Ocean circulation along the southern Chile transition region (38° −46°S): Mean, seasonal and interannual variability, with a focus on 2014–2016

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Abstract

Satellite and atmospheric model fields are used to describe the wind forcing, surface ocean circulation, temperature and chlorophyll-a pigment concentrations along the coast of southern Chile in the transition region between 38° and 46°S. Located inshore of the bifurcation of the eastward South Pacific Current into the equatorward Humboldt and the poleward Cape Horn Currents, the region also includes the Chiloé Inner Sea and the northern extent of the complex system of fjords, islands and canals that stretch south from near 42°S. The high resolution satellite data reveal that equatorward currents next to the coast extend as far south as 48°−51°S in spring-summer. They also display detailed distributions of forcing from wind stress and wind stress curl near the coast and within the Inner Sea. Between 38°−46°S, both winds and surface currents during 1993–2016 change directions seasonally from equatorward during summer upwelling to poleward during winter downwelling, with cooler SST and greater surface chlorophyll-a concentrations next to the coast during upwelling, opposite conditions during downwelling. Over interannual time scales during 1993–2016, there is a strong correlation between equatorial El Niño events and sea level and a moderate correlation with alongshore currents. Looking more closely at the 2014–2016 period, we find a marginal El Niño during 2014 and a strong El Niño during 2015 that connect the region to the tropics through the oceanic pathway, with some atmospheric connections through the phenomenon of atmospheric blocking (as noted by others).
The period also includes a Harmful Algal Bloom of the dinoflagellate *Alexandrium catenella* during early-2016 that occurred during a sequence of physical conditions (winds, currents and temperatures) that would favor such a bloom. The most anomalous physical condition during this specific bloom is an extreme case of atmospheric blocking that creates a long period of calm in austral autumn after strong upwelling in austral summer. The blocking is related to the 2015–2016 El Niño and an unusual coincident positive phase of the Southern Annular Mode.

**Keywords**

Eastern boundary currents; Wind-driven circulation; Seasonal variations; El Niño phenomena; Phytoplankton; Red tides

1. **Introduction and background**

   In this paper we describe the mean, seasonal and interannual variability in the wind forcing and ocean surface circulation next to the west coast of southern Chile in the transition zone between the Humboldt and Cape Horn Currents. We focus primarily on the coastal ocean from 38°S to 46°S during 1993–2016, with a closer look at anomalous conditions during 2014–2016. The latter period includes an ‘interrupted’ or ‘marginal’ El Niño during 2014, a strong equatorial El Niño in 2015–16 and a Harmful Algal Bloom (HAB) of the dinoflagellate *A. catenella* during the first half of 2016 that extended farther north than previously recorded. To our knowledge, the mean, seasonal and interannual variability in the wind forcing and circulation along southern Chile have not been well resolved in previous studies.

1.1. **Large-scale mean circulation**

   The classic description of large-scale mean surface circulation in the southeast Pacific Ocean is provided by Wooster and Reid (1963): the eastward South Pacific Current (SPC) bifurcates as it approaches southern Chile between 40°S and 50°S, flowing into the equatorward Humboldt Current to the north and the poleward Cape Horn Current to the south. Fig. 1 depicts these currents using a mean dynamic topography derived from satellite and *in situ* data (discussed in more detail below). Since that early description, a great deal of effort has clarified the physics, chemistry, biology, geology and fisheries oceanography of the Humboldt Current north of 40°S, where upwelling drives a productive ecosystem and fishery (See reviews in Montecino et al., 2006; Thiel et al., 2007; Montecino and Lange, 2009; Quiñones et al., 2010; Kämpf and Chapman, 2016; Several special volumes also focus on this region: Bertrand et al., 2008; Fréon et al., 2009; Escribano and Morales, 2012). South of 40°S, most of the physical, chemical and biological research has occurred in the channels of the complex fjord and island system of the Chilean Patagonia region (Sievers and Silva, 2008; Silva and Palma, 2008 and other papers in the same volume; Pantoja et al., 2011 and other papers in the same volume).

   Moving from north to south, the geometry of the coastline transitions from simple to more complex south of Chiloé Island. A region of relatively straight coast lies between 38°S and 41.5°S (Fig. 2), north of the area of fjords, islands and channels. South of the narrow
Chacao Channel along its north side, the western coast of Chiloé Island follows the nearly continuous coastline from farther north to approximately 43.5°S. South of Chiloé Island, between 43.5°S and 44.0°S, the wide channel known as the Boca del Guafo connects the open ocean to the large inner sea east of Chiloé Island. South of 44.0°S in the offshore Aysén Region, the “coast” consists of the outer boundaries of a multitude of small islands and archipelagos. The Chiloé Inner Sea (CIS, also called the Mar Interior de Chiloé) is 50–60 km wide and extends from approximately 41.5°S to 44.5°S.

Several descriptions of the coastal surface circulation off southern Chile differ in their details. Talley et al. (2011, their Figures 10.1 and 10.2) show the SPC to approach the continent near 43°S, with little information about the direction of the currents near the coast. A synthesis of surface drifter velocities, altimeter derived geostrophic currents and wind-driven Ekman transports by Maximenko et al. (2009) presents several variations of the large-scale circulation. Drifter movements (their Fig. 2a) appear to show poleward flow next to the coast south of 42°−44°S; combinations of the drifters, altimeter and Ekman flow, however, leave a blank next to the coast between approximately 40°S-48°S. Another view is presented by Strub et al. (2013), using the surface velocity fields from the global Ocean-GCM for the Earth Simulator (OFES) model (Ohfuchi et al., 2007). The long-term mean surface velocity fields from the model next to the coast appear to have an equatorward component as far south as 50°S and do not include a poleward Cape Horn Current anywhere between the coast and 90°W.

1.2. Large-scale seasonal variability

In the qualitative schematics of Strub et al. (1998), based on the temperature and salinity climatology of Levitus and Gelfeld (1992), the SPC deflects northward in winter, feeding into poleward currents that extend from 35°S to the south; in summer the offshore SPC dips south, connecting to equatorward coastal currents extending north from 42°S. The modeling study of Aguirre et al. (2014) provides information on the summer circulation as far south as 42°S, depicting an equatorward geostrophic coastal jet at the surface with speeds of 10–15 cm s⁻¹ and a width of 40–60 km between 37°S and 42°S. Their summer surface circulation fields do not extend south of 42°S along the east coast of Chiloé Island or within the CIS.

The large-scale seasonal variability of mean surface wind forcing along southern Chile is produced by the seasonal migrations of the offshore South Pacific High (SPH) pressure system and its frictional interactions with the continent. Synoptic variability is superimposed on the seasonally varying winds, caused by the eastward passage of cyclonic storms (Rahn and Garreaud, 2013) and poleward propagating coastal lows (Garreaud et al., 2002; Garreaud and Rutllant, 2003).

