^aInfluence of Climate Variability on Extreme Ocean Surface Wave Heights Assessed from ERA-Interim and ERA-20C

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ABSTRACT

Extreme ocean surface wave heights significantly affect coastal structures and offshore activities and impact many vulnerable populations of low-lying islands. Therefore, better understanding of ocean wave height variability plays an important role in potentially reducing risk in such regions. In this study, global impacts of natural climate variability such as El Niño-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), and Pacific decadal oscillation (PDO) on extreme significant wave height (SWH) are analyzed using ERA-Interim (1980-2014) and ECMWF twentieth-century reanalysis (ERA-20C; 1952-2010) datasets for December-February (DJF). The nonstationary generalized extreme value (GEV) analysis is used to determine the influence of natural climate variability on DJF maxima of SWH (Hmax), wind speed (Wmax), and mean sea level pressure gradient amplitude (Gmax). The major ENSO influence on Hmax is found over the northeastern North Pacific (NP), with increases during El Niño and decreases during La Niña, and its counter responses are observed in coastal regions of the western NP, which are consistently observed in both Wmax and Gmax responses. The Hmax response to the PDO occurs over similar regions in the NP as those associated with ENSO but with much weaker amplitude. Composite analysis of different ENSO and PDO phase combinations reveals stronger (weaker) influences when both variability modes are of the same (opposite) phase. Furthermore, significant NAO influence on Hmax, Wmax, and Gmax is observed throughout Icelandic and Azores regions in relation to changes in atmospheric circulation patterns. Overall, the response of extreme SWH to natural climate variability modes is consistent with seasonal mean responses.

1. Introduction

Extreme ocean surface waves can have a substantial contribution to coastal flooding, the destruction of offshore structures (such as harbors), and coastal sea level

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extremes. Typically, extreme (ocean surface) wave heights are associated with high surface wind speeds and generated from storms (Young 1999; Wang et al. 2010; Vinoth and Young 2011; Young et al. 2012). For example, the tropical coastal regions of East Asia and U.S. coastlines are severely affected by several typhoons and hurricanes, which generate extreme wave heights (Bromirski and Kossin 2008; Hemer et al. 2011; Park et al. 2014). Ocean surface waves are an important part of the climate system (Cavaleri et al. 2012; Babanin et al. 2012). Recent studies suggest that wave heights could be affected by anthropogenic forcing. For example, satellite observations showed increasing global trends of

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surface wind speed and wave heights during the most recent decades (Young et al. 2011, 2012). Additionally, Wang et al. (2009) demonstrated a clear detection of external influence on Northern Hemisphere wave heights and atmospheric storminess, using data for the 1955–2004 and 1958–2001 periods.

The majority of studies investigating historical trends are based on dynamical reanalyses of ocean surface wave heights over the past half century (Wang and Swail 2001; Wang et al. 2010; Dodet et al. 2010; Semedo et al. 2011). However, historical increasing trends in wave heights have also been shown to occur over much longer time scales, such as in the East China Sea from the early twentieth century (Wu et al. 2014). Global wave height trends during the late twentieth century (i.e., 1958–97) were analyzed by Cox and Swail (2001), while at the same time, Wang and Swail (2001) identified significant trends in seasonal extremes of wave height over the North Pacific (NP) and North Atlantic (NA). More recently, several studies report that such significant increasing trends are continuing (Caires et al. 2006; Wang et al. 2010; Semedo et al. 2011; Wang et al. 2012; Bertin et al. 2013). However, although extreme wave heights have generally increased over the northeastern NA, there has been a decrease in the midlatitudes (Wang et al. 2012). These historical trends in ocean wave heights are basically projected to continue during the twenty-first century with reorganized patterns of changes depending on regions and seasons (e.g., Dobrynin et al. 2012; Semedo et al. 2013; Fan et al. 2013). Studies based on multimodel simulations participating in phases 3 and 5 of the Coupled Model Intercomparison Project (CMIP3 and CMIP5) suggest, on average, increasing trends in wave heights over the tropics and southern high latitudes (Hemer et al. 2013; Wang et al. 2014).

