Detection and Attribution of Climate Change Signal in Ocean Wind Waves

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ABSTRACT

Surface waves in the ocean respond to variability and changes of climate. Observations and modeling studies indicate trends in wave height over the past decades. Nevertheless, it is currently impossible to discern whether these trends are the result of climate variability or change. The output of an Earth system model (EC-EARTH) produced within phase 5 of the Coupled Model Intercomparison Project (CMIP5) is used here to force a global Wave Model (WAM) in order to study the response of waves to different climate regimes. A control simulation was run to determine the natural (unforced) model variability. A simplified fingerprint approach was used to calculate positive and negative limits of natural variability for wind speed and significant wave height, which were then compared to different (forced) climate regimes over the historical period (1850–2010) and in the future climate change scenario RCP8.5 (2010–2100). Detectable climate change signals were found in the current decade (2010–20) in the North Atlantic, equatorial Pacific, and Southern Ocean. Until the year 2060, climate change signals are detectable in 60% of the global ocean area. The authors show that climate change acts to generate detectable trends in wind speed and significant wave height that exceed the positive and the negative ranges of natural variability in different regions of the ocean. Moreover, in more than 3% of the ocean area, the climate change signal is reversible such that trends exceeded both positive and negative limits of natural variability at different points in time. These changes are attributed to local (due to local wind) and remote (due to swell) factors.

1. Introduction

Wind waves control the basic physical processes, such as heat, momentum, and mass exchange between ocean and atmosphere (Cavaleri et al. 2012). Wind waves generate additional turbulence, modify ocean currents, and control the state of the sea surface. All these processes affect the air–sea exchange, general circulation patterns, and wind in the atmosphere, which in turn control wind waves. This cycle establishes a wind–wave feedback loop in the Earth system. In line with increasing global temperature, rising sea level, and melting polar ice, variations in atmospheric circulation under ongoing climate change affect the global distribution of wind. With changing winds, a new distribution of wind waves can lead to an intensification or weakening of exchange processes of ocean–atmosphere interaction in many regions over the globe.

Observations (e.g., Young et al. 2011; Allan and Komar 2000) and modeling studies (e.g., Semedo et al. 2013; Shimura et al. 2013; Hemer et al. 2013; Dobrynin et al. 2012; Wang and Swail 2006) indicate short- and long-term trends in wave height over the recent past and
in future climate projections. Regional trends contain signatures driven by local wind patterns in wave regimes and of remotely generated swell. Following the seasonal cycles in Earth’s climate, ocean surface waves have a strong variability because of changing wind conditions and propagating swell. For instance, in a recent modeling study, Dobrynin et al. (2012) demonstrated that wave generation regions such as the Southern Ocean are projected to undergo pronounced increase in the wind speed and significant wave height in the near future under a future scenario of climate change, RCP8.5 (Stocker et al. 2013; van Vuuren et al. 2011), which represents a high-CO₂ world. Ocean waves are also projected to increase in the ice-free future Arctic Ocean (Khon et al. 2014). In contrast, for the main part of the North Atlantic, a decrease of wind speed and significant wave height is projected. As a result of increasing winds and waves, the signal of climate change propagating from the Southern Ocean can be found also in regions with no significant changes or even with a decrease in wind speed (e.g., in the southern Pacific). Attribution of changes in wave climate therefore has two components: (i) local wind waves and (ii) propagating swell.

Together with variability and changes in wind speed and direction, contraction and expansion of sea ice cover in the Arctic and the Southern Ocean also affect wave climate. Generally, the generation of sea ice hampers the evolution of waves, whereas ice melting has the opposite effect. Also, waves propagating through sea ice induce ice breaking (Kohout et al. 2014; Thomson and Rogers 2014). These interactions between sea ice and ocean waves are not well understood and currently are not commonly included in model simulations. Furthermore, by the end of the twenty-first century, intensified heat transport into the Arctic as a result of climate change (Koenigk and Brodeau 2014) is projected to open presently ice-covered regions in the Kara Sea, Laptev Sea, and Barents Sea in boreal winter and to remove the ice cover completely in boreal summer. These changes in sea ice lead to a farther north- and eastward propagation of Atlantic swell and wind-generated waves into the polar regions. In contrast to the Arctic, the sea ice extent around Antarctica has had a positive trend over the last few decades (1985–2010; Turner and Overland 2009). According to a modeling study based on the EC-EARTH model (Bintanja et al. 2013), this tendency is projected also for the coming years/decades. The mechanism behind extending sea ice in the Southern Ocean is not well understood but could be related to the accelerating warming of the subsurface Southern Ocean and melting intensification of the Antarctic shelf ice (Hellmer et al. 2012; Yin et al. 2011), which in turn contributes to the cooling of the sea surface temperature. This trend, however, is not commonly reproduced in global climate models (Stocker et al. 2013).