Using sea level (SLP) fields from the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) models, Rahn and Garreaud (2013) describe the seasonal movement of the SPH from ~100°W, 35°S in summer to ~90°W, 27°S in winter. For coastal winds, however, the important aspect of surface pressure is the latitude at which the maximum pressure occurs at the coast, which is south of the latitude of the offshore SPH maximum. Saavedra and Foppiano (1992) use a 30-year (1911–1940) time series of surface pressure observations along the coast to demonstrate the latitudinal
movement of this coastal point from 42.5°S in summer (January-February) to 35°S-36°S in winter (May-August). The general relationship is that alongshore winds are poleward (equatorward) south (north) of this region, resulting in the observation that alongshore winds between approximately 35°S-45°S reverse direction between summer (equatorward) and winter (poleward). More specifically, Sobarzo et al. (2007) and Rahn and Garreaud (2013) state that north of 37°S (Punta Lavapié), alongshore winds are equatorward, alternating in direction south of that latitude. Several authors place the maximum in equatorward winds between 35°–37°S in summer. Using scatterometer data, Aguirre et al. (2014) find the actual summer wind maximum at about 150 km offshore of the coast, stretching from 34°S to 38°S, just north of our primary domain of interest (Fig. 2). Due to the proximity of the Andes mountains to the coast, a geostrophic balance cannot be maintained and the coastal winds often take the form of a low-level ageostrophic equatorward jet, driven in part by an alongshore pressure gradient (Muñoz and Garreaud, 2005; Garreaud and Muñoz, 2005; Garreaud and Falvey, 2009; Belmadani et al., 2013; Rahn, 2012; Rahn and Garreaud, 2013). South of 37°S, the winds return to a more geostrophic balance.

Aguirre et al. (2014) also present the mean summer fields of sea surface temperature (SST), based on their model and satellite SST fields (2000–2007). A band of cool water lies next to the coast due to upwelling, along with a general cooling from north to south. From 35°S to 42°S, the front between the cool upwelled coastal water and the offshore warmer water extends southward, parallel to the coast and approximately 100 km from it. Near 42°S, at the northern end of Chiloé Island, the front turns offshore (westward) and runs east-west along the axis of the South Pacific Current, separating the warmer water in the north from the cold water in the south.

1.3. Large-scale interannual variability

Interannual variability in the ocean and in the atmospheric weather/climate system next to western South America is most commonly attributed to the effects of El Niño-Southern Oscillation (ENSO) events. As in the Northern Hemisphere, ENSO effects reach the west coast of South America through both oceanic and atmospheric pathways, with the oceanic pathway dominating at lower latitudes (Barber and Chavez, 1983; Fuenzalida, 1992) and the atmospheric pathway gaining importance in the subtropics (Rutllant and Fuenzalida, 1991; Garreaud, 2018). We note that in this paper our reference to ‘equatorical El Niño’ events includes both Central Pacific and Eastern Pacific El Niños, but not the more localized ‘coastal El Niños’ along the coasts of Ecuador and Perú as experienced in 2017 (Garreaud, 2018).

Equatorial Kelvin waves associated with El Niño events raise sea levels and depress pycnoclines next to the coast at low latitudes, with associated increases in SST and poleward currents. The poleward spread of high sea levels in the Northern Hemisphere of the Americas is well documented by the early tide-gauge studies of Enfield and Allen (1980) and Chelton and Davis (1982), with poleward speeds of 80–150 km/day, slower than first mode of coastal trapped waves (Enfield and Allen, 1983).

In the Southern Hemisphere off Ecuador and Perú, the 1976–1977 and 1982–1983 El Niño events are well documented. In 1982, a first pulse of high sea level occurred during
September-October, with a second pulse in January-February. The first pulse increased SST by 4°–6°C and raised sea levels by 30 cm, depressing mean pycnocline depths from 20 to 80 m. Coastal currents were anomalously poleward during the first two months of the 1982 event, similar to observations during the 1976 event (Smith, 1983; Huyer et al., 1987). Upwelling-favorable winds often continue or even increase during El Niño events at these lower latitudes (Halpern, 2002; Kessler, 2006), although Rahn (2012) finds a decrease in equatorward winds farther south between 30° and 40°S during El Niño conditions. Off Chile at 20°S, Fuenzalida (1992) documents SST anomalies in 1976–77, 1982–83, 1986–87 and 1991–92. Fonseca (1985) and Bilbao (1992) observe higher sea level signals during El Niño at 33°S, but not at 36°S.

During the 1997–98 event, Strub and James (2002a) use altimeter data to describe two pulses with different timing than 1982–83: a first pulse in May-June 1997 and a second in October-December 1997, both associated with poleward velocities and positive anomalies of sea level and SST, along with depressed isotherms (Blanco et al., 2002; Carr et al., 2002). The first pulse of high sea level moved quickly and coherently down the coast of South America to its tip (55°S) in May, aided by poleward autumn winds. The second pulse in October-December was opposed by spring-summer equatorward winds, especially at the higher latitudes. At mid-latitudes, anomalous conditions were quickly reduced during January-March 1998.

Moving to the atmosphere, the Intertropical Convergence Zone (ITCZ) drifts farther south during strong El Niño conditions next to the South American coast. Consistent with a negative Southern Oscillation Index, the subtropical anticyclone weakens, particularly in austral winter/spring, allowing the mid-latitude storm tracks to be displaced north (Rahn, 2012). At the same time there is an increase in the occurrence of blocking anticyclones west of the Antarctic Peninsula, (e.g. Rutllant and Fuenzalida, 1991, Garreaud, 2018). The displacement of the mid-latitude storm tracks toward the sub tropics and the weakening of the subtropical anticyclone generate a low pressure anomaly (L) between 30° and 40°S, contrasting with the warm-core, high-pressure system (H) that blocks the westerly flow at higher latitudes. This meridional dipole anomaly is a frequent configuration during El Niño periods (Renwick, 1998), related to the Pacific South America (PSA) atmospheric teleconnection pattern (Mo and Higgins, 1998). The dipole structure may also be increased in strength during positive phases of the Southern Annular Mode (Garreaud, 2018). At the nodal point between the two anomalies (near 45°S), easterly wind anomalies weaken the climatological westerlies, which are forced to flow around the dipole, producing enhanced westerly flow at 25°–30°S and 55°–60°S. The high pressure anomaly may also enhance equatorward winds along the coast between 40° and 50°S if the high moves west of the coast. In our analysis, we quantify the strength of blocking by forming a blocking index (BI) based on this conceptual model, as formulated by Ummenhofer et al. (2013).

1.4. The Inner sea, fjords and channels, physics and HABs

Within the CIS and the complex channels interlaced between the thousands of islands south of the CIS, there is limited information on the patterns of horizontal circulation. We restrict our discussion to the CIS east of Chiloé Island and the channels leading into the CIS
from the open ocean, primarily to the south of Chiloé Island (Fig. 2). The CIS consists of two ‘micro-basins.’ The northern microbasin is comprised of the Estero Reloncaví and two ‘gulfs’: Golfo Reloncaví and Golfo Ancud. The southern microbasin includes the Golfo Corcovado and the Boca del Guao channel (Palma et al., 2011; Martínez et al., 2015). The northern and southern basins of the CIS are interconnected by narrow passages between the Desertores Islands stretching across the middle of the CIS (Rodrigo, 2008). The location of the islands is indicated in Fig. 2, although the islands themselves are not displayed on the map.