While many studies have focused on long-term trends in wave heights as described above, natural climate variability such as El Niño–Southern Oscillation (ENSO), the Pacific decadal oscillation (PDO), and the North Atlantic Oscillation (NAO) can also exert significant impacts on wave heights through associated large-scale atmospheric circulation patterns. However, there have been limited studies on the global impacts of ENSO and PDO on wave heights. The influence of ENSO on extreme wave heights is even more limited, particularly based on extreme value analysis, except for regional studies of ENSO's impact on the coastal shoreline of California (e.g., Seymour 1998; Storlazzi and Griggs 2000).

The NAO is a well-established dipolar mode of climate variability, where variation in mean sea level pressure (SLP) has an inverse relationship between the subtropical Atlantic (Azores) high and Icelandic or Arctic low (Wanner et al. 2001; Hurrell et al. 2003). Previous studies show that significant variations of wave height in the NA are closely associated with contemporaneous seasonal SLP variations, especially in winter (Kushnir et al. 1997; Bauer 2001; Wang and Swail 2001; Woolf et al. 2002; Semedo et al. 2015). Further, it is crucial to analyze the influence of climate variability on extreme wave heights, which exert stronger impacts than seasonal means. Recently, extreme value theory has been increasingly used to examine the influence of climate variability on extremes (e.g., Kharin and Zwiers 2005; Zhang et al. 2010; Min et al. 2013) considering the nonnormality of extreme variables, for which simple linear regression analysis, typically used for mean variables, cannot be utilized.

In this study, we analyze the interannual (ENSO and NAO) and decadal (PDO) influences on extreme wave heights during Northern Hemisphere winter [December-February (DJF)] when two dominant variabilities (ENSO and NAO) have strongest amplitudes. We apply a generalized extreme value (GEV) analysis to significant wave height (SWH) data taken from two reanalyses: ERA-Interim (Dee et al. 2011) for 1979/80-2013/14 (referred to as 1980-2014) and ECMWF twentiethcentury reanalysis (ERA-20C; Poli et al. 2013) for 1951/52-2009/10 (referred to as 1952-2010). Further, the spatial patterns of wave height responses are compared with (surface) wind speed and SLP gradient amplitude patterns to assess the mechanisms associated with wave height responses to climate variability. In addition, responses in seasonal extremes of SWH, wind speed, and SLP gradient amplitude are compared with corresponding responses in seasonal means. Composite analysis of mean and extreme SWHs for different combinations of ENSO and PDO phases is also carried out to assess the role of PDO in strengthening and weakening the ENSO influences.

The remainder of the paper is structured as follows. We describe the data used and methodology of the GEV analysis in section 2. The indices of natural climate variability (interannual and decadal) and related teleconnection patterns are presented in section 3. In section 4, the DJF climatology and variability patterns as well as influence of climate variability on wave heights are analyzed and compared with wind speed and SLP gradient patterns. Conclusions are given in section 5.

2. Data and methodology

a. Data

We consider the Northern Hemisphere winter season (DJF) in our analysis of extreme wave heights. We

derived seasonal means and maxima of SWH (Havg and Hmax, respectively) from 6-hourly SWH data taken from ERA-Interim (Dee et al. 2011) and ERA-20C (Poli et al. 2013). Similarly, DJF mean and maximum wind speeds (Wavg and Wmax, respectively) were calculated from 6-hourly wind speeds at the 10-m level taken from each reanalysis dataset (ERA-Interim and ERA-20C). DJF mean and maximum of the SLP gradient amplitude (Gavg and Gmax, respectively) were also calculated from the 6-hourly SLP data to examine the relationship of geostrophic wind forcing to surface wind speed and SWH (Wang et al. 2009).

All data from ERA-Interim and ERA-20C have a horizontal resolution of 1° longitude \times 1° latitude, which have been directly downloaded from the ECMWF website (http://apps.ecmwf.int/datasets/, accessed 15 November 2015). Note that the original resolutions are slightly different depending on variables and reanalyses: ERA-Interim has a resolution of approximately 110 km for SWH and approximately 80 km for wind speed and SLP (Dee et al. 2011), and ERA-20C has a resolution of approximately 125 km for all variables.