In summary, it is still uncertain whether observed and projected trends in ocean waves are the result of natural climate variability of sea ice, wind speed, and significant wave height or transient climate change. The limited time of wave observations generally conducted in the last few decades deliver datasets in which natural variability and climate change signals are superimposed. Hence, it is impossible to separate them based on observations alone. Numerical models are useful tools to reconstruct the wave climate under the preindustrial climate state, which is necessary to estimate the natural variability in wind and waves.

Comprehensive Earth system models (ESMs), such as those used within phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012), have served as the basis for the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (Stocker et al. 2013). These CMIP5 projections provide long (several hundreds of years) and internally consistent datasets of climate parameters. In ESMs, climate change as a consequence of natural variability and anthropogenic activities is represented by transient forcing by aerosols and greenhouse gases, as well as by varying incident solar radiation. In the frame of the CMIP5 project, models simulate Earth’s climate starting from preindustrial conditions in the year 1850 and into the future, through the year 2100 and beyond. These projections include a historical simulation (covering the period of 1850–2005) and a number of future climate change scenarios termed as representative concentration pathways (RCPs; van Vuuren et al. 2011) spanning the years 2006–2100. Additionally, the CMIP5 set of experiments includes a control simulation with constant forcing (i.e., mean solar radiation and representative greenhouse gas and aerosol concentrations for 1850). This control simulation ensures that neither the reference wind speed nor other model parameters have any significant long-term trends related to climate change. Because the control simulation was performed without anthropogenic forcing, it produces natural (unforced) climate variability. Consequently, the historical simulation and future climate change projections result in a climate including man-made signals; therefore, they represent a forced climate variability.

Wind speed and direction, as well as sea ice cover from the CMIP5 datasets, provide forcing data for wave model applications, which can be used for the investigation of complex interactions between natural variability and anthropogenic impacts in the wave climate. Following the
approach of Hasselmann (1993), the climate response can be represented as a superposition of a natural unforced and an external forced variability. In terms of wind–wave climate, its variability is a response to changes and variability in atmospheric circulation and climate states. The CMIP5 set of experiments can be used to obtain information on these unforced and forced variabilities.

Here we focus on the response of the global wave climate to ongoing climate change. We investigate effects caused by changing wind and sea ice cover. The goal of our study is to determine natural variability in the wave climate under preindustrial conditions and detect at which point in time in a future scenario a climate change signal significantly deviates from the natural variability. Note that modeling studies indicate both regions with increasing and decreasing waves in the future climate. Thus, in our study, we focus on finding statistically significant and therefore detectable positive and negative trends in the wave climate. Furthermore, we make an attempt to attribute the changes in wave climate to changes in wind (indicating local changes) and swell (indicating remote changes). Effects of rising sea level are limited to the near-coastal zones and have not been taken into account.

2. Methodology

a. Models

We used the output of an Earth system model EC-EARTH (Hazeleger et al. 2012) produced within CMIP5 to force the global Wave Model (WAM; WAMDI Group 1988). EC-EARTH includes atmospheric, oceanic, sea ice, and land components. The atmospheric model of the EC-EARTH’s version 2.3 (Hazeleger et al. 2012) that has been used in the CMIP5 experiments is based on the ECMWF Integrated Forecast System (IFS) cycle 31r1 (ECMWF 2006), with a modified convection scheme and a new land surface hydrology of the Tiled ECMWF Scheme for Surface Exchanges over Land (H-TESSEL) from the more recent cycles [see Hazeleger et al. (2012) for more details and references]. The ocean component of EC-EARTH is the Nucleus for European Modelling of the Ocean version 2 (NEMO2) developed by L’Institut Pierre-Simon Laplace (Madec 2008) with the Louvain-la-Neuve Sea Ice Model version 2 (LIM2; Fichefet and Maqueda 1997; Bouillon et al. 2009) and the Ocean Atmosphere Sea Ice Soil, version 3 (OASIS3) coupler (Valcke 2006). The ocean component features a higher nominal horizontal resolution (1°) compared to the atmospheric model (1.125°) which is higher in the tropics (being 0.33° in the ocean). The model has 42 vertical z-layers in the ocean, whereas the atmospheric component uses 65 vertical layers characterized by isobars.

