shown by Palmer *et al* (2018, their 'case studies'). They simulate three very different historical surge events (affecting the UK east coast, the UK south and west coasts, and the Scottish coast, respectively) under present-day mean sea level and a range of increased mean sea levels, including a 50 cm increase. In all three cases, at all affected sites, they find that the effect of MSL change on the simulated skew surge is small (less than 2 cm under a 50 cm MSL increase). Since the peak still water level is the sum of the peak astronomical tide and the skew surge, this shows that the addition of atmospheric forcing has only a small impact on the *change* in the tidal range, which is mostly driven by the change in the MSL, in all three cases at all of the sites considered in the case studies (a total of 26 sites, distributed around the UK). Further support comes from Lowe *et al* (2001), who found that to a first-order approximation, modest amounts of time-mean sea level rise and changes in surge can be added linearly around the United Kingdom. Sensitivity tests by Howard *et al* (2010) for larger time-mean sea level increases drew a similar conclusion, even for mean sea level rise in excess of 2 metres. Other studies which have taken a similar approach include Woth *et al* (2006) and Vousdoukas *et al* (2018), who provide further justification and supporting references.

In addition to the evaluation described in section 2.2, the CS3 model has been extensively evaluated as an operational model, and a version (CS3X) is used for UK storm-tide forecasting. Model results are routinely compared against observations, and show typical RMS errors of around 10 centimetres (e.g Furner *et al* 2016). Further details of model evaluation are given by Palmer *et al* (2018).

Our metric of tidal change is based on UKCP18-Marine projections of change in the standard deviation of the tide. To give a sensible first-order metric to describe tidal changes, we make an interpretation of the standard deviation change to give an estimated change in the high tides, based on simple assumptions of a sinusoidal tide, and that the change is shared symmetrically between high and low tides. This metric—the estimated change in the high tides—is labelled 'T' in figure 3.

For UKCP18-Marine, projections of tidal change were presented for a range of MSL rise projections. To make a like-for-like comparison we wish to represent the tidal change corresponding to the RCP8.5 projection of MSL rise. To achieve this we interpolate between the projected tidal change under 0.5 metres MSL rise and 1 metre MSL rise to obtain an estimate of the tidal change under 0.6 metre of MSL rise. This is a representative 21st-century figure for the UK under RCP8.5 (because the mean of the central estimate of 21st century MSL increase over the UK mainland tide gauges considered here is 0.607 metres under RCP8.5).

Our tidal simulations are relatively crude because we do not include any models of coastline change or sediment morphology. The increased MSL is simply represented by increasing the bathymetric depth of the existing active grid boxes of the European shelf model (in other words, the existing *inactive* grid boxes act as vertical side walls, allowing no further inundation). Thus, our simulation of the increased MSL does not accommodate the sort of changes in the coastline that might be expected under substantial MSL rise, which might require additional active grid boxes in locations that are currently assumed inactive (dry). In this context, Austin (1991) used a numerical tidal model simulation to study the effects of a uniform depth change of the order of the Holocene eustatic variation on the M2 tide of the NW European continental shelf. Under these large sea-level changes they found amplitude changes from ~0.5 metres to present-day values of ~2 metres for the Southern Bight of the North Sea, following the breach of the Strait of Dover. Thus our use of vertical side-walls could be a significant caveat to our tidal simulations. This caveat is acknowledged by Palmer *et al* (2018) for their tidal change projections under millennial-scale MSL rise of up to 10 metres. However, in spite of the simplicity of the approach, they find spatial patterns and typical magnitudes of change in tidal range that are comparable to those found in more sophisticated studies (e.g. Pickering *et al* 2012). Furthermore, we consider here results under a MSL rise of only 0.6 metre, so we do not expect this limitation to be very important in our case. A paleotidal study (Uehara *et al* 2006) found sensitivity of the European shelf tides to the open ocean tidal changes associated with the large Holocene eustatic sea-level change (~130 metres). We used (modelled) present-day open ocean tides as the open boundary conditions in our tidal simulations and thus further work would be needed to test this sensitivity in our case of much smaller sea-level change. However, previous studies of the effect of century-scale MSL change (for example Idier *et al* 2017, Flather and Williams 2000) have typically