We also used monthly mean sea surface temperature (SST) from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al. 2003) and SLP from the Hadley Centre Sea Level Pressure dataset (HadSLP2r, near-real-time update of HadSLP2; Allan and Ansell 2006) to analyze the teleconnection patterns of selected climate modes (ENSO, NAO, and PDO) for 1952–2014, which covers both the ERA-Interim (1980– 2014) and ERA-20C (1952–2010) periods. Teleconnection patterns obtained from ERA-20C SST and SLP provide essentially the same results (not shown). Further, in composite analysis of ENSO and PDO, the monthly mean HadISST and HadSLP2 data were used for the ERA-20C period (1952–2010) for the comparison of mean SST and SLP responses with SWH patterns.

b. Analysis of GEV distribution

To analyze the influence of climate variability on extremes, the nonstationary GEV analysis is carried out following previous studies (e.g., Kharin and Zwiers 2005; Zhang et al. 2010; Min et al. 2013). In this approach, it is assumed that extreme samples are well described by the GEV distribution, which is the limit distribution of block maxima of independent and identically distributed random variables (Coles 2001). Results from a goodness-of-fit test [based on a parametric bootstrap Kolmogorov-Smirnov test; refer to Kharin and Zwiers (2005) for details] show that the GEV distribution provides a reasonable fit to seasonal extremes of Hmax, Wmax, and Gmax at each grid point. Some exceptions are tropical ocean areas with too-small Gmax, which are excluded from our analysis (see below). The cumulative density function of the nonstationary GEV distribution is given as follows:

$$F(x,\mu_t,\sigma_t,\xi_t) = \begin{cases} \exp\left[-\exp\left(-\frac{x-\mu_t}{\sigma_t}\right)\right], & \xi_t = 0\\ \exp\left[-\left(1+\xi_t\frac{x-\mu_t}{\sigma_t}\right)^{-\xi_t^{-1}}\right], & \xi_t \neq 0, 1+\xi_t\frac{x-\mu_t}{\sigma_t} > 0, \end{cases}$$
(1)

where μ_t , σ_t , and ξ_t are the location, scale, and shape parameters, respectively. The location parameter represents the near-center position of the GEV distribution, corresponding to the mean of the normal distribution. The scale parameter indicates the spread or width of GEV distribution, which is equivalent to the standard deviation of the normal distribution. The shape parameter determines the types of the GEV distribution depending on its sign: $\xi_t = 0$, $\xi_t > 0$, and $\xi_t < 0$ represent Gumbel, Fréchet (having a heavier upper tail), and Weibull (having a finite upper tail) distributions, respectively (Coles 2001; Kharin and Zwiers 2005).

Now the climate variability v_t (ENSO, NAO, or PDO indices, detrended and normalized), which varies with time *t*, can be used as a covariate of GEV parameters.

For instance, the location parameter μ_t is made as a function of climate variability such that

$$\mu_t = \mu_0 + \mu_1 (v_t - v_0), \qquad (2)$$

where μ_0 is the location parameter at time t_0 , and μ_1 is the regression coefficient denoting the relationship between climate variability and location parameter. This makes the GEV distribution shift left and right, experiencing the influence of climate variability. Similarly, scale and shape parameters (spreads and shapes of the distribution) can be made as a function of climate variability such as $\log \sigma_t = \log \sigma_0 + \sigma_1(v_t - v_0)$ and $\xi_t = \xi_0 + \xi_1(v_t - v_0)$, where σ_0 and ξ_0 are the scale and shape parameter value at time t_0 , and σ_1 and ξ_1 are corresponding regression coefficients.

To determine the statistical significance of the influence of climate variability on extremes, the likelihood ratio test is applied. The test compares the log likelihood for a nonstationary GEV model with the log likelihood for a corresponding stationary GEV model and assesses the significance of the ratio (Kharin and Zwiers 2005; Zhang et al. 2010). The likelihood ratio test can also be conducted for two nonstationary GEV models with different assumptions-for example, one model with varying location parameter only and the other model with both location and scale parameters varying with time responding to climate variability. Based on likelihood ratio tests, influences of climate variability on extremes of SWHs, wind speed, and SLP gradient amplitudes analyzed below are found to occur largely through location parameters (shift of the distributions) with negligible influences on scale and shape parameters. Therefore, results shown below are from the nonstationary GEV model based on Eq. (2). To estimate GEV parameters, the method of maximum likelihood is used following Kharin and Zwiers (2005). It should be noted that since impacts of climate variability modes are dominated by location parameters, changes in return values such as a 50-yr event will be approximately the same as changes in location parameter given fixed scale and shape parameters (Kharin and Zwiers 2005; Min et al. 2009).