Maps of sea ice extent are derived from the ice concentration distribution by setting a minimum value of 0.3, which represents a limit of 30% ice cover of the area of a grid cell. We have derived 3-hourly maps for sea ice extent from the output of EC-EARTH’s sea ice model component LIM2, which is available on the NEMO2 ocean model grid. The wave model treats ice-covered points as land, effectively blocking wave energy transport across ice-covered waters and limiting the fetch of newly generated waves that would otherwise grow unobstructed in ice-covered waters. Sea ice extent and concentrations are comparatively well represented by the EC-EARTH climate model, except for the summer values in the Southern Hemisphere, which are underestimated because of a warm bias that has been traced back to intensified solar heating and missing near-surface mixing in the upper ocean (Sterl et al. 2012).

The global setup of WAM has a resolution of 1° and uses the ETOPO1 ocean bathymetry (Amante and Eakins 2009). The model area covers a region from 84°N to 79°S. The wave spectrum covers the range of frequencies from 0.042 to 0.802 Hz (logarithmically scaled into 32 bins) and the directions from 7.5° to 325.5° (24 directions with a resolution of 15°). Separate analysis of wind sea and swell fractions makes use of the classification scheme implemented into WAM cycle 4.5 (Komen et al. 1994). The wind sea– and swell-related fractions of the wave energy spectra $F (f, \theta)$ (i.e., the frequency $f$ and direction $\theta$ dependent mean variance spectra of the surface elevation) are defined by the separation value $c_s = 33.6 U / \cos(\theta - \theta_o)$ of the phase speed $c_p$ of linear waves; where $c_p \leq c_s$ characterizes the wind sea and $c_p > c_s$ defines the swell part of the spectra. The significant wave height $H_s = 4.04 / \sqrt{m_0}$ is then provided by the zeroth spectral moment $m_0 = \iint F(f, \theta) df \, d\theta$ of either the full spectra (total sea) or the spectral components of wind sea and swell.

Winds at 10-m height from the EC-EARTH model have been used to force WAM. The 6-hourly EC-EARTH wind fields have been interpolated bilinearly in space onto the WAM grid as a preprocessing step, and are interpolated linearly in time during the run of WAM. Note that the 6-hourly wind forcing may filter out some extreme wind events, which are shown to have an effect on the wave climate (e.g., Chawla et al. 2013). Performance of the model WAM has been evaluated in a number of previous studies, in which it was compared to observations and reanalysis data (e.g., Semedo et al. 2013; Dobrynin et al. 2012; Weisse and Günther 2007). The setup used here has been evaluated against altimeter data by Dobrynin et al. (2012).
b. Numerical experiments

First, we simulated the natural variability of ocean winds and waves under unforced preindustrial climate conditions. A 100-yr wave simulation (control simulation; Table 1) was conducted using 10-m winds from the control simulation of EC-EARTH. This period was considered to be sufficient to include the effects of major multidecadal climate oscillations on waves.

Second, we simulated variability of ocean winds and waves under forced climate conditions based on: (i) the historical simulation from 1850 to 2010 (historical simulation; Table 1), and (ii) a future climate change projection from 2010 to 2100 (future scenario RCP8.5; Table 1). In the historical simulation, EC-EARTH followed the historical emissions of CO\textsubscript{2} from 1860–2005. To have a smooth transition from the historical simulation to the future projection, the historical simulation in EC-EARTH was extended with RCP4.5 forcing until 2010. For the future climate change projection, we use the RCP8.5 scenario, which represents a high-CO\textsubscript{2} world in order to understand the response of wave climate at the upper range of climate change given under CMIP5 projections. The integrated wave parameters, including significant wave and swell heights and wind speed, have been stored with an output time step of 6 h. Along with changing winds, also changes in the sea ice cover were considered in our model simulations.

In the following, we refer to the control simulation as unforced climate variability and to both the historical simulation and the future climate change projection as forced climate variability (Table 1). Note that both “forced” simulations inherently contain both forced and unforced variability.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Length, years</th>
<th>Period</th>
<th>Type of variability</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>100</td>
<td>—</td>
<td>Unforced</td>
</tr>
<tr>
<td>Historical</td>
<td>160</td>
<td>1850–2010</td>
<td>Forced</td>
</tr>
<tr>
<td>Future scenario</td>
<td>90</td>
<td>2010–2010</td>
<td>Forced</td>
</tr>
<tr>
<td>RCP8.5</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

In a first step, we calculated the range of natural variability of significant wave and swell heights, and wind speed in the control simulation (Table 1). For each horizontal grid point, we calculated an ensemble of trends varying in length from 1 to 50 yr (herein, 1-yr trends will be referred to as short-term, 50-yr trends as long-term). Note that the maximum trends length is determined by the length of the control simulation so that all 50-yr-long trends fully fit within the 100-yr-long control simulation. For each trend period, the entire control simulation is subsampled, allowing for overlapping samples. As a result, the number of samples from the 6-hourly output of the 100-yr control simulation ranges from about $1.45 \times 10^4$ for 1-yr trend periods to $7.3 \times 10^4$ for 50-yr trend periods. Note that since the control simulation uses constant forcing, it represents “model time” rather than real calendar years.