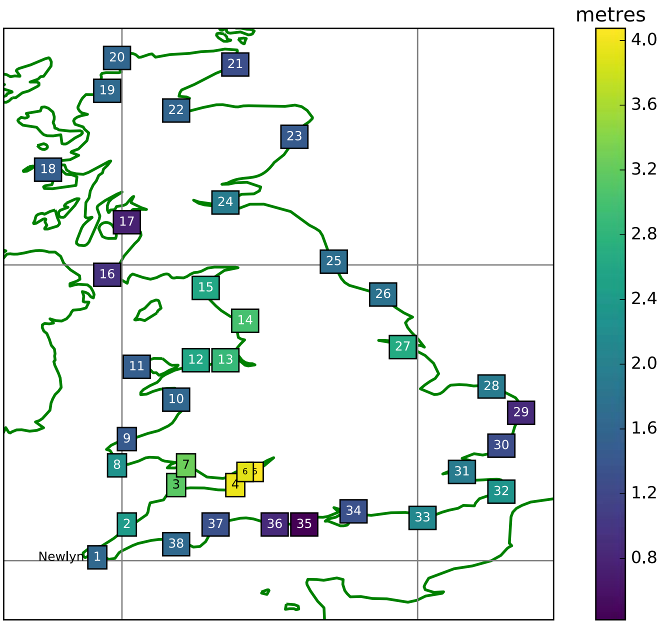

Figure 1. Locations around the UK mainland coast shown by their indices (these correspond to the indices on the *X*-axis of figure 2). The labels are coloured by a simple characterisation of model tidal amplitude (1.41 × standard deviation of elevation from a ∼19 year tide-only simulation), intended to give a value of half the typical tidal range, and consistent with the metric described in section 2.4. The locations are tabulated in table A1, and also listed here for ease of reference. 1: Newlyn, 2: Padstow, 3: Ilfracombe, 4: Hinkley, 5: Avonmouth and 6: Newport (in extra small font due to close proximity at this scale), 7: Mumbles, 8: Milford Haven, 9: Fishguard, 10: Barmouth, 11: Holyhead, 12: Llandudno, 13: Hilbre Island, 14: Heysham, 15: Workington, 16: Portpatrick, 17: Millport, 18: Tobermory, 19: Ullapool, 20: Kinlochbervie, 21: Wick, 22: Moray Firth, 23: Aberdeen, 24: Leith, 25: North Shields, 26: Whitby, 27: Immingham, 28: Cromer, 29: Lowestoft, 30: Felixstowe, 31: Sheerness, 32: Dover, 33: Newhaven, 34: Portsmouth, 35: Bournemouth, 36: Weymouth, 37: Exmouth, and 38: Devonport. Straight grey lines show the zero (Greenwich) meridian, the 5 degrees west meridian, and the 50 and 55 degree north latitudes.

Table 1. The contributions considered here, and the labels used for them.

Label	21st-century change
MSL	Time-mean relative sea level
S_H	1-year return level of skew surge from HadGEM2-ES-RCA4 simulation
S_G	1-year return level of skew surge from GFDL-ESM2M simulation
W	Annual mean Significant Wave Height from EC-EARTH regional simulation
T	Amplitude of tide

assumed negligible effects on the open-ocean tides. Table 1 summarizes the metrics of the potential 21st century contributions to change in sea-level extremes that we use here. In all cases, the results effectively represent long-term mean change over the 21st century. Inter-annual variations have effectively been smoothed out.

3. Results and discussion

We consider locations around the UK mainland coast, as shown in figure 1. These are based on the locations of 38 mainland tide gauges. We compare sizes and patterns of the four contributions for these locations in figure 2. Figure 2 panel (a) shows projected changes in skew surge, using the metrics described in section 2.2. Panel (b) shows projected changes in SWH, using the metric described in section 2.3. Panel (c) shows projected changes in tide, using the metric described in section 2.4. All panels also show the projected MSL change as described in section 2.1.

Under our selected RCP8.5 scenario, MSL is projected to increase all around the UK by an amount of the order of half a metre. Uncertainties are also of the order of half a metre and are not symmetrical about the central estimate, showing some positive skew (e.g. (5th, 50th and 95th) percentiles of 21st century projected change for Newlyn: (0.49, 0.71, 1.02) metres). Spatial variation is of the order of twenty centimetres, and is most pronounced on the large spatial scale, with the largest increases in the south and the smallest in the north: the central estimates vary spatially from 0.46 metres (at Millport, in the north west) to 0.71 metres (at Newlyn, in the south west). This spatial variation mostly comes from the spatial pattern of glacial isostatic adjustment.