3. Climate variability modes

a. Indices

We have obtained climate variability indices for ENSO, NAO, and PDO during 1952-2014 from data centers (see below for detailed sources) to compare our results with previous studies. ENSO is a dominant coupled ocean-atmosphere phenomenon occurring over the equatorial Pacific, which affects global climate variability on interannual time scales. There is high confidence that ENSO will remain the dominant mode of natural climate variability in the twenty-first century (Collins et al. 2010; Stevenson 2012). To analyze ENSO influence, the Niño-3.4 index (SST anomaly averaged over 5°S-5°N, 170°-120°W), calculated based on the Extended Reconstructed SST, version 4 (ERSST.v4; Huang et al. 2015), was obtained from the National Oceanic and Atmospheric Administration (NOAA)/ Climate Prediction Center (CPC) (http://www.cpc.ncep. noaa.gov/data/indices/, accessed 15 November 2015).

The NAO indicates the most prominent teleconnection pattern over the NA, generally strongest during the Northern Hemisphere winter (Hurrell 1995; Hurrell et al. 2003), which consists of north–south dipole SLP anomalies with one centered in Iceland and the other center in the central latitudes of the NA between 35° and 40°N. We use standardized 3-month running mean values of the NAO index, defined as the leading mode from rotated empirical orthogonal function (REOF) analysis of monthly mean 500-hPa height (Barnston and Livezey 1987), obtained from the NOAA CPC (http://www.cpc.ncep.noaa.gov/data/teledoc/nao.shtml, accessed 15 November 2015).

A PDO index, derived as the leading mode of monthly SST anomalies in the NP, poleward of 20°N, is obtained from the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) at the University of Washington (http://research.jisao.washington.edu/pdo/, accessed 15 November 2015; Zhang et al. 1997). Monthly mean global average SST anomalies are removed to isolate the PDO pattern of variability from any global warming signal that may be present in the data. The index is constructed using the Met Office (UKMO) historical SST dataset for 1900–81 and Optimum Interpolation SST version 1 for January 1982–December 2001 and version 2 for January 2002–present (Reynolds et al. 2002).

When interpreting the impacts from individual modes of climate variability, we need to consider that they can interact with each other (e.g., Cai et al. 2011; Min et al. 2013). It is suggested that the PDO can exacerbate or mitigate the impacts of ENSO according to its phase (Yu and Zwiers 2007). Therefore, when both ENSO and PDO are in phase, El Niño or La Niña impacts may be intensified. If ENSO and PDO are of opposite phases, they are likely to cancel each other. However, we still need to investigate the interrelation of climate variability modes, or which climate mode has a stronger or weaker influence than others. To elucidate interactions between climate variability modes, correlation coefficients between DJF mean detrended ENSO, PDO, and NAO indices are calculated for 1952-2014 (Fig. 1, black lines). A significant positive correlation is found between ENSO and PDO (r = 0.49, p < 0.01), while only weak nonsignificant correlations are found between ENSO and NAO (r = -0.11) and NAO and PDO (r = -0.17).

To separate independent impacts due to each climate variability mode, we remove the dependency of one index on another by using a simple linear regression scheme. For example, removing the ENSO influence from PDO is attained by first regressing the PDO index onto the ENSO index and then removing the regressed component from the original PDO index. The result is a residual PDO index that is linearly independent of ENSO. Figure 1 shows detrended and normalized DJF mean time series for ENSO (Niño-3.4), ENSO independent of PDO (denoted as ENSO|_{PDO}; Fig. 1a), NAO (Fig. 1b), PDO, and PDO independent of ENSO (denoted as PDO|_{ENSO}; Fig. 1c). When comparing