No significant transport flows through the narrow and shallow Canal Chacao (Fig. 2) north of Chiloé Island (4 km wide, 50 m deep). The estuarine circulation (surface outflow over subsurface inflow) described by Sievers (2008) occurs primarily through the Boca del Guao south of the island (66 km across, 150 m deep, (Palma et al., 2011)). Fresh water enters the estuary from the boundaries through rivers, fjords and land runoff, with greater inflow in the northern basin, where it forms a 20–30 m surface layer (Iriarte et al., 2016). Vigorous tidal mixing homogenizes the water column in passages through the Desertores Islands in the middle of the CIS, transitioning to a deep upper mixed layer of fresher water in the southern basin that exits through the Boca del Guao over saltier inflow. More detailed descriptions of the vertical structure are provided from winter and spring cruises in 2004, 2005 and 2006 by Montecino and Pizarro (2008) and Palma et al. (2011), with a similar overall characterization.

Tides are of mixed semidiurnal type with a range of 1.5 m, propagating southward along the Chile coast and into the CIS through the Boca del Guao (Fierro, 2008). Tidal ranges increase to the north of the Boca Del Guao, reaching approximately 6 m at the mouth of the Estuario Reloncaví. A current meter in the Boca del Guao south of the southeast corner of Chiloé Island observed reversible surface currents to be strongly polarized in the east-west, along-channel direction, with maximum velocities of ~1.40 m s\(^{-1}\) (Castillo and Valenzuela, 2008).

The phytoplankton community abundance, size composition and north-south spatial structure in the CIS are influenced by lower light during winter and lower nitrate concentrations and freshwater inflow in the upper ocean in summer (Iriarte et al., 2007; Martínez et al., 2015; Iriarte et al., 2016). Euphotic zone depths of ~20 m are reported by Montecino and Pizarro (2008) in the fjord regions, with high correlations between the pigment concentrations at 10 m and the surface. Thus, satellite observations represent pigment concentrations in a surface layer similar to the depth of the estuarine surface layer of fresher water. Maps of satellite-derived CHL concentrations (Lara et al., 2016) characterize the CIS as a northern basin with high CHL and a southern basin with lower CHL, separated by the Desertores Islands near 42.6°S. They also show a band of higher CHL along the west coasts of Chiloé Island and the mainland to the north. The seasonal changes in the coastal band are expected to follow the pattern described by Yuras et al. (2005), who found a summer maximum next to the coast and a winter maximum farther offshore.
Harmful algal blooms (HABs) are economically important in the southern channels and CIS, where aquaculture of salmon and shellfish has become a major industry. Dinoflagellate species such as *Alexandrium catenella* (*A. catenella*) were found primarily between 53° and 55°S (Guzmán and Campodonico, 1975) in 1972 but expanded northward to 50°S in 1981 and to 43°–48°S by 1992. To better understand the spread of HAB species, field monitoring was initiated with cruises in 1994 by the Ministry of the Economy (Programa de manejo y monitoreo de las mareas rojas en las regiones de Los Lagos Aysén y Magallanes). These cruises documented the arrival of *A. catenella* within the CIS in 1998. The first observations of *A. catenella* along the western and northwest outer coast of Chiloé Island occurred during October-November 2009 (Mardones et al., 2010). During January 2016, however, *A. catenella* was reported for the first time north of Chiloé Island along the coast of Chile to 39°–40°S (Buschmann et al, 2016; León-Muñoz et al., 2018; Garreaud, 2018).

1.5. Questions to be addressed

Our analyses (below) make use of time series of satellite data (primarily) to show the spatial structure of fields over three time scales: (1) long-term means (zero frequency), (2) seasonal changes (roughly annual periods), and (3) interannual variability at scales mostly longer than a year. Higher-frequency intraseasonal variability is removed by filtering. We examine vector fields of surface wind stress and geostrophic currents, along with scalar fields of wind stress magnitude, wind stress curl, sea surface heights, temperatures and chlorophyll-a pigment concentrations. Although the different satellite data are not all available over the same time spans, the data sets with the longest records (reanalysis winds and altimeter fields) are used to compare seasonal cycles derived over the different time intervals. The resulting seasonal cycles are not sensitive to the use of the different record lengths.

The primary questions addressed in our analyses are:

1. What are the long-term means, seasonal progressions and interannual variability of forcing by wind stress (including wind stress curl) along the outer coast of Chile in the region between 38°S and 46°S?

2. What are the means, seasonal cycles and interannual variability in the geostrophic surface currents in this ‘transition region’ between the Humboldt and Cape Horn Currents? Do the alongshore currents follow the spatial patterns of the reversals of the seasonal alongshore winds? How do the coastal currents connect to the eastward SPC in the offshore open ocean and to coastal currents north and south of our regions of interest?

3. How do the seasonal cycles and interannual variability of sea level, surface temperature and chlorophyll-a concentrations relate to the wind forcing and surface current fields; how are they related to tropical ENSO signals?

4. What can these satellite data resolve about the patterns of seasonal wind forcing and response within the more restricted region of the CIS?

5. What atmospheric and oceanographic physical conditions occurred during the anomalous 2014–16 period? How do these relate to the equatorial El Niño events and to the extended HAB off southern Chile in the first half of 2016?
2. Data and methods

The baseline period considered covers the 24 years from 1993 through 2016 (see Table 1), during which data are available for altimeter sea level anomalies (SLA), and geostrophic velocities calculated from SLA ($V_g$). Over the same time span, we have coarse sea surface temperature (SST, 0.25° gridding) and coarse surface wind stress from reanalysis products ($\tau$, 0.67° gridding). Higher-resolution satellite data are available during shorter periods for $\tau$ and wind stress curl (WSC), SST and chlorophyll-a concentrations (CHL). Raw data with different sampling intervals are averaged to form monthly time series for consistent analyses and comparisons. Seasonal cycles are calculated from monthly calendar means during integral numbers of whole years. The HAB distributions of the dinoflagellate A. catenella are derived from available publications and reports.

2.1. Altimeter data

2.1.1. Gridded SLA and mean dynamic topography (MDT)—The “AVISO 2014 Delayed Time (DT14)” gridded (0.25°) SLA data, along with geostrophic velocity anomalies derived from the SLA, are obtained from the European Copernicus Marine Environment Monitoring Service (http://marine.copernicus.eu). In the final analyses we use monthly mean data from the “all altimeters” data set to obtain the highest spatial resolution. Absolute sea surface height (SSHA) fields are created from the SLA fields by adding the mean dynamic topography (MDT) field from CNES (CNES-13), obtained from the AVISO web site, https://www.aviso.altimetry.fr. The MDT field is constructed by CNES from a combination of satellite data (altimeter SLA and gravity) and in situ data (surface drifters and hydrographic profiles) (Rio et al., 2007). Geostrophic currents derived from gradients of the MDT fields are also added to the geostrophic velocity anomalies to form absolute geostrophic currents. Both SLA and MDT fields are available on a 0.25° grid. The altimeter data cover the base period of 1993–2016.

2.1.2. Alongtrack SLA from the reference missions—These data are obtained from the RADS (Radar Altimeter Database System, http://rads.tudelft.nl/) web site, maintained at TU Delft. We use the final research-quality alongtrack SLA data and various environmental ‘corrections’ as recommended to construct those data. Only the TOPEX and Jason reference missions are used, with ~10-day repeats. These data cover the 1993–2016 period with alongtrack resolution of ~6 km.