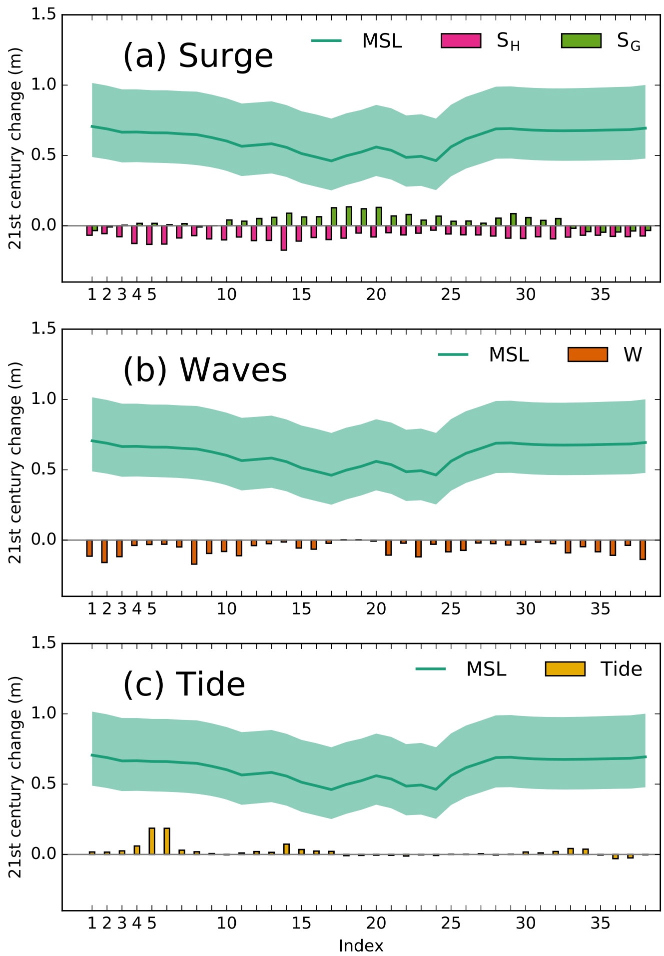

Figure 2. Contributions to 21st-century extreme sea-level change clockwise around the UK mainland coast from Newlyn. (a): Surge changes. S_H: HadGEM2-ES-RCA4. S_G: GFDL-ESM2M. (b): Wave changes: change in mean SWH. (c): Change in tidal amplitude. All panels also show projected MSL change (green line) with uncertainty shown by the green shading. Further details are in section 2. Indices on the X-axes refer to the locations shown in figure 1.

Surge changes are of order 10 centimetres, but models disagree on the sign of change. However, there is a high level of spatial coherence within each model projection: the HadGEM2-ES-RCA4 projection is negative around the whole of the UK coastline, whilst the GFDL-ESM2M projection is positive except on the south coast, where there is a weak but uniform negative signal.

Our projection exhibits a decrease in offshore SWH of between zero and 17 centimetres all around the UK, with substantial small-scale spatial variations. The largest decreases are on the south and south-west coasts. The general pattern of reduction in mean SWH seen in figure 2 is reflected in an ensemble (shown in UKCP18-Marine) of wave model simulations based on a subset of the CMIP5 models. This consistency across the ensemble, and the spatial coherence of this metric, increases our confidence in the mean offshore SWH change projection. However, the limited sample size and, more importantly, the uncertainty in projections of changes in atmospheric circulation means that we cannot be confident about the sign of future changes in offshore wave climate. Furthermore, as discussed in section 3.2, some recent studies (e.g. Arns *et al* 2017) have shown that the reduction in the coastal depth-limitation of the waves, associated with MSL rise, may cause significant increases in the wave energy reaching the shore. Thus, although we project a decrease in mean offshore SWH at all sites, this may be less relevant inshore as the increased MSL will oppose this change, by reducing the effect of coastal depth-limited breaking. This is discussed further by Bricheno and Wolf (2018). They note that this caveat may be more important for changes in the extreme waves (which are not discussed here).

Our projected tidal amplitude changes are smaller than the other contributions except in the Bristol Channel, which is further discussed below.

We show details of three example sites in figure 3. This shows time series of simulated annual maximum skew surge and bar charts comparing the representative metrics of potential contributions to extreme sea-level change as described in section 2. The time-series of simulated MSL rise (relative to a baseline of the 1980–2000 mean) is shown for comparison (although the skew surge does not add directly to the MSL, but rather to the high