FIG. 1. Detrended and normalized DJF mean time series for (a) ENSO (Niño-3.4) and $ENSO|_{PDO}$, (b) NAO, and (c) PDO and $PDO|_{ENSO}$ for the period of 1952–2014. Dashed horizontal lines show the plus or minus one-half standard deviation threshold used to define significant event years.

original (black lines) with residual indices (red lines in Figs. 1a,c), it is evident that interannual variations of ENSO are affected by decadal fluctuations of PDO, and vice versa. We use the independent residual indices below to isolate influences of ENSO and PDO from each other. To complement this approach, composite analysis is carried out based on the original indices and results are compared with those based on residual indices.

b. Mean teleconnection patterns

Mean teleconnection patterns of climate variability modes are first examined using independent indices (ENSO|_{PDO}, NAO, and PDO|_{ENSO}) before analyzing responses of ocean surface wave heights. Linear regression patterns of the mean SST and SLP onto climate variability modes are examined for the period 1952– 2014 DJFs (Fig. 2). As expected, ENSO is the mode of interannual climate variability (Figs. 2a,b), with its origin in the tropical Pacific Ocean and atmosphere, but its teleconnection reaches far beyond the tropical Pacific (Neelin et al. 1998; Hu et al. 2012). Significant signals in SST are observed over the central to equatorial eastern Pacific and tropical Indian Oceans (warmer conditions during El Niño) and over the tropical western Pacific



FIG. 2. Linear regression patterns of (left) SST [°C (std dev)⁻¹] and (right) SLP mean [Pa (std dev)⁻¹] onto (a),(b) ENSO|_{PDO}, (c),(d) NAO, and (e),(f) PDO|_{ENSO} indices for 1952–2014 DJF seasons. The hatched area represents grid points with statistically significant regression coefficients at the 5% level. All climate variability indices are detrended and normalized prior to analysis, and ENSO|_{PDO} and PDO|_{ENSO} represent the results using independent residual indices (see Figs. 1a,c).

(colder conditions). Further, the ENSO influence on SLP (Fig. 2b) is mostly significant in the northeastern Pacific (anomalous low pressure during El Niño) and tropical western Pacific and Indian Oceans (anomalous high pressure). Thus, anomalous high or low pressure systems can generate strong wind speed, leading to extreme ocean surface wave heights.

The NAO has the greatest influence on SSTs in the NA region (Fig. 2c) with a tripolar pattern of cold SST

anomalies in the subpolar NA, a warm anomaly in the west Atlantic between 20° and 45°N, and a cold anomaly between the equator and 30°N in the east Atlantic during its positive phase (Grossmann and Klotzbach 2009). A corresponding dipole response of SLP appears in the NA with an anomalous low centered in Iceland and an anomalous high over 35°–40°N (Fig. 2d). The PDO influence is associated with warmer SST over the northeastern NP and colder SST over the central to



FIG. 3. Climatology patterns of DJF (top) Havg, Wavg, and Gavg and (bottom) Hmax, Wmax, and Gmax for ERA-Interim (1980–2014) and ERA-20C (1952–2010). Note that the location parameter μ_0 from the stationary GEV fit is given for seasonal extremes.

northwestern NP and anomalous low pressure over the northern NP during its positive phase (Figs. 2e,f).

4. Influence of climate variability on ocean surface wave height

a. Climatology and variability

The global patterns of DJF seasonal mean and maximum SWH, wind speed, and SLP gradient amplitude from ERA-Interim (1980–2014) and ERA-20C (1952–2010) are displayed in Fig. 3. Location parameters of GEV distribution are shown as mean intensity of extremes. Mean climatology patterns of Havg and Wavg are broadly consistent with previous studies (Young 1999). Large Hmax values are evident over the NP and NA and the Southern Ocean, reflecting locations of active storm tracks. In the northern NA, the climatological mean of DJF extreme SWH is larger than 8 m (Hmax > 8 m), and it is slightly less (~6 m) in the high-latitude NP regions. The spatial correlation coefficients between DJF mean patterns of the three fields are high (0.88, 0.83, and 0.76)