2.2. Satellite and reanalysis wind fields

2.2.1. European center for medium range weather forecasts (ECMWF) 10-m surface winds—Wind stress is calculated from the 6-hour ERA-Interim ECMWF model reanalysis surface winds (https://rda.ucar.edu) at approximately 2/3° resolution, then averaged into monthly means. We use a drag coefficient that depends on wind speed ($C_D = 10^{-3} \times ((2.7/U10) + 0.142 + 0.0764 \times U10)$, Large et al. (1995)), where $U10$ is the 10 m wind speed. The ECMWF winds are available for 1993–2016. During the strong anomalies of 2015 and 2016, we look at daily time series of the ECMWF surface 10-m winds. We note that ECMWF reanalysis methods assimilate the available scatterometer data from all
scatterometers throughout the 1993–2016 period and so should be a consistent substitute for the scatterometer fields, although coarser in resolution.

2.2.2. Scatterometer surface winds—QuikSCAT wind fields are available from all individual swaths for November 1999 through October 2009 at 12.5 km resolution at the NASA Jet Propulsion Laboratory (JPL) Physical Oceanography Distributed Active Archive Center (PODAAC) (https://podaac.jpl.nasa.gov). Surface wind stress is calculated in each swath using the same drag coefficient as for ECMWF winds. Three different versions of the JPL processing were examined in efforts to maximize the amount of data available close to land and within the CIS. This also maximizes the coverage of fields of WSC. Version 3.0 uses conservative processing, eliminating data within a fixed distance of land. This results in the greatest loss of data, retrieving no winds within the CIS. Version 3.1 retrieves more data next to land by allowing a certain percent of its “footprint” to lie over land (presently 1%). It retrieves data closer to the coast, including some data from within the CIS. QuikSCAT version 4.0 is similar to version 3.1 except that it estimates the reflectivity (sigma-0) of the land surfaces based on the full QuikSCAT data set. It then estimates the percent of the returned power that is reflected by land in the footprint and allows the retrieval if that percent is less than some amount (1% at present) [Bryan Stiles, pers. comm., May 2017]. Our tests confirm that version 4.0 retrieves a significantly greater amount of data within the inland sea and next to land, also producing more realistic fields of WSC. We use version 4.0 in these analyses. Away from land, all three versions are nearly identical. We also evaluated the use of winds from the European Advanced Scatterometer (ASCAT), to extend the period of data into the present. Coverage from ASCAT is similar to QuikSCAT version 3.1, providing less coverage than QuikSCAT v4.0 next to land and within the CIS. For that reason, we do not use the ASCAT data, relying on the ECMWF fields for consistent coverage during the longer baseline period.

The scatterometer retrieves the surface winds at irregular locations within each swath covered by each satellite pass, which we first convert to wind stress. At each gridpoint of a fixed 0.1° grid, we fit the retrievals within a set radius to a polynomial within the region. The coefficients of the polynomial are used to obtain values of wind stress at the grid point and the gradients of wind stress components at the same point. The gradients are used to estimate the curl of the wind stress. Next to land, the retrievals will come from the part of the target region not over land. The estimate then, will not apply to the center of the target region, although it will be “registered” as coming from the center. This is a minor inaccuracy and we judge it acceptable in our attempt to examine the monthly means of wind stress and WSC next to the coasts and within the inland sea. The gridded swath data are averaged within each month to form the monthly means. At each grid point away from land, there are approximately 50 independent samples during each month.

2.2.3. Upper level winds—We calculated daily blocking indices (BIs, see below) using 500 hPa daily zonal winds (U) at 12 UTC from the NCEP Climate Forecast System Reanalysis (CFSR) (Saha et al. 2014), with horizontal resolution of ~0.31°. We note that independent calculations using ECMWF winds gave similar results. These upper level winds determine whether synoptic storms are ‘steered’ into or away from the Chilean coast at
mid-latitudes. Zonal winds are downloaded across the SE Pacific from 120° to 50°W for the 39-year period: 1979–2017.

2.3. **SST and surface chlorophyll-a pigment concentration (CHL) fields**

2.3.1. **MODIS Aqua SST**—We use the monthly composite daytime SST fields from the MODIS Aqua satellite. These are available for 2003–2016 and we form the monthly mean climatology for the 14-year period 2003–2016. The MODIS-Aqua SST data are from JPL (ftp://podaac-ftp.jpl.nasa.gov/allData/modis/L3/aqua/11um/v2014.0/4km/monthly), monthly mapped (4 km) Daytime fields (v2014.0). As expected, cloud cover reduces the available data in autumn-winter, primarily affecting the months of May-July, when a majority of the pixels may be missing during a given 8-day field. Using the monthly composites, most of those months during the 14-year averaging period had more than 50% of the data missing. The other months of the year were well sampled during most years. These data are used in the analysis of seasonal cycles of SST and snapshots during 2016.

2.3.2. **Reynolds OI SST**—We use the longer time series of SST from the NOAA Reynolds OI analysis to investigate interannual variability in SST over the 1993–2016 period. This has coarser 0.25° gridding. Non-seasonal time series are created by subtracting the monthly seasonal cycles derived from the Reynolds OI SST (not the MODIS seasonal cycles found above). NOAA Reynolds OI SST are from https://podaac.jpl.nasa.gov/dataset/A VHRR_OI-NCEI-L4-GLOB-v2.0 (the JPL PODAAC web site).

2.3.3. **MODIS and Globcolour CHL**—CHL fields are available from MODIS Aqua for the 2003–16 period at 4 km resolution. A longer combined (SeaWiFS, MODIS, MERIS) Hermes Globcolour (NASA-ESA) data set with 4 km gridding is available from http://hermes.acri.fr/, covering the period from September 1997 to the present. To choose a CHL data set, we compared fields from Globcolour to those from the single-sensor MODIS Aqua data (Level-3 mapped monthly CHL data at 4 km resolution, obtained from https://oceandata.sci.gsfc.nasa.gov), derived using both the chl-a and ‘GSM’ algorithms. CHL data from both data sets suffers the same loss of data as SST, with the greatest loss during the May-July months. For seasonal cycles and interannual variability, we use the Globcolour data (also derived using the GSM algorithm), due to the fact that it produces the most coherent seasonal cycle EOFs, with less noise evident in the time series, than the two MODIS data sets. The longer period of the Globcolour data also covers part of the 1997–1998 El Niño, providing an additional comparison to 2014–2016. The Globcolour processing produces more conservative and lower values of CHL, with fewer ‘spikes’. It also eliminates more data, which results in blotchy individual monthly mean fields. For this reason, we use the MODIS data to form individual monthly mean snapshots of the development of the phytoplankton blooms in 2016.

2.4. **Methods**

2.4.1. **Blocking index**—We adopt the blocking index (BI) used in Ummenhofer et al. (2013), based on Pook and Gibson (1999), which is calculated at each longitude spanning 120°−50°W as follows:
\[BI = 0.5 (U_{25} + U_{30} - U_{40} - 2U_{45} - U_{50} + U_{55} + U_{60}),\]
where \(U_{nn}\) represents the wind at the grid point closest to southern latitude \(nn\). The westerly (positive U) wind components, north of the cyclonic low-pressure circulation anomaly (L) are averaged over 25° and 30°S (0.5 \((U_{25} + U_{30})\)); similarly, south of the anticyclonic high pressure anomaly (H) the westerly wind maximum is represented as 0.5\((U_{55} + U_{60})\), both contributing to positive BI’s. Negative (easterly) wind components between the L and H, contributing also to positive BI’s, are centered at 45°S: 0.5\((-U_{40} - 2U_{45} - U_{50})\). BIs are grouped into 73 pentads per year. In order to remove synoptic and intraseasonal variability and assess persistent blocking conditions, 3-month running means of the BI pentads are calculated.