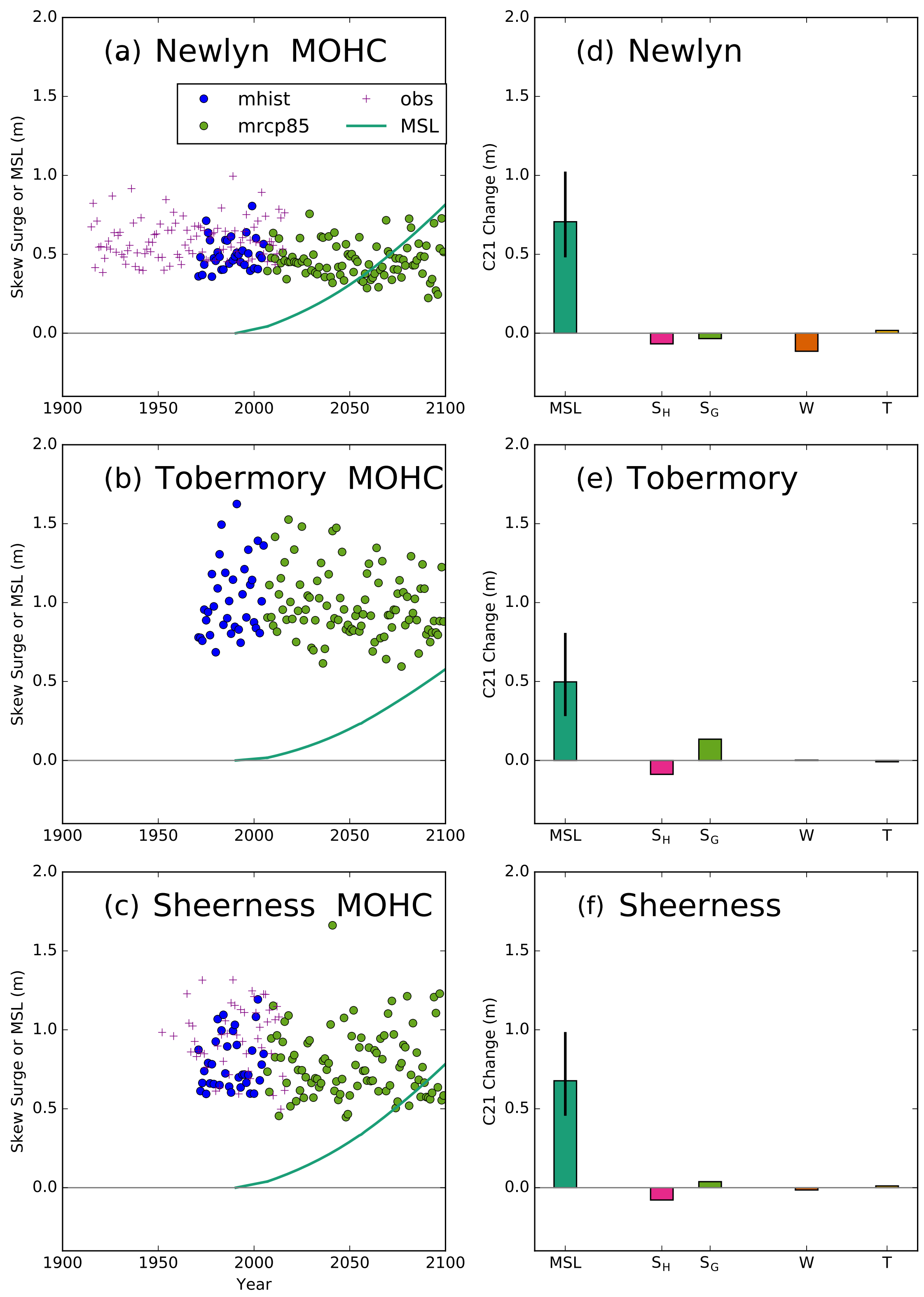

Figure 3. (a)–(c): Time series of modelled annual maximum skew surge for the historical period (points labelled 'mhist') and the RCP8.5 21st century (points labelled 'mrcp85') from the HadGEM2-ES-RCA4 simulation at (a) Newlyn, (b) Tobermory and (c) Sheerness. Time series of simulated MSL rise under RCP8.5 (relative to a baseline of the 1980–2000 mean) at the same location is shown by the curve ('MSL'). Observed skew surge annual maxima ('obs') are also shown for Newlyn and Sheerness. (Please see main text for a note on the simulated versus observed annual maxima.) (d)–(f): representative 21st-century projections of the contributions to extreme sea-level change from time mean sea-level (MSL), surge (S_H, S_G), waves (W), and tides (T).

tides). Inclusion in figure 3 of observed skew surge annual maxima for two of the sites illustrates the realism of the model simulations. The simulated and observed annual maximum for any given year of the historical period are not directly comparable because although the model includes historical forcing by greenhouse gases, ozone concentrations, solar variations, volcanoes and aerosols, the year-to-year variability will not be in phase with the