To describe the range of natural variability, a probability distribution function (PDF, normal distribution) is calculated for each trend period. The PDF confidence intervals of 0.05 and 0.95 are used as threshold values to define the upper and the lower limits of the unforced natural variability, as in Baehr et al. (2008). This approach provides an estimate of the range of unforced model variability under preindustrial climate conditions.

In the second step, we calculated linear trends in wind speed and significant wave and swell height data within the two forced simulations (Table 1). Starting from 1960 for the historical simulation, and from 2010 for the climate change scenario, linear trends (ranging from 1 to 50 yr) are estimated as before.

In a third and final step, we calculated when a forced simulation significantly deviates from the estimated range of natural variability. For each trend period, the trends within the respective forced simulation are compared to the upper and lower limits of natural variability estimated from the control simulation. If a trend from the forced simulation exceeds the upper or lower limits of natural variability for a given trend period, the difference between forced and unforced wave climate projections becomes statistically significant. We refer to this trend period as the detection period and to the year at which the unforced and forced variability are significantly different as the detection time. The detection time is determined from the starting point of the respective forced simulation plus the detection period over which the difference became detectable for the first time. In this way, our detection time depends on the reference years 1960 and 2010 for the historical and future climate change simulations, respectively.

In our detection analysis, we did not consider regions covered by ice in the control simulation. This detection method requires that the linear trends are calculated.
continuously in time. In regions in which sea ice cover has changed during the simulation (i.e., over the historical period and in the future climate change scenario) this condition is not fulfilled, and thus variability limits cannot be estimated.

3. Variability of wave climate

Increasing global mean values of significant wave height and wind speed (Fig. 1) are produced by the complex combination of unforced (natural) and forced (climate change) signals. In the historical simulation, the global mean of significant wave height and wind speed indicates different climate regimes. A quasi-“no trend” period (after the preindustrial unforced period) from 1850 to 1930 under a low-CO₂ concentration climate and only a small increase of CO₂ emissions is continued by a clear positive trend period until 2010, which partially extends into the future scenario RCP8.5. Starting from 1950, the climate change signal is pronounced in the global mean significant wave height and wind speed (Fig. 1). Time series of significant wave height and wind speed have clear, pronounced short- and long-term oscillations (Fig. 1). The spatial distribution of the short-term variability of waves is illustrated by the seasonal mean values of significant wave and swell heights (Fig. 2). Typical for the global wave distribution is an opposite storm dynamic in the Northern and the Southern Hemispheres. The maximal differences of the climatological seasonal values of significant wave height in the boreal winter season [December–February (DJF)] and the boreal summer season [June–August (JJA)] vary from 2.9 m in the Northern Hemisphere to 1.6 m in the Southern Hemisphere. Swell height varies from 2.0 to 1.1 m in the Northern and Southern Hemispheres, respectively. The difference between the boreal summer (JJA) and boreal winter (DJF) patterns strongly varies in the northern Atlantic, with stronger westerly winds and wind sea (not shown) during boreal winter season. In the southern Atlantic and Pacific, wind sea and swell are well developed during all seasons (most dominant during austral summer) and the natural short-term variability is lower.

The ongoing warming of the Arctic Ocean, as a consequence of intensified ocean heat transport, mainly through the inflowing water into the Barents Sea, and, to a smaller extent, through an increase in volume flux (Koenigk and Brodeau 2014), redistributes wave patterns in this region. Most of the water that enters the Barents Sea is exported into the Kara Sea between Franz Josef Land and Novaya Zemla, leading to intensified warming, sea ice reduction on the continental shelf, and unobstructed wave propagation. The ocean heat transport into the Barents Sea has grown substantially already in the twentieth century, and the growth is projected to intensify until the end of the twenty-first century, leading to reduced sea ice concentration and higher atmospheric temperature (Stocker et al. 2013).
FIG. 2. Climatological seasonal (top) mean and (bottom) anomalies of significant wave and swell heights calculated from the historical simulation in the period from 1850 to 1870 for (left) boreal winter (DJF) and (right) boreal summer (JJA). Seasonal anomalies of significant wave and swell heights are calculated from 20 yr of the future scenario RCP8.5 from 2080 to 2100 in comparison to the historical reference period from 1850 to 1870. Gray shaded area represents sea ice coverage.
The long-term decadal component of wave variability responds to major cycles in the Earth system that control the wind distribution, such as the North Atlantic Oscillation (NAO) or Pacific Decadal Oscillation (PDO; Tokinaga and Xie 2011). In our detection analysis of the 100-yr-long control simulation, the trend lengths of 1 to 50 yr include the natural variability driven by these oscillations. Most of these oscillations have a period of several years; therefore, the calculated global mean range of natural variability of significant wave height and wind speed [Fig. 3 (shadowed areas) and Table 2] shows a rapid drop from the annual to 5-yr period. It furthermore shows a moderate decrease in the following 20 yr and a low-to-constant rate of decrease in the last 20 yr. The range of variability varies from 1.90 to 0.12 m for the significant wave height and from 4.81 to 0.40 m s\(^{-1}\) for wind speed, respectively.