### 2.4.2 Separation of temporal scales

Seasonal and interannual variability are our primary focus. To isolate these and eliminate intraseasonal variability, we first find the mean fields for each calendar month. These form the true seasonal cycle, which may have sharper transitions than present in harmonic fits to the data. The monthly mean seasonal cycles are removed from the raw monthly time series, leaving “non-seasonal” variability that contains both shorter “intraseasonal” and longer “interannual” periods. To eliminate the intraseasonal variability, we low-pass filter the non-seasonal time series using a loess filter with a 6-month half-span (approximately half-power) period. This filter removes signals with time scales even longer than the 2–3 months generally associated with intraseasonal scales (Rahn, 2012; Garreaud et al., 2008).

### 2.4.3 EOFs

Empirical orthogonal functions (EOFs) are used to characterize the interannual variability of the low-pass filtered time series of spatial fields. A form of Factor Analysis, these are calculated from the eigen-functions of the matrix of temporal covariances between the time series at all grid points. They produce statistically independent modes, each mode consisting of a spatial pattern and a time series (Richman, 1986; Preisendorfer, 1988). The eigenvalues of the matrix quantify the amount of variance represented by each mode over the entire region considered. It is also possible to calculate the percent of variance explained at each spatial grid point of the domain. We discuss these in the interpretation of the EOFs. If the seasonal cycles are not removed, the first 1–2 EOFs usually represent the seasonal variability (these are presented in the Supplemental Material [SM]). Note that single conventional EOFs cannot describe propagating signals.

### 2.4.4 HAB species distributions

The data base for *A. catenella* presence/absence was obtained from the Red Tides Management and Monitoring Program in the Los Lagos and Aysén Regions. As reported by that program, the historical relative abundance (RA) of microalgae refers to species number in an aliquot of 100 μL under a cover slip of 18 × 18 mm (3×) on a slide. Cell numbers are quantified on a scale of 10 levels of relative abundance from 0 (absent) to 9 (mega-abundant). This parameter is more sensitive to the presence of microalgae than cell density. The ‘standard’ periodic sampling sites are those in the CIS area during 2013–2015. Those outside the CIS and farther north reaching 39°S correspond to extraordinary sampling sites undertaken because of the exceptional bloom reported in 2016. We use the RA data only to characterize the number of sites at which *A. catenella* is present within a larger region during a specific period.

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3. Results

3.1. Distributions of variance at different temporal scales

In Table 2 we present the mean variances of the monthly time series of each variable, averaged over our domain of interest (Fig. 2, 38°–46°S) at each time scale. These are discussed in more detail in sections below. General characteristics of the variance can be summarized as:

- Seasonal variability is moderately large to very large (41–82%) for SST, CHL and all of the wind variables except QuikSCAT WSC;
- Interannual variability is greatest (57%) for SLA and moderately large for \( V_y \) (34%), SST (25%) and CHL (20%); the amount of variance in the wind variables at interannual time scales is very low, relative to intraseasonal and seasonal variability; this agrees with previous studies that find the greatest variability for winds at synoptic and intraseasonal time scales, the least variability at interannual scales;
- Intra-seasonal variability (which we remove) is largest (40–78%) for \( V_y \) (47%), CHL (39%) and all wind variables except ECMWF wind stress magnitude (\(|\tau|\)).

We primarily analyze the mean, seasonal and interannual signals over the longer time periods. In discussing the anomalies during 2014–16, however, we examine shorter event-scale phenomena (over days to months), related to the El Niños and atmospheric blocking during that period.

3.2. Mean fields of SLA, circulation, wind, SST, and CHL

Fig. 1 presents the absolute MDT height field and the geostrophic velocities calculated by AVISO. A subset of the same data in Fig. 2 provides more detail between 38°S and 46°S. The white vectors are averages of the individual velocities in boxes covering 2° (Fig. 1) or 1° (Fig. 2), providing a more schematic view of the circulation.

The eastward SPC extends broadly from 30°S to 50°S (Fig. 1). Offshore, the maximum values of 4–5 cm s\(^{-1}\) are found near 40°S. The large-scale offshore bifurcation into the Humboldt and Cape Horn Currents occurs west of Chiloé Island near 43°S, 77°–80°W. As the currents approach the coast to the southeast of that point, however, they connects to weak mean equatorward flow of 1–2 cm s\(^{-1}\) as far south as 46°S (Fig. 2). Thus, the mean coastal currents of the equatorward Humboldt Current extend south of Chiloé Island. In Fig. 1 north of 40°S, the average currents within 50–300 km of the coast are weaker than those farther offshore in the central Humboldt Current, due to the greater variability of currents within the Coastal Transition Zone off Chile, described by Hormazabal et al. (2004).

Within the CIS, lower sea levels along the eastern side of the enclosed sea (Fig. 2) create mean northward geostrophic currents. Given the greater uncertainty of the MDT field in this smaller basin, we consider direction and (especially) the magnitudes of this northward flow to represent a hypothetical result, needing further confirmation by in situ observations or models. However, the northward geostrophic flow is consistent with the reported estuarine circulation in the basin: subsurface northward transport that dominates the vertical mean,
underneath a shallow layer of fresher flow toward the south and out the mouth of the CIS through the Boca del Guafo (Sievers and Silva, 2008).

Fig. 3 presents the other annual mean fields of wind stress, SST and CHL in our domain of interest. The 10-year QuikSCAT (Fig. 3a) and 24-year ECMWF (Fig. 3c) wind stress fields depict predominant westerly (eastward) wind stress south of 40°S, increasing in strength to the south. Next to the coast, the wind stress vectors include a slight meridional component in many regions, poleward south of approximately 41°S and equatorward north of approximating 40°S, becoming more strongly equatorward north of 38.5°S. Where these meridional components are small, the orientation of the coast determines whether the eastward winds are slightly upwelling- or downwelling-favorable. The surprise from the scatterometer fields is within the CIS and the wide enclosed region to its south, where the magnitudes of the wind stresses appear as strong or stronger than in the open ocean west of Chiloé Island. Their orientation is slightly more poleward than west of the island, producing upwelling on the western boundary of the CIS, downwelling on the east. Although the ECMWF wind stress values resembles the scatterometer field in both direction and magnitude in the open ocean, their decrease in magnitude appears to be too great next to the coast north of 41°S. The most severe difference occurs in the CIS, where the resolution of the model reanalysis winds is too coarse and the magnitudes of the winds are far too weak.

Regions of negative WSC (Fig. 3a, colors) produce Ekman pumping and upwelling, such as found strongly in the coastal areas between roughly 38°−39°S and 40°−41°S, with regions of weaker negative curl located within a degree of longitude from the coast between 38° and 44°S. The regions of moderate positive and negative curl in the offshore areas correspond to the anticyclonic (counter-clockwise) and cyclonic (clockwise) curvature in the wind fields, respectively, whereas the stronger values of curl are usually due to horizontal shear near coasts. Within the CIS, strong curl-driven upwelling occurs in the central and southwest portion of the basin, while the northern half of the CIS experiences strong positive WSC (downwelling-favorable), as does the Boca del Guafo and a band stretching to the southwest.