for Havg and Wavg, Wavg and Gavg, and Gavg and Havg, respectively). In the case of extremes, the spatial correlations between variable fields are even higher (0.95, 0.94, and 0.90 for Hmax and Wmax, Wmax and Gmax, and Gmax and Hmax, respectively). This indicates close association between wave height, wind speed, and SLP gradient in terms of both winter mean and extremes. There is also strong agreement between climatology patterns of seasonal means and extremes, with spatial correlations between Havg and Hmax, Wavg and Wmax, and Gavg and Gmax of 0.93, 0.89, and 0.96, respectively. Therefore, extreme patterns of SWH, wind speed, and SLP gradient are closely associated with the mean pattern. ERA-Interim and ERA-20C show overall consistent patterns but with slightly stronger amplitudes in ERA-Interim than in ERA-20C over the central NP, NA, and Southern Ocean. This is mostly due to the different periods used between ERA-Interim and ERA-20C results. Almost the same patterns are seen when applying the common recent period of 1980-2010 (not shown).

The global patterns of interannual variability for DJF mean and extreme SWH, wind speed, and SLP gradient amplitude are displayed for ERA-Interim and ERA-20C (Fig. 4). In Gmax, we exclude regions with weak SLP gradients less than 1.5 Pa km⁻¹ as the GEV distribution does not properly fit for too-low values. Interannual variability of Hmax (scale parameter of GEV distribution; see above) is strongest over NP, NA, and Southern Ocean regions while weak in tropical regions for both reanalyses. The interannual variability of Wmax and Gmax shows consistent patterns with Hmax variability, with larger variability over NP and NA regions. This suggests that interannual variation of ocean surface wave height is intrinsically linked with variability in wind speed energy, which in turn is a response to changes in SLP gradient amplitudes. Spatial patterns of the interannual variability are overall consistent between DJF means and extremes, again indicating similar mechanisms working for both. The difference in interannual variability arising from two reanalysis datasets is due mainly to the different analysis period (not shown) as mentioned above for the climatology patterns.

b. ENSO and PDO influence

The spatial patterns of regression coefficients for mean and extreme SWH (Havg and Hmax), wind speed (Wavg and Wmax), and SLP gradient amplitude (Gavg and Gmax) regressed onto ENSO|_{PDO} are shown in Fig. 5. The mean responses of Havg, Wavg, and Gavg are based on simple linear regression, while extreme responses of Hmax, Wmax, and Gmax are obtained by using GEV analysis, where ENSO|_{PDO} index is used as the covariate. The ENSO|_{PDO} influence on Havg is statistically significant (5% level) mainly over the central equatorial Pacific region (positive) and the Maritime Continent (negative). Statistically significant responses of Wavg and Gavg are found over the northeastern NP, as well as along the intertropical convergence zone (ITCZ) and the South Pacific convergence zone (SPCZ). The mean response of ENSO|_{PDO} for ERA-Interim is consistent with ERA-20C responses, except that the Havg response to ENSO|_{PDO} for ERA-Interim is somewhat stronger than the ERA-20C results probably because of increased ENSO intensity in the ERA-Interim period (see above).

Similarly, the ENSO|PDO influence on Hmax is evident over the northeastern Pacific and along the ITCZ and SPCZ. The increase in Hmax along the SPCZ region seems to be related to enhanced tropical cyclone activity during El Niño during DJF (Kuleshov et al. 2008; Vincent et al. 2011). The Wmax response patterns to ENSO_{PDO} are largely consistent with the Hmax response in the SPCZ and ITCZ regions, indicating that increased extreme wave height is indeed induced by increased wind speed energy. There is also significant spatial correlation between means and extremes for SWH (r = 0.59), wind speed (r = 0.42), and SLP gradient (r = 0.42), indicating the strong association between extreme and mean responses to ENSO. The intensity of extreme responses is, however, approximately 4 times stronger than the corresponding mean response. Overall, the mean and extreme responses to $ENSO|_{PDO}$ for ERA-Interim are consistent with those for ERA-20C.