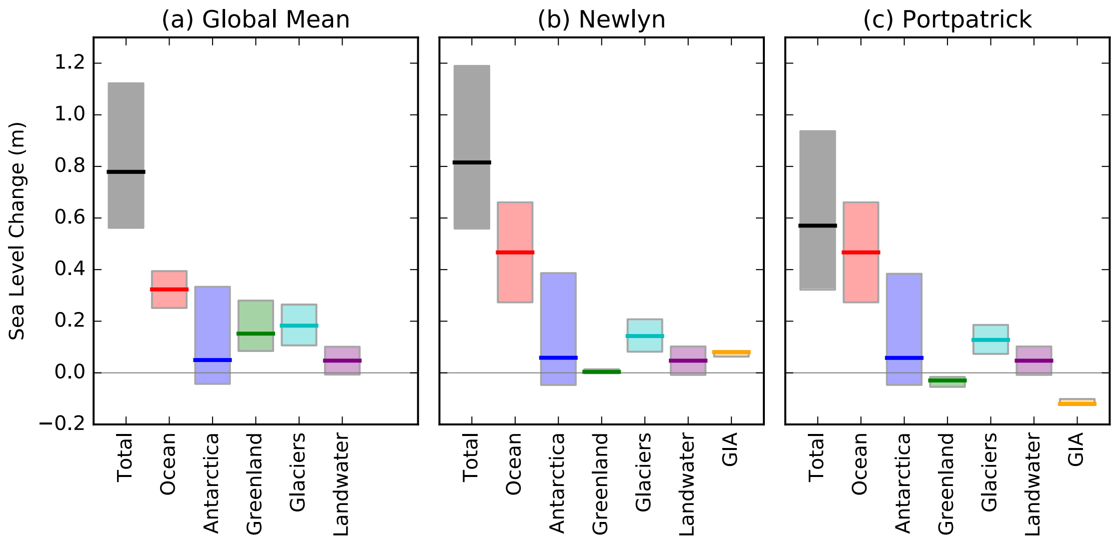

Figure 4. (a) Global 21st-century MSL change projections and component terms from UKCP18-Marine under the RCP8.5 climate change scenario; (b) and (c) corresponding regional sea-level projections for Newlyn and Portpatrick (locations 1 and 16, respectively). Horizontal lines indicate the 50th percentile and the shaded regions represent the 5th to 95th percentile range. 'Ocean' refers to global thermal expansion at the global scale and also includes regional oceanographic effects for Newlyn and Portpatrick.

observations. However, in addition to the visual similarity seen in figure 3 between the model and observed distribution of annual maxima at these two sites, Palmer *et al* (2018) show further details of satisfactory surge model evaluation at a selection of sites. In particular, the inter-annual variability is generally well simulated.

It is apparent from both figures 2 and 3 that the MSL rise is the projected dominant 21st-century contribution to the change in the extremes. This is consistent with other projections for the UK (e.g. Lowe *et al* 2009), for the north-west European coast (e.g. Sterl *et al* 2009, Vousdoukas *et al* 2017), and for many other locations worldwide (Garner *et al* 2017, Church *et al* 2013). In terms of the historical evidence, Menéndez and Woodworth (2010) showed that MSL change made the dominant contribution to changes in the extremes over the 20th century.

Our projections show MSL rise for our chosen UK mainland locations to be typically around five times the size of any of the other contributions over the 21st century. There is a spatial pattern in the projected MSL change, with the largest changes in the south and the smallest in the north.

In order to illustrate the drivers of spatial variations in projected MSL around the UK, we present the breakdown of the different components for two example locations alongside the corresponding global MSL projections (figure 4). The comparison of global and regional projections illustrates an overall increase in uncertainty at local scales and particularly the strong attenuation of the Greenland signals, which is associated with the UK's relative proximity to that ice sheet (close to the zero line in the associated mass fingerprints). The contribution to regional UK MSL change is dominated by the ocean term (including the effects of global thermal expansion and local oceanographic effects) and the future loss of ice mass from Antarctica. These two terms also dominate the overall uncertainty, with the skewness in the Antarctica contribution giving rise to the skewness in the projections of total MSL. Spatial variations in projected MSL rise around the UK are dominated by differences in GIA, with a small additional contribution from spatial gradients in the Greenland mass fingerprints across the UK. Our GIA projections come from a 15-member ensemble produced as part of the NERC BRITICE-CHRONO project (Bradley *et al* 2018), which utilised a recently-updated regional sea-level database. The aim of this study was to use only ice sheet reconstructions that fitted to the observed data across UK and Europe (Bradley, pers. comm.) It is noticeable that the uncertainty estimate from this source is smaller than from the other sources, and smaller than the uncertainty estimates of some recent global studies (e.g. Caron *et al* 2018). It is likely that two factors contribute to this difference in uncertainty estimates: first, the British-Irish Ice Sheet is relatively well-constrained observationally, and secondly the inability of current global models to simulate horizontal variations in mantle viscosity (Lambeck, pers. comm.) seems likely to introduce larger uncertainties in global projections.

Since publication of the IPCC AR5 a number of research studies have focused on the potential for accelerated sea-level rise associated with dynamic instability of the West Antarctic Ice Sheet (e.g. Ritz *et al* 2015, Golledge *et al* 2015, DeConto and Pollard 2016, Ruckert *et al* 2017). In particular, DeConto and Pollard (2016) proposed a controversial new mechanism called 'marine ice-cliff instability' (MICI), which led to projections of sea-level rise from Antarctica of over 1 metre by 2100 under the RCP8.5 scenario. However, subsequent work (Edwards *et al* 2019) has questioned the validity of including MICI in projections of future sea-level rise and demonstrated that the majority of projections are broadly consistent with the estimates of Levermann *et al* (2014) that were used in UKCP18.