SST and CHL values in the 100 km next to the coast north of 43°S show the effects of both upwelling and advection. As described by other studies (Aguirre et al., 2014), the mean coastal SST values are cooler by 1°−2°C than farther offshore, with a slight break in the cool coastal band between 38.5°S and 39.5°S (Fig. 3b). This break occurs where surface currents turn cyclonically and advect warmer offshore water east and toward the coast north of the sheltering cape (Punta Galera, 40°S). This region is also between the negative (upwelling-favorable) extremes in wind stress curl. Mean CHL concentrations are higher (> 1.0 mg m$^{-3}$) in a similar coastal band north of 42°S. Farther offshore, there is a north-south SST gradient, cooling by 3°−4°C from north to south, with a modest increase in CHL concentrations in the south. South of 44°S, the currents dip southward before connecting to the equatorward advection within the cool band of higher CHL waters.

Within the CIS, the southern half (G. Corcovado) is cooler by 1°−2°C than the northern half (G. Ancud), while the G. Reloncaví is even warmer. The CIS is generally warmer on the eastern side than on the west, consistent with upwelling on the west and Ekman transports to
the northeast, as implied from the scatterometer wind stress. The southwest part of the CIS next to Chiloé Island is especially cooler on average (the location of negative curl-driven upwelling). Increased tidal and wind mixing around the Desertores Islands in the middle of the CIS and quasi-homogenation of the southern G. Corcovado could also account for lower SST and CHL concentrations in the middle and southern half of the CIS. The cool waters in the SW part of the CIS connect to cool water in the Boca del Guafo and around the islands to the south (Fig. 3b), where tidal mixing is expected to be energetic. CHL concentrations are higher in the northern CIS (G. Ancud and Reloncaví) and south of the CIS in the C. Moraleda, lower in the G. Corcovado and lowest around and south of the central Desertores Islands.

3.3. Seasonal variability

Seasonal variability in wind forcing, circulation, SST and CHL are presented below. We present maps of the surface fields during the months of January, April, July, and October to represent the austral seasons of summer, autumn, winter and spring, although we refer to rapid changes during other months. Maps of all months can be found in the SM. Overlapped line plots of the full monthly seasonal cycles of different parameters at representative locations provide a complementary view. ‘Seasonal’ EOF spatial fields and their mean monthly time series are also included in the SM, which provide a more concise definition of the seasonal transitions and their dominant patterns of spatial covariability.

3.3.1. Maps of seasonal surface wind stress and WSC—Physical forcing of the circulation at the surface is primarily due to fields of surface wind stress and WSC, associated with the surface atmospheric pressure fields. An examination of summer and winter mean monthly fields of SLP, averaged over the 1993–2016 period using ECMWF ERA-Interim data (not shown), produces fields consistent with previous analyses described above (Saavedra and Foppiano, 1992; Strub et al., 1998; Garreaud et al., 2008; Rahn, 2012).

The most realistic portrayal of seasonal variations in wind stress and WSC is given by high-resolution maps of the scatterometer’s wind fields for the four representative months (Fig. 4a–d). Seasonal anomalies (long-term mean removed) for January and July are included in Fig. 4e and f. These show that the seasonal variation (creating the seasonal variance in monthly fields) is primarily in the alongshore component of the surface wind stress, which is well represented by the meridional ($\tau_y$) direction except south of 44°S. It is greatest in the north, minimum in the offshore region between 42° and 44°S, with a relative increase again south of 44°S. The spatial average of the seasonal variance in $\tau_y$ in Table 2 is 1.14·10$^{-3}$ (N m$^{-2}$)$^{-2}$, corresponding to a regional mean standard deviation of 0.034 N m$^{-2}$. This is consistent with the alternation in the wind stress magnitudes of 0.05 N m$^{-2}$ in the northern regions of Fig. 4e and 4f, much less in the south.

The January (summer) anomaly wind field (Fig. 4e) is predominantly equatorward everywhere but mostly confined to the region north of 42°–43°S. Within the CIS and along the islands south of 43°S, the predominant wind direction in January is eastward in the full field. The equatorward coastal anomalies in the south (Fig. 4e) are due to the reduction of the slight poleward components (Fig. 3a). There are, however, slight equatorward
components within the CIS that should produce minor downwelling and upwelling along its western and eastern boundaries, respectively. Negative WSC (favoring upwelling) in January is found next to the outer west coast from the southern tip of Chiloé Island to the northern boundary of our domain, caused by a reduction in the strength of the wind stress in a narrow (50 km or less) band next to the land. Negative WSC also appears along the southwest border of the CIS, remaining mostly positive in the north. For reference, the maximum WSC value on our color bar, \(-0.4 \times 10^{-6} \text{N m}^{-2}\), corresponds to upwelling velocities of order 10 m mo\(^{-1}\).

As the position of the SPH moves to the north in autumn and winter, the alongshore component of the wind changes sign in April south of approximately 42°S (Fig. 4b), becoming poleward (in an absolute sense) everywhere south of 38°S in July (Fig. 4c). The reduction of wind stress next to the coast again produces a narrow band of more intense WSC along the outer west coast (this time positive) throughout much of our domain north of 43°S in July. The same pattern is found along the eastern edge of the CIS, where the strongest components of poleward alongshore wind stress and WSC occur in winter. In October, the SPH begins to move back south and upwelling favorable conditions return along the west coast north of about 40°S. Although we are primarily interested in the meridional wind stress for its effect on upwelling, we note the July westward anomalies south of 44°S, consistent with the conclusion that the eastward winds between 45° and 55°S are strongest in summer (Garreaud et al., 2008).

In the SM we include the EOFs of monthly QuikSCAT and ECMWF \(\tau_y\), along with the first EOF of QuikSCAT curl \(\tau\), with the seasonal cycles retained. These clarify or emphasize several points:

- The autumn transition to downwelling between April-May is rapid, while the spring transition to upwelling during September-November is more gradual;
- The first EOF for ECMWF \(\tau_y\) is similar to that of QuikSCAT over the open ocean, showing their agreement in that domain;
- The time series of the first EOF of QuikSCAT curl \(\tau\) is coincident with the time series of \(\tau_y\) EOF-1, reflecting the creation of the narrow band of wind stress curl by the shear of the meridional wind stress next to the coast;
- Maps of the percent variance explained by the first EOF of scatterometer \(\tau_y\) (Figure S3a) show that this EOF explains over 50% of the variance in the CIS, giving us confidence in the scatterometer’s seasonal cycles there.

The fact that the variability in the monthly wind stress is predominantly in the north-south component and that this variability is concentrated in the north may seem counterintuitive, given that the winter storms are expected to be stronger in the south. The explanation is that the individual monthly means average out the alternating northerly and southerly winds of the storms. To demonstrate this, we first high-pass filter the 6-h ECMWF wind stress time series with a half-power point of 7 days, to keep only the winds associated with passing synoptic storms. We then find the absolute value of the 6-h wind stress and average those values over monthly periods. Fig. 5 presents the spatial pattern of the first EOF of these
monthly storm-related wind stress magnitudes, with and without the seasonal cycles. The spatial pattern is concentrated in the southern half of the domain, with maximum (minimum) values in the south during winter (summer), as expected. We return to this figure when analyzing the anomalous wind forcing during 2014–2016.