The positive and negative limits of natural unforced variability of wind speed and significant wave heights calculated from the control simulation were averaged for oceanic basins (Fig. 4). The range of variability changes by one order of magnitude depending on the period of variability such that short-term (1–5 yr) variability determined by seasonal and interannual signals is higher than the long-term (several decades) variability (Fig. 4, shadowed area; blue and red lines are discussed below). The highest ranges of 1-yr natural variability of 3.44 m for the significant wave height in the Arctic and 6.59 m s\(^{-1}\) for the wind speed are found in the North Pacific. Other regions with relatively high short-term variability are the North Atlantic, the Indian Ocean, and the Southern Ocean (Table 2). The long-term variability of significant wave height on the 50-yr scale varies from a maximum of 0.23 m in the Arctic to 0.06 m in the equatorial Atlantic (Table 2). In the North Atlantic and Pacific, the long-term variability of both the winds and the waves indicates the wind-driven character of the wave dynamics. In contrast, the equatorial Atlantic and Pacific demonstrate a low variability of the significant wave height, indicating a weaker dependency on the local wind than in other regions, whereas the wind speed variability is rather high. Wind speed variability intensifies southward (i.e., in the South Atlantic, the South Pacific, and the Southern Ocean,) and the long-term variability of significant wave height becomes stronger as well. The South Atlantic and South Pacific demonstrate an increase of the variability of significant wave height on top of low variability of the wind speed in

### Table 2. Range of variability of significant wave height (Hs) and wind speed (U) for the periods of 1 and 50 yr and mean values from the control simulation.

<table>
<thead>
<tr>
<th>Region</th>
<th>Hs, m</th>
<th>U, m s(^{-1})</th>
<th>Mean Hs, m</th>
<th>Mean U, m s(^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arctic</td>
<td>3.44–0.23</td>
<td>3.62–0.42</td>
<td>2.05</td>
<td>5.98</td>
</tr>
<tr>
<td>North Atlantic</td>
<td>2.94–0.16</td>
<td>5.89–0.42</td>
<td>2.32</td>
<td>7.83</td>
</tr>
<tr>
<td>North Pacific</td>
<td>3.17–0.15</td>
<td>6.59–0.46</td>
<td>2.69</td>
<td>7.83</td>
</tr>
<tr>
<td>Indian Ocean</td>
<td>2.12–0.10</td>
<td>5.25–0.37</td>
<td>2.91</td>
<td>7.67</td>
</tr>
<tr>
<td>Equatorial</td>
<td>1.17–0.06</td>
<td>5.58–0.36</td>
<td>1.87</td>
<td>5.95</td>
</tr>
<tr>
<td>Arctic</td>
<td>1.12–0.08</td>
<td>5.17–0.42</td>
<td>2.19</td>
<td>5.94</td>
</tr>
<tr>
<td>Equatorial</td>
<td>1.10–0.08</td>
<td>5.17–0.42</td>
<td>2.19</td>
<td>5.94</td>
</tr>
<tr>
<td>Pacific</td>
<td>1.12–0.08</td>
<td>5.17–0.42</td>
<td>2.19</td>
<td>5.94</td>
</tr>
<tr>
<td>South Atlantic</td>
<td>1.66–0.09</td>
<td>3.90–0.31</td>
<td>2.81</td>
<td>8.09</td>
</tr>
<tr>
<td>South Pacific</td>
<td>1.13–0.11</td>
<td>3.70–0.37</td>
<td>3.05</td>
<td>7.68</td>
</tr>
<tr>
<td>Southern</td>
<td>2.12–0.20</td>
<td>4.39–0.47</td>
<td>3.70</td>
<td>8.91</td>
</tr>
<tr>
<td>Ocean</td>
<td>1.90–0.12</td>
<td>4.81–0.40</td>
<td>2.67</td>
<td>7.25</td>
</tr>
<tr>
<td>Global</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
comparison to the equatorial regions, indicating a swell-dependent character of variability (Table 2).