The work presented here does not consider so-called 'high-end' scenarios for MSL, which explore plausible but very unlikely future levels of rise. Decision makers with a high level of risk aversion, for example those who design to a ten-thousand year return level, should consider such scenarios in addition to the projections described here and presented in UKCP18-Marine (2018). As noted by Hinkel *et al* (2015), 'high-end' scenarios have been used very effectively in coastal decision making, for example in the Thames Estuary 2100 project (Ranger *et al* 2013, Lowe *et al* 2009).

3.1. Surge

A high degree of spatial coherence is seen in S_H (figure 2). However, this is just one model from an ensemble of five CMIP5 models downscaled through RCA4 under the Euro-CORDEX experiment and used to produce surge change simulations. This small ensemble did not agree on the sign of change. In view of the disagreement and the relatively small signal in the individual models, Palmer *et al* (2018) concluded that a central estimate of no change is the best representation of the storm-surge contribution to extreme sea-level change over the 21st century. We note that the sign of change in the GFDL-ESM2M (S_G) simulation opposes that of the S_H simulation around most of the UK coastline. Thus we regard S_H and S_G as being representative of typical magnitudes of change which might be seen over the 21st century, rather than robust projections of climate change. In particular, the S_G projection should not be viewed as an upper limit to the 21st century change in skew surge. However, the 21st-century surge increase exhibited by the GFDL-ESM2M simulation is consistent with an increase in atmospheric storm activity in that model as measured by two different metrics of storminess (band-pass filtered atmospheric pressure at mean sea level, and storm count density; for details see Palmer *et al* 2018). Furthermore, the spatial pattern of increase around the UK mainland coast seen in figure 2, with strongest increases to the north, is consistent with the spatial pattern of increase in storm count density in that model (Lee, R., University of Reading URL cited 2019). The HadGEM2-ES model does not exhibit similar strong increases in these metrics of storminess. This consistency between atmospheric storminess change metrics and our metrics of storm surge change gives more confidence that the modelling system and our metrics of storm surge change are successfully identifying changes in the driving atmospheric models and thus identifying the disagreement between the driving atmospheric models regarding 21st century change. The atmospheric changes in turn may reflect disagreement between climate models regarding the response of the storm tracks to global warming (Shaw *et al* 2016, Shepherd 2014), or they may simply be attributable to long-term variability in the storm tracks, or both.

Woth *et al* (2006), using a barotropic tide-surge model driven by a small ensemble of atmospheric models of the CMIP3 generation, projected increases in extreme storm surge for the eastern coast of the North Sea. However, this increase was not seen on the east coast of the UK. Sterl *et al* (2009) found no significant increase in the most extreme surges over the 21st century along the Dutch coast based on a 17-member initial condition ensemble of a single model of the CMIP3 generation, which was used to drive a barotropic storm surge model. Vousdoukas *et al* (2016) used an ensemble of eight CMIP5 atmospheric models to drive a state-of-the-art storm surge model in order to investigate potential changes over the 21st century. One of the models was GFDL-ESM2M, and considering their projections for RCP8.5 for the end of the 21st century around the UK we see some similarities to our GFDL-ESM2M projection, with increases to the north which are not duplicated in the south of the UK coastline. A common factor in their work, the present study, and Lowe *et al* (2009), is that the projected changes on the English east coast are small and/or negative for the end of the 21st century under RCP8.5.

Lerwick, on the Shetland Isles in the far north-east of the UK (not shown here), and Newlyn, in the far south west, are relatively open-ocean sites, where surges are less dependent on generation by wind stress, and more dependent on the inverse barometer effect, compared to typical UK sites. Thus, if the small changes in surge were coming primarily from changes in atmospheric pressure variability, we might expect to see an amplification in the small signal of change at these sites relative to other UK sites. Conversely, if they were coming primarily from changes in the surface wind, we might expect to see an attenuation of the small signal of change at these sites. However, in view of the uncertainties and small-spatial-scale noise in the patterns, we cannot clearly identify any such amplification or attenuation in the patterns of change in the either GFDL-ESM2M (S_G) or HadGEM2-ES (S_H) simulation: the spatial pattern simply continues through Lerwick and Newlyn in both cases. Thus, based on our results we cannot readily separate the effects of changes in the inverse barometer effect from the effects of changes in the wind stress. Further 21st-century simulations (pressure-driven-only and wind-driven-only) would be required to address this question.