3.3.2. Maps of seasonal ocean response—Moving from wind forcing to ocean circulation, maps of absolute dynamic height fields (SLA + MDT) and associated geostrophic velocities are presented in Fig. 6a on a larger scale (25°S-55°S) to provide the context for our domain of interest. In January (summer), the SPC between 45° and 50°S near 80°W deflects to the south before connecting to equatorward coastal currents, which can be found consistently next to the coast as far south as approximately 48°S, with local regions of equatorward currents found as far south as 51°S. Thus, characterizing the coastal transports off southern Chile as uniformly poleward south of the SPC bifurcation is a mistake during austral summer. This pattern reverses in July (winter), when the eastward SPC near 40°S takes a slight northward meander before connecting to the poleward coastal current that extends south along the coast to the tip of the continent (Fig. 6a). In the long-term mean (Fig. 1), both of these branches of the SPC are visible: between 40°–42°S and 46°–48°S.

Removing the MDT field leaves just the altimeter-derived seasonal SLA and velocity anomalies (Fig. 6b), which demonstrate that the seasonal change in January is due to a drop in sea level in a coastal band south of approximately 35°S. This band of low SLA is found progressively offshore as one moves north from 35°S. Thus the maximum in equatorward currents in Fig. 6a also moves offshore north of about 35°S in summer.

Likewise, the July field in Fig. 6b reveals a band of high SLA next to the coast south of approximately 35°S, associated with poleward geostrophic current anomalies along its western boundary. In Fig. 6a, these anomalies reduce, but do not reverse the band of low absolute dynamic heights and equatorward velocities during July north of 37°S, the same latitude identified in previous studies as the northern limit of seasonally reversing wind stress.

Returning to 38°S-46°S, monthly climatological mean maps of high-resolution satellite SST and CHL distributions (Figs. 7 and 8) are overlaid by climatological mean monthly vector averages of the absolute vector surface currents from the altimeter heights (Fig. 7) and monthly climatologies of ECMWF reanalysis winds (Fig. 8). Besides the full fields for the four months representing the seasons (panels a-d), we also present January and July climatological anomalies, with the temporal means removed. In these figures, the years used to form the seasonal cycles of surface currents (2003–2016) and wind stress (1998–2016) are the same as those available to form the seasonal cycles of SST (from MODIS) and CHL (from Globcolour), respectively. In the SM, we show similar climatological overlays for all calendar months, this time using the full period (1993–2016) for the calculation of the surface current and wind climatologies. A comparison between the SM figures and Figs. 7 and 8 confirms that the seasonal cycles of currents and winds do not change significantly when the shorter or longer periods are used.
As is evident in Fig. 7, most of the seasonal variance in surface velocity is due to temporal variability in the north-south (alongshore) surface velocity component ($V_y$), spatially concentrated next to the coast (Fig. 7e and f). Seasonal extremes show absolute velocity values of $\pm 5$–$8$ cm s$^{-1}$, with the stronger currents in the north and in the CIS (Fig. 7a and c). The temporal variance in seasonal $V_y$ averaged over the entire region (Table 2) represents 19% of the total variance; the variance value of 5.50 cm$^2$ s$^{-2}$ represents a spatially-averaged standard deviation of magnitude 2.3 cm s$^{-1}$.

The seasonal deflections of the offshore currents is even clearer in Fig. 7a and c, north and south of 42°S, as is the seasonal reversal of the coastal currents. Examination of the full sequence of months in the SM clarify that the coastal currents north of 44°S become systematically equatorward in October, stronger in a coastal band in the middle of the region (Fig. 7d). The equatorward coastal currents increase in the north to reach a maxima next to the coast in December-January (Fig. 7a), expanding offshore during January-April (Fig. 7b). There is a rapid autumn transition to poleward coastal currents in the north in May, which extend southward in June and reach a maximum in July. Although the poleward coastal velocities in winter are nearly as strong as the coastal equatorward velocities in summer, the band of winter velocities is more concentrated toward the coast, so the one-degree average (white) poleward vectors in winter are not as strong as the average equatorward vectors in summer, especially in the south. After July, coastal poleward magnitudes decrease at the coast and spread offshore, reversing to equatorward currents near the coast in October. Also in the SM, the first EOF of seasonal $V_y$ confirms that the ~100 km wide coastal band of meridional velocities explain 30–60% of the variance, as they also do in the western CIS, where the seasonal changes in velocity follow the pattern in the large-scale coastal band. The second mode explains another 10–15% of the variance in three narrower bands: offshore of the first mode band; very nearshore; and in the eastern CIS. Together, the two modes combine to describe the seasonal migration of equatorward velocities that develop right next to the outer coastline north of 42.5 °S (Mode-2) in October-December and move offshore to occupy a wide coastal band in January-April. This band reverses to poleward during May-August, as the equatorward currents continue to move westward during July-December.

Spatial patterns in high-resolution SST in Fig. 7a–d are somewhat hidden by the larger seasonal pattern of heating and cooling. However, cooler temperatures occur next to the west coasts in all seasons except spring (here represented by October). The full annual sequence of months in the SM depicts warmer temperatures in the 100 km next to the coast in September-October between 40° and 44°S. Thus, seasonal changes in currents carry warmer water south next to the coast during winter cooling and downwelling, as currents carry cooler water to the north during summer warming and upwelling. The net result is that the seasonal changes in SST next to the coast are less than those farther offshore: Anomalies in SST next to the coast in January (Fig. 7e) are +1° to +2 °C, while those offshore are +2° to +3 °C. In July the coastal anomalies are −1° to −2°C, compared to −2° to −3°C offshore (note the change in color bars in Fig. 7e and f). The contrast between the lower seasonal ranges in SST along the coast compared to offshore is summarized well by the seasonal EOFs of SST in the SM. In Table 2, the seasonal variance in SST is from the coarser Reynolds SST over the longer period (1993–2016) and represents 70% of the overall
monthly variance. The spatial average of the variance over the entire domain of 3.04 \((^\circ C)^2\) corresponds to a standard deviation of 1.7 \(^\circ C\).

Within the CIS, the cooler region in the southwest and warmer northern basin are evident in all seasons except winter. This results in a greater range of seasonal variability (similar to offshore values) in the northern basin (G. Ancud and G. Reloncaví) and in the C. Moraleda to the south, with lower variability (similar to the outer coasts) in the central basin and around the Desertores Islands. Seasonal changes in currents within the CIS are small compared to the equatorward long term means. Given uncertainty in those values, we do not further interpret the seasonal changes in currents within the CIS.

In Fig. 8, the monthly climatological mean wind stress values from the ECMWF reanalysis are similar to the scatterometer winds in Fig. 4, except for the lower values next to the coast and in the CIS. We primarily show them for different periods to demonstrate that the specific sets of years used in the climatologies do not produce qualitatively different seasonal cycles. Here we use 1998–2016. In the full monthly seasonal cycles in the SM, we present the seasonal monthly means for the full 1993–2016 period.