4. Detection of climate change signal

Strong regional variability of wind speed and significant wave height leads to different regional impacts of climate regimes on wave climate. In some ocean basins at the higher latitudes, like in the North Atlantic and North Pacific, atmospheric circulation results in high values of wind speed and significant wave and swell height on top of a pronounced seasonal cycle. The North Atlantic and North Pacific are regions affected by strongly pronounced long-term (multiyear) atmospheric variability; therefore, climate change in these regions is stronger than in other regions. The consequences could be far

FIG. 4. Detection of the climate change (forced) signal in significant wave heights (Hs) and wind speed (U) illustrated for various ocean basins. Shadow represents unforced variability calculated from the 100-yr control simulation. Lines are linear trends over detection time for the historical period of 1960–2010 (blue line) and the future scenario RCP8.5 from 2010 to 2060 (red line).
reaching for the global wave climate if these regions would be exporting long-traveling swell to other seas. In such a case, dominant trends in wind speeds, as well as changes in wind and wind sea wave direction would result in a changing swell climate and would affect the neighboring seas. As shown by Dobrynin et al. (2012), the negative anomalies of wind speed and significant wave height appear in a vast area of the global ocean.

We applied our detection technique for every time series at each grid point of the global model domain in both directions in order to detect whether and when the positive and negative threshold values of natural variability are exceeded. Detection time was calculated as the first crossing of the upper or the lower limit of natural variability by forced trends for the significant wave height, swell height, and wind speed (see section 2c). The detection time was estimated for every grid cell and then averaged zonally for certain periods to demonstrate the global progress of climate change. This approach results in a detection map (Fig. 5), which defines regions where the climate change signal is detectable in the near future in the RCP8.5 scenario.

Overall, the climate change signal is detectable for the significant wave height in the next 5 decades in 60% of the area of the global ocean. Propagation of the climate change signal in time can be illustrated by the zonal mean area fraction in which positive and/or negative signals emerge (Fig. 5a). In the Arctic and Southern Ocean, a similar increase from 20% up to more than 50% of the area where the change in significant wave height becomes detectable from one decade to another can be seen. Note that only a very small fraction of the Arctic Ocean could be considered in our detection method (see section 2c). This indicates a continuously changing wave climate in these regions. For other regions, such as between 30° and 40°N in the Pacific and Atlantic Oceans, the changes in significant wave height become detectable in the first and second decades for 25%–30% of the area, with no further increase of the area in the subsequent decades. Note that the only region where the changes become detectable for 100% of the area by the end of 2060 is the Southern Ocean. For the rest of the global ocean, the climate change signal may increase in the following decades, and the area where changes become detectable can reach 100% as well. However, this is not captured in our detection analysis, because the maximal detection period allowable from our approach is 50 yr (see section 2c).

We analyzed three types of detectable climate change signals: (i) when trends exceed the positive limit of natural variability, (ii) when they exceed the negative limit of natural variability, and (iii) reversible signals (i.e., when trends exceed first one and then the other limit at a different point in time). In the following, we discuss these three types of detection in details.

a. Type I: Trends exceed the positive limit of natural variability

In a high-CO2 world in which the global mean significant wave height and wind speed are projected to increase (Fig. 1), this is the most common type of detection—when a positive trend for any specific trend length becomes strong enough and crosses the positive limit of natural variability. In other words, in a forced simulation for the future scenario RCP8.5, a positive trend can be found that was not found in the control simulation under preindustrial climate conditions. In major parts of the global ocean (35%, Fig. 5b, red part of the color scale), the positive limit of natural variability is exceeded, indicating intensification of positive trends in significant wave height. Increasing local wind speed in combination with additional locally or remotely generated swell leads to an increase of significant wave height. This type of climate change signal is already detectable in the equatorial Pacific and Indian Oceans, as well as in the East China Sea in the next decade (2010–20). Starting from the second decade, the signal becomes more pronounced in the southeastern and equatorial Atlantic and in the Southern Ocean. In the following two decades, until 2060, the climate change signal is detectable in the entire Southern Ocean and major areas of the equatorial Pacific and Atlantic. It is notable that in the regions of high variability, such as the North Pacific and Atlantic, the climate change signal is hardly detectable over the analyzed period.

b. Type II: Trends exceed the negative limit of natural variability

Intensification of negative trends is detectable when the negative limit of natural variability is exceeded. In 25% (Fig. 5b, blue part of the color scale) of the global ocean, the negative trends are significantly different from the control simulation, and the climate change signal is detectable in the next decades. In parts of the North and equatorial Atlantic, Arctic, and Southern Oceans, the climate change signal is already detectable in the next two decades. Later, in the third and fourth decades, this type of climate change signal becomes detectable in the equatorial Pacific, Indian, and northeast Atlantic Oceans. This type of detection means that the decrease in significant wave height is more rapid than occurs in the control simulation.