It is apparent from figure 3 that the inter-annual variation in the annual maximum skew surge is comparable to the projected MSL rise at the three sites shown. The uncertainty in the projected MSL rise is also of a similar size. Combining the uncertainties in projections of MSL rise with the variability in still water extremes in a meaningful way is not straightforward (Hunter *et al* 2013) and it is anticipated that this will form the basis for future work. In the meantime, Palmer *et al* (2018) and Howard *et al* (2019) provide projected future return level curves of still water level based on a simple addition of MSL rise to the present-day return level curves.

The implications of MSL change in terms of return period for a given location will depend on the variability in the extremes at that location. To illustrate this, consider two contrasting locations: Avonmouth, which experiences high variability, and Lerwick, which experiences low variability. Now consider our RCP8.5 central estimate for 2100 as an example scenario of MSL rise. Under this scenario, at Avonmouth the 2100 one-year return level would match the present-day forty-year return level, but at Lerwick the 2100 one-year return level would be higher than the present-day ten-thousand-year return level.

3.2. Waves

Our projection of a decrease in mean SWH agrees with the results of Aarnes *et al* (2017), who also note a smaller reduction in higher percentiles of SWH, suggesting an increase in the variance. Bricheno and Wolf (2018) discuss an alternative metric of wave changes, based on the mean of twenty annual maxima in SWH at the beginning and end of the 21st century. At some locations the projected change in that metric is positive (increasing) and comparable to the projected MSL change. It is possible for the extremes to increase whilst the mean decreases: this only needs a widening of the distribution. However, there are three reasons for having less confidence in that result. Firstly, for most plausible distributions of environmental metrics such as SWH, we expect the maxima to be more noisy (variable) than the means, in view of the law of large numbers. Secondly, there is no clear consensus among ensemble members (Palmer *et al* 2018) regarding the change in that metric. Thirdly, there is little spatial coherence in the change in that metric.

The general picture of a projected reduction in mean SWH was also see in Lowe *et al* (2009), but not over the same geographical extent: in fact Lowe *et al* (2009) reported a small increase in winter mean SWH to the south of the UK. They noted that their pattern of change was consistent with a reduction in northerly winds and a strengthening of westerlies. Grabemann *et al* (2015), using an ensemble approach, also projected decreases in SWH in the western North Sea, but accompanied by increases in the eastern North Sea.

The effect of atmospheric storminess change on the offshore waves is only one of the climate-change factors which may affect the future impact of waves at the coast. For example, Arns *et al* (2017) studied the non-linear interactions between tides, surge, waves and MSL change. In particular they found that projected changes in wave characteristics (and, to lesser extent, tides) caused by sea-level rise, amplified the required design height changes by ~50% compared to the required changes due to projected sea-level change alone, primarily due to the reduction in the coastal depth-limitation of the waves. Chini *et al* (2010, 2011) used a numerical modelling approach to investigate the interplay of MSL change, tides, surge, waves and changing sediment transport for a case study area in East Anglia, UK. They found a number of complex, sometimes compensating, responses to increased MSL. One clear finding was that, in the absence of any bathymetry changes (which might arise as sandbanks adapt to sea-level rise by capturing more sediment), MSL rise has more effect on extreme wave heights at locations that are currently protected by offshore sandbanks than at locations which are currently more exposed to the open sea (Chini and Stansby 2015). This level of detailed hydrodynamic modelling is beyond the scope of the present work and so we do not translate our projected offshore SWH changes into a metric which is directly comparable with MSL, although we note that Vousdoukas *et al* 2017, use a generic approximation of the wave setup as 20% of the SWH. Since our projections for change in SWH are uniformly negative, we suggest that, based on the current generation of climate models, it seems this offshore component is unlikely to add to the inshore hazard.

3.3. Tide

Our projected changes in tide under the 0.6 metre of MSL rise representative of RCP8.5 (figure 2 panel (c)) are small except around Avonmouth and Newport, near the top of the Bristol Channel. The Bristol Channel is near-resonant with the dominant (M2) tidal constituent. Projections of tidal change using CS3 (our projections, and those of Flather and Williams (2000)) show an increase in tidal range here with increasing MSL. However, some other similar experiments with different shelf models have exhibited the opposite response, i.e. a decrease in tidal range (Pickering *et al* 2012, Pickering 2017, Pelling *et al* 2013) in the Bristol Channel, even though the patterns and magnitudes of change show a broad agreement elsewhere. Presumably the CS3 model channel moves closer to resonance, whilst the model channel in some other models moves away from resonance, and we speculate that this difference is associated with the model resolution and/or the bathymetry dataset used. Furthermore, as noted by Idier *et al* (2017), our projections of tidal change suffer from the lack of a credible model of century-scale morphological change. Some areas within out model domain (such as North Sea tidal sandbanks) have construction doubling times of the order of centuries and thus would not be expected to keep pace with sea-level rise, whereas other regions such as coastal dunes might do so (Idier *et al* 2017). Thus, we have low confidence in specific geographic details of changes in tidal range, but greater confidence in the results as providing an indication of the magnitude of change which might arise due to this contribution under ~0.6 m of MSL rise. Given larger MSL rise, simulations suggest that the changes in tidal range are not proportional to the MSL rise, for some locations (see UKCP18-Marine 2018).