Figure 6 shows the PDO|_{ENSO} (PDO independent of ENSO) influence on SWH, wind speed, and SLP gradient amplitude. There appears to be significant influence of PDO|_{ENSO} on Hmax over the central NP from the longer period (ERA-20C), where consistent responses occur for Wmax and Gmax. The spatial patterns of PDO|_{ENSO} influence on mean responses (Havg, Wavg, and Gavg) are similar to the case of extreme responses with areas with statistical significance seen over the central NP in the ERA-20C results. Generally, PDO|_{ENSO} appears to have a weaker influence than ENSO|_{PDO}, especially in the central NP and NA. PDO|_{ENSO} responses for ERA-20C display a clearer pattern than those of ERA-Interim, which seems to be partly due to its longer period.

c. Composite analysis of ENSO and PDO

Individual influences of ENSO and PDO are analyzed above by applying regression analysis of independent variability modes to each other. To further examine combined influences of ENSO and PDO, we carry out composite analysis of SST, SLP, Havg, and Hmax



FIG. 4. Geographical distributions of interannual variability of DJF (top) Havg, Wavg, and Gavg and (bottom) Hmax, Wmax, and Gmax for ERA-Interim (1980–2014) and ERA-20C (1952–2010). Shape parameter σ_0 from the stationary GEV fit is given for extremes. The gray region indicates very small values for the SLP gradient ($\mu_0 < 1.5 \text{ Pa km}^{-1}$).

anomalies over the period 1952–2010 for different combinations of ENSO and PDO phases. First, we select years of El Niño (ENSO+), La Niña (ENSO-), and positive and negative phases of PDO (PDO+ and PDO-) from the detrended and normalized time series of original indices (Fig. 1, black lines). This results in 9, 3, 3, and 11 sample years for a combination of El Niño/ PDO+, La Niña/PDO+, El Niño/PDO- and La Niña/ PDO-, respectively, based on a threshold value of ± 0.5 for ENSO and PDO indices (Table 1). Figure 7 shows the composite patterns of SST and SLP anomalies for the four combinations of ENSO and PDO samples. The tropical Pacific SST exhibits typical anomalies for El Niño and La Niña events (Hoerling et al. 1997; Yeh et al. 2014). For El Niño/PDO+ events, SST warming becomes stronger in the central and eastern Pacific region, while La Niña/PDO- shows stronger cooling, indicating a strengthening of ENSO impact when PDO has the same sign of phase as ENSO. When ENSO and PDO are of opposite phase, the patterns are characterized by weakening



FIG. 5. ENSO influence on (top) Havg, Wavg, and Gavg and (bottom) Hmax, Wmax, and Gmax for ERA-Interim (1980–2014) and ERA-20C (1952–2010), based on ENSO|PDO. Regression coefficients of seasonal means are obtained from simple linear regression while those of extremes are estimated from GEV analysis. The regions statistically significant at the 5% level are hatched with black lines.

of the ENSO impact mainly over the eastern equatorial Pacific and with opposite SST anomalies in the NP to when the two climate variability modes are in phase. In the Atlantic and Indian Oceans, SST anomalies are generally of the same sign during El Niño events regardless of PDO phase but differ during La Niña events depending on PDO phase. For SLP, similar to SST, anomalies become stronger in the NP for El Niño/PDO+ (low pressure) and La Niña/PDO- (high pressure) event years as compared to years with opposite phases (Fig. 7, bottom). This confirms that the PDO plays a significant role in intensifying or reducing ENSO influences on SST and SLP, as previously reported (Yu and Zwiers 2007).

Composite analysis of mean and extreme SWH anomalies during different ENSO and PDO phase combinations is given in Fig. 8. Generally, Havg increases during El Niño/PDO+ event years and decreases during La Niña/PDO- events over the tropical Pacific Ocean. In the central NP, an El Niño (La Niña)



FIG. 6. As in Fig. 5, but for PDO|_{ENSO}.

associated with a positive (negative) phase of PDO leads to significant increases (decreases) of Havg and Hmax anomalies. However, the La Niña/PDO+ composite depicts a weaker pattern than the El Niño/PDO+ composite, and similarly, the El Niño/PDO- composite exhibits a somewhat weaker pattern than the La Niña/ PDO- composite. The impact of El Niño/PDO+ on extreme SWH covers a larger area in the NP compared to mean responses. This suggests that during positive phases of ENSO and PDO, the central NP experiences increases in Hmax, while La Niña and PDO- reduce Hmax in this region. The influence of opposite phases of ENSO and PDO on mean and extreme SWH will depend on the intensity of ENSO and PDO phases.