c. Type III: Reversibility of climate change signals

According to our simulations, we found areas of the global ocean where the detection type is reversible. In
3% of the ocean area (Fig. 5b, dotted area), type I of the detection was found in the first decade (cf. Fig. 5b, dotted area, and Fig. 5a), and later, in the next decades, the type of detection was reversed to type II (negative). In this case, a positive trend on the short-term scale is a part of longer negative trends that dominate at the end of the detection period. Reversibility may indicate also that oscillations longer than several decades are present in the wave climate; this needs to be investigated additionally over longer detection periods (i.e., longer than 50 yr).

In summary, the detection analysis indicates that, on the scale of ocean basins, a forced climate change signal is detectable for significant wave height (Fig. 5) in the South Atlantic and the Southern Ocean in the next 40–45 yr, where it goes out of the natural variability range (Fig. 4, red lines). Wind speed does not show the same behavior as waves, and the climate change signal is not...
detectable at all when values averaged over the whole ocean basin are considered. This outlines the strongly pronounced local character of the trends in the wind and wave fields, which would be filtered out by averaging over larger domains. Therefore, our results suggest that on the global scale a forced climate change signal in wave climate is not detectable: neither in the historical period from 1960 to 2010, nor in the future scenario RCP8.5 for the next decades (Fig. 3, red and blue lines are not crossing the edges of the shadowed area).

5. Toward attribution of changes in wave climate

Changes in the atmospheric circulation lead to a redistribution of wind patterns and wave regimes. It is quite evident that an increase in local wind speed causes an enhancement of wind wave generation. But in regions like the South Pacific, Atlantic, and Indian Oceans, the increase of significant wave height is not related to the increase of the local wind. In fact, in these regions the wind speed is decreasing, but additional swell coming from the Southern Ocean increases the significant wave height (Dobrynin et al. 2012). Wave systems in most areas of the global ocean are generally a superposition of wind sea and swell. Therefore, we attributed the changes in wave climate to two major factors: variations in the local wind and the impact of wind on swell generation and propagation.

To illustrate the impacts of these two factors, we calculated the anomalies of climatological seasonal mean values of significant wave and swell heights over the period 2080–2100 in the RCP8.5 scenario to the reference period 1850–70 (Fig. 2, lower panel). For significant wave height, the area of the global ocean covered by positive anomalies is 48.5%, which is only slightly smaller than the area covered by negative anomalies (51.5%). For swell height, this distribution is reversed, being 52.1% for positive anomalies and 47.9% for negative. This is a consequence of the differences in seasonal storm patterns between the Northern and Southern Hemispheres: the different directions of the winds and the wind-generated waves and topography. Oceans in the Southern Hemisphere are open to swell generated in the Southern Ocean. In the Northern Hemisphere, wind with a dominant westerly direction generates swell, but the continental configuration blocks swell propagation to other regions. The increase of swell generation in the Southern Ocean and its propagation toward the north leads to an increase of significant wave height in the austral winter period (JJA) in most areas of the Indian, South and equatorial Pacific, and Atlantic.

These changes in swell distribution can have two main sources related to the wind conditions in the regions of swell origin and on the pathways of swell. First, an increase in significant swell height is in line with higher wind speed in the swell-generating regions. Second, changes in swell direction are due to shifts in wind directions. The dominant swell direction is calculated as the most frequent direction within one of the 16 directional sectors (i.e., starting from north with a step of 22.5°). Changes of the pathways of swell propagated from the Southern Ocean are notable already in the historical period (not shown). More pronounced shifts of swell pathways are projected in the future RCP8.5 scenario (Fig. 6d).

The wave energy is a function of wave height $E = \frac{1}{16\rho g H^2}$ ($\rho$ is water density, $g$ is gravity acceleration, and $H$ is wave height), and, because of wave energy conservation, the total energy can be written as the sum of swell and wind sea contributions:

$$H^2 = SH^2 + WH^2,$$

where $H$ is the significant wave height, $SH$ is significant swell wave height, and $WH$ is significant wave sea height. We defined three periods for the analysis of anomalies of swell contribution [calculated as $100(SH^2/H^2)$] and compared these to the reference period 1850–70 with preindustrial climate state. In the first period, from 1950 to 1970 (Fig. 6a), the changes in swell distribution become notable in different regions. In the equatorial Pacific and Indian Ocean, and partly in the Southern Ocean, a clear negative tendency shows that, because of an increase of local wind, the wave spectrum is shifted toward wind sea waves. During the second period, at the end of the historical simulation (1990–2010, present climate state, Fig. 6b), the negative anomalies in the Southern Ocean are more pronounced. In the Southern Ocean, the contribution of swell is decreasing in line with the pathways of stronger winds forming a positive anomaly of swell contribution in the regions opened to the Southern Ocean swell. In the future period (2080–2100, RCP8.5 scenario, Fig. 6c), the wind sea becomes dominant in the Southern Ocean. More swell, corresponding to up to a 5% increase of wave energy, is projected in the equatorial regions, the North Atlantic, and the eastern North Pacific.