Since we focus on projected century-scale changes, we have not considered the interplay of the mean sea-level change and the 18.6-year nodal cycle in tide amplitude, which can be expected to be important for the nearer-term. For example, Talke *et al* (2018), in a study focused on Boston, Massachusetts, concluded that this interaction produces a decadal-scale fluctuation in sea-level hazard.

4. Summary and conclusions

We have presented a synthesis of projections of 21st century change in sea-level extremes associated with changes in time-mean sea level (MSL), waves, tide and surge, developed for the United Kingdom Climate Projections 2018 Marine Report, and we have compared the sizes of these contributions.

Our simulations show that projections of 21st century MSL change (and its uncertainty) dominate over the sources (surge and waves) which depend on changes in the storm tracks. In contrast to the projected changes in surge and waves, the inter-annual variation in the extremes even under present-day conditions *is* comparable to the projected MSL change uncertainties in many locations.

We conclude that, whilst there is an ongoing need for research leading to a scientific consensus on future storm track changes, until such consensus begins to emerge coastal sea-level research effort might most usefully focus on constraining the MSL change projections, and on the implications of the MSL change uncertainties for coastal planning— for example by combining uncertainties in projected MSL change with a probabilistic expression (as is usually shown by a return-level curve) of the present-day distribution of extremes.

Based on our surge simulations, we conclude that a central estimate of no change is the best representation of the storm-surge contribution to extreme sea-level change over the 21st century.

Projected tidal changes are also small compared to the MSL contribution. However, in contrast to the surge and wave projections, tidal changes do not depend on the highly-uncertain and poorly-understood atmospheric changes, and in that sense they can be considered more robust. The physical processes driving this contribution are fewer, better understood, and easier to model than the drivers of atmospheric circulation changes and so it is probable that uncertainties in this small contribution will respond more readily to improvements in models and experimental design.

We note two caveats to the results presented here: first, although our metric of SWH shows a 21st century reduction, Bricheno and Wolf (2018) discuss a less-robust metric based on the extremes of SWH, which shows an increase for some UK locations. Second, recent work has indicated the importance of non-linear effects of MSL rise on the other components of change in sea-level extremes, particularly the effect on the inshore wave climate, and more research is required on these non-linear interactions.

These new simulations provide updated information for adaptation planning, and there is work to do within coastal climate services (Cozannet *et al* 2017) to get this information to policy makers. Furthermore, given that the largest contribution relates to MSL rise, there is an urgent need to both continue to monitor this metric and narrow the major uncertainty—from land ice melt.

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Author contributions

MDP made the time-mean sea-level change simulations and lead on the United Kingdom Climate Projections 2018 Marine Report. LMB made the wave change simulations. TH made the surge change simulations and lead on the writing here.

Table A1. This table shows the projected SWH change metric used in figures 2 and 3.

Index	Site	ΔSWH (m)	Index	Site	ΔSWH (m)
1	Newlyn	−0.114 3	20	Kinlochbervie	−0.007 2
2	Padstow	−0.159 2	21	Wick	−0.106 7
3	Ilfracombe	−0.117 5	22	Moray Firth	−0.021 7
4	Hinkley	−0.037 8	23	Aberdeen	−0.119
5	Avonmouth	−0.031	24	Leith	−0.029 4
6	Newport	−0.029 3	25	North Shields	−0.083 1
7	Mumbles	−0.048 6	26	Whitby	−0.072 9
8	Milford Haven	−0.170 9	27	Immingham	−0.020 7
9	Fishguard	−0.094 7	28	Cromer	−0.024 9
10	Barmouth	−0.080 8	29	Lowestoft	−0.035 1
11	Holyhead	−0.110 9	30	Felixstowe Pier	−0.031 3
12	Llandudno	−0.038 9	31	Sheerness	−0.014 5
13	Hilbre Island	−0.025 3	32	Dover	−0.025 1
14	Heysham	−0.013 6	33	Newhaven	−0.090 3
15	Workington	−0.056 3	34	Portsmouth	−0.046 8
16	Portpatrick	−0.064 5	35	Bournemouth	−0.082 8
17	Millport	−0.022 5	36	Weymouth	−0.108 1
18	Tobermory	0.002 2	37	Exmouth	−0.037 1
19	Ullapool	0.002 2	38	Devonport	−0.137 3

Appendix. Wave data

Table A1 shows the projected SWH change metric used in figures 2 and 3.

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