d. NAO influence

The NAO accounts for much of the interannual and longer-term variability evident in the Northern Hemisphere. During DJF, for instance, the NAO accounts for more than one-third of the total variance in SLP over the NA (Hurrell et al. 2003). Here, we examine NAO influence on mean and extreme SWH, wind speed, and SLP gradient amplitude for ERA-Interim and ERA-20C (Fig. 9). Regions that show significant influence of

TABLE 1. Classification of years when a combination of El Niño or La Niña and/or positive or negative PDO greater than plus or minus one-half standard deviation occurred during the period 1952–2010 (based on detrended time series; black lines in Fig. 1). The years represent years of January–February of the Northern Hemisphere winter season (DJF).

	El Niño (ENSO+)	La Niña (ENSO–)
PDO+ PDO-	1958, 1977, 1978, 1980, 1983, 1987, 1988, 1998, and 2003 1969, 1991, and 1995	1985, 1996, and 2006 1956, 1965, 1971, 1972, 1974, 1976, 1989, 1999, 2000, 2008, and 2009

NAO cover large parts of the NA for all three variables. The NAO has a significant impact on Havg and Hmax over the Icelandic region (increases during the positive phase of NAO) and Azores region (decreases during the positive phase of NAO), consistent with Wang and Swail (2001). Significant responses to NAO can also be seen in Wavg and Wmax, as well as Gavg and Gmax, over the same regions, highlighting the mechanism by which



FIG. 7. Composite patterns of (top) SST (°C) and (bottom) SLP (Pa) anomalies for the various combinations of ENSO and PDO—El Niño/PDO+, La Niña/PDO+, El Niño/PDO-, and La Niña/PDO- years (see Table 1)—for ERA-20C period of 1952–2010.



FIG. 8. As in Fig. 7, but for (top) Havg and (bottom) Hmax anomalies (m).

extreme SWH changes occur due to NAO—that is, significant variations in anomalous SLP patterns (a north–south shift of Atlantic storm tracks) resulting in changes to wind speed. Good agreement between ERA-Interim and ERA-20C results suggests robustness of the NAO impacts on wave heights and associated mechanisms.

5. Conclusions

In this study, we have examined the influence of natural climate variability modes on extreme SWH using ERA-Interim and ERA-20C datasets. ENSO, NAO, and PDO are analyzed for their influence on wave heights during the DJF season, and GEV analysis is used



FIG. 9. As in Fig. 5, but for NAO influence.

to identify areas with significant impacts associated with each climate variability mode. In addition, response patterns of wind speed and SLP gradient amplitude are used to highlight the mechanisms leading to such changes in wave heights. Overall, ENSO influence on extreme SWH occurs most strongly over the northern NP, while the strongest NAO influence appears over the Icelandic and Azores regions in the NA. Spatial regression patterns of extreme SWH, wind speed, and SLP gradient responses to ENSO and NAO are found to be very similar to seasonal mean response patterns. In contrast, the influence of decadal variability, such as PDO on extreme SWH, is shown to be much weaker than interannual variability associated with ENSO. However, composite analysis suggests that the PDO is able to enhance (reduce) the ENSO influence on mean and extreme SWH significantly when it has the same (opposite) sign of phase as ENSO, in support of previous studies on SST and SLP responses. Results are largely consistent between ERA-Interim and ERA-20C datasets.

Overall, we show that there is a close relationship between ocean surface wave height, wind speed, and SLP gradient amplitude in spatial patterns of climatology and interannual variability, as well as response patterns to climate variability modes for both seasonal mean and seasonal extremes during DJF. This clearly demonstrates that extreme wave height during the Northern Hemisphere winter is driven by increased wind speed energy, which is largely a result from increased SLP gradients over the northern extratropics, which can be induced mainly by ENSO and NAO, with small modifications by PDO. A boreal summer analysis (June–August) is being carried out and will be reported in a separate study, in which more importantly the influence of tropical cyclones on extreme wave events needs to be considered (Bromirski and Kossin 2008; Park et al. 2014).

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