Additional estimated detection times for the wind speed (Fig. 5f) and swell height (Fig. 5d) indicate regions where local or remote factors dominate the wave climate change. This can be estimated by identifying and comparing similar patterns in the detection map for significant wave height (Fig. 5b) and the detection maps for significant swell height and wind speed (Figs. 5d,f). For instance, typical swell-induced changes can be seen in the Gulf of Guinea, where the changes are detectable...
in the significant wave (Fig. 5b) and swell heights (Fig. 5d), but not in the wind speed (Fig. 5f). Following this approach, we attributed changes in the wave climate to be swell-induced on the west coast of North and South America between 40°N and 10°S, the west coast of South America between 30° and 50°S, and in the area around the southern end of Africa in both the South Atlantic and Indian Oceans. The equatorial Pacific from 165°E
toward the coast of America shows only wind-induced changes. Here, the climate change signal is detectable in significant wave height (Fig. 5b) and wind speed (Fig. 5f), but not in the significant swell height (Fig. 5d). The China Sea, Hudson Bay, and Gulf of Alaska are other regions where wave climate changes are dominated by the local changes in the wind speed. However, most of the global ocean has a mixed effect from both wind changes and swell changes. In the North Atlantic, the negative signal in the wave climate is detectable (Fig. 5b), but it cannot be clearly attributed to wind or swell changes. The Southern Ocean and Arctic are other regions with mixed effects of wind speed and swell.

6. Concluding remarks

The goal of this study was to detect the forced external variations in wave climate caused by climate change and to compare them to the natural (unforced) variability. We estimated the natural variability in significant wave and swell heights, and wind speed using a control simulation of the Earth system model EC-EARTH and a wave model WAM under preindustrial climate conditions. Statistically significant changes were analyzed in simulations spanning the historical period and a future climate change scenario RCP8.5. Using a simple fingerprint approach, we found that, in our experiments, the present wave climate is already affected by climate change in terms of significant wave and swell heights and wind speed. We found detectable climate change signals in the current decade (2010–20) in the North Atlantic, equatorial Pacific, and Southern Oceans. Projections for the near future in the RCP8.5 scenario show an extension of climate change signal to the new regions. For the detection period of 50 yr to the end of 2060, 60% of the global ocean area is affected by climate change. In our analysis, both negative and positive limits of natural variability in waves and wind were exceeded, indicating reversibility of the climate change signal, and leading to positive and negative types of detection. Therefore, ongoing climate change acts not only to increase significant wave height and wind speed, but also to significantly decrease them regionally. For example, in the North Atlantic and South Pacific (between 140°E and 130°W), continuously decreasing linear trends (negative detection) in the significant wave height are projected in the future RCP8.5 scenario. Positive detection dominates over the remaining global ocean areas, except for the 3% of the area where a reversibility of the climate change signal is detectable (equatorial Pacific and partly in the South Pacific and Southern Oceans).

Detectable changes in the wave climate can be attributed to a combination of changes due to local and remote components. Locally induced changes are accompanied by an increase of in situ wind speed (e.g., in the Arctic Ocean and in the Southern Ocean between 40° and 60° S). Our results highlight the role of the Southern Ocean as a major source of swell and thereby as a generator of wave climate change signal in adjacent regions. Swell generated in the Southern Ocean propagates northward, contributing to changes in significant wave height in the Indian, South Atlantic, and Pacific Oceans, where changes in wind and waves are decoupled. The pathways of swell are affected by climate change as well, indicated by shifts in the energy distribution over the full wave spectrum and changing wave systems.

Note that, because of methodology constraints (see section 2), parts of the Arctic and Southern Ocean that were covered by sea ice in the control simulation were not included in the detection analysis. Yet vast areas in the Arctic and Southern Ocean, where sea ice is projected to retreat in the future, show pronounced positive anomalies in significant wave and swell heights. Although these anomalies are caused by climate change, this is not captured in our detection method. Furthermore, our results have been produced by one model only. The method used here is model dependent, and hence detection time can differ in another model simulation. A multimodel detection analysis would be a straightforward extension of this study.

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