The CHL fields display a relatively simple pattern of higher concentrations in a wide coastal band in January (a clearer summer coastal pattern than the cool SST). Coastal CHL values decrease in April (autumn) and becomes weaker and more patchy in July (winter), then increase in the coastal regions in October (spring). CHL values are significantly higher (> 1 mg m\(^{-3}\)) than offshore values in this wide coastal band during all seasons, with a minimum in offshore CHL concentrations in summer, when the water column stratifies and the surface waters become depleted of nutrients. The seasonal cycle represents about 41\% of the monthly variance in Table 2, with a value of 0.45 (mg m\(^{-3}\))^2, corresponding to a standard deviation of about 0.67 mg m\(^{-3}\). With the mean values of each pixel removed, Fig. 8e and f demonstrate the concentration of the variance in the ~100 km coastal band (north of ~44\°S), with seasonal ranges of ± 1 mg m\(^{-3}\) next to the coast. That band widens in the south, where there may be a greater degree of variance due to noise (cloud contamination). In the SM, the first two (seasonal) EOFs combine to explain only about 23\% of the monthly variance, demonstrating the much more variable nature of CHL data compared to SST. The full set of monthly fields and EOFs in the SM confirm minimum coast values in winter (June-September) and maxima during November-February. Within the CIS, there is a consistent pattern of higher CHL concentrations in the northern half, with lowest concentrations just south of the Desertores Islands in the center, and somewhat low concentrations in the southern basin, especially in winter. CHL concentrations are also consistently high in the Canal Moraleda south of the CIS.

### 3.3.3. Seasonal time series at selected locations

To evaluate temporal relationships between the different parameters, we provide overlays of time series of the full seasonal cycles from the eight locations shown in Fig. 9g: along six ‘Offshore’ and ‘Inshore’ lines west of the outer coast at three latitudes (Fig. 9a–d); and at two locations within the CIS (Fig. 9e and f). The temporal mean has been removed from all variables over the longest period possible: 1993–2016 for ECMWF \(\tau_y\), altimeter SLA and \(V_y\); 1998–2016 for CHL; 2003–2016 for MODIS SST; and November 1999 through October 2009 for QuikSCAT.
\[ \tau_y \] . Using the shorter periods associated with MODIS and QuikSCAT produces only minor changes in the seasonal cycles of ECMWF \[ \tau_y \], SLA, and \[ V_y \].

The ‘Offshore’ locations (Fig. 9a and b) are at approximately 75°W, characterizing the middle of the region of interest shown in the previous figures, rather than the extreme offshore locations. The seasonal patterns are typical of open ocean locations in eastern boundary currents at mid-latitudes: higher values of SST and surface heights occur during summer heating and steric expansion of the warmer water, lower values at the end of winter. Ranges are \( \pm 2–3 \) °C and \( \pm 2–3 \) cm. Surface current and wind stress anomalies are equatorward (positive) in summer and poleward in winter, with meridional winds leading currents by approximately one month in summer. Ranges are \( \pm 2–4 \) cm s\(^{-1}\) and \( \pm 0.04 \) N m\(^{-2}\). The CHL pattern is weaker, with spring (September-November) maxima at 41°S and 42.5°S, and a weaker spring maximum at 43.6°S. The ranges is \( \pm 1 \) mg m\(^{-3}\).

At the ‘Inshore’ coastal locations next to the mainland at 41°S and Chiloé Island at 42.5°S, wind and current anomalies are again more equatorward in summer, poleward in winter (Fig. 9d), without clear lead/lag relationships between winds and currents. Surface temperature patterns are also simple (Fig. 9c): warm in summer and cool in winter. Unlike farther offshore, however, sea levels have only a minor relative peak in summer (January-February), due to onshore advection of denser water at depth during summer upwelling. Sea levels peak next to the coast in late-autumn and early winter (May-June) as winds become poleward and drive onshore Ekman transport. CHL patterns have clear winter minima; maxima are less uniform, occurring in spring-summer. The ranges at the Inshore locations are similar to, or slightly less than, those at the Offshore locations.

Within the CIS (Fig. 9e and f), the north-south scatterometer wind stress values are similar in timing and strength to those west of the coast, but 3–4 times as strong as the ECMWF winds (note the change in scales in Fig. 9f). Anomalies of altimeter-derived currents at the northern location follow the winds in summer, with a slight lag and values of \( \pm 1 \) cm s\(^{-1}\), less than the regions west of the island. Currents are not available at the southern location. Surface heights and temperatures are like those at the ‘Inshore’ locations west of the island with similar values, maximum during winter downwelling, minimum in spring and autumn, with a very minor relative peak in summer. CHL concentrations in the northern CIS display spring (November) and autumn (March-May) peaks, with a greater range (\( \pm 2 \) mg m\(^{-3}\)) than west of the island. In the south of the CIS, there are spring-summer (November-January) peaks in CHL concentrations. Seasonality of the CHL signal at the southern CIS location is similar to that to the west in the Boca del Guafo (Fig. 9c).

### 3.4. Summary of seasonal variability and comparison to the California Current

The closest analog to the southern Chile region is the northern California Current System (CCS), extending into the Alaska Gyre along the British Columbia coast, inshore of the bifurcation of the North Pacific Current. One might even consider Vancouver Island and the inland Salish Sea to be the counterparts of Chiloé Island and the CIS. Regarding the other wind-driven, mid-latitude, eastern boundary current (EBC) systems in the global ocean, the northeast Atlantic geometry next to Europe is much more complicated, while the southeast
Atlantic does not extend to the equivalent latitudes of our domain of interest (Mackas et al., 2006).

The major differences between the southern Chile system and its North American counterpart are due to two factors: (1) the seasonal change in the location and strength of the high pressure system in the southeast Pacific is much more limited and weaker than that in the northeast Pacific; and (2) the bifurcation of the ‘West Wind Drift’ current (South Pacific Current) occurs at a lower latitude (approximately 43°S) than that of the North Pacific Current (approximately 50°N).

In the southeast Pacific Ocean, the movement of the center of the SPH between winter and summer is from approximately 27°S to 35°S, with the location of highest coastal SLP moving from approximately 35–36°S to 42°S. The net result is that winds change direction seasonally between ~34°S-45°S, in relatively good agreement with previous results (Saavedra and Foppiano, 1992; Sobarzo et al., 2007; Garreaud et al., 2008; Rahn and Garreaud, 2013). In the northeast Pacific, the center of the North Pacific High (NPH) moves from ~28°N in winter to ~40°N in summer, with the greatest coastal SLP moving from 36°N to 52°N (Huyer, 1983; Strub and James, 2002b). This produces a much larger region of seasonal wind reversals between ~35°N and 53°N. Note that the winter locations of the pressure and wind systems are similar. The difference is the much greater poleward extension of the North Pacific pressure and wind systems in summer.

The winds that result from seasonal changes in the SLP systems between 38°S to 46°S are revealed by the new Version 4.0 QuikSCAT surface wind data set, in details not available in previous versions near land (Figs. 3 and 4). The dominant direction of the mean winds changes from equatorward north of 40°S to eastward south of 41°S, another difference between the southern Humboldt and northern CCS, which lacks the strong eastward wind component. Farther north off Vancouver Island and the British Columbia coast north of the island, the strength of the eastward (or cross-shore) component of wind stress does increase, suggesting again that our southern Chile region includes the equivalent of the southeast Gulf of Alaska. Off Chile, the meridional wind stress changes from strongly to slightly equatorward as one moves from 38° to 44 °S in summer, driving upwelling along the outer coast of Chiloé Island and the mainland farther north, as well as along the eastern boundary of the CIS. In winter, poleward winds force downwelling everywhere south of 38°S and along the eastern boundary of the CIS. The monthly wind stress magnitudes off Chile are somewhat less than off Oregon and Washington. The long-term mean values in Fig. 3 off southern Chile are between 0.05 and 0.1 N m², compared to 0.1 N m² or more off OR-WA. Seasonal monthly mean magnitudes off OR-WA rise to 0.2 N m² in winter and 0.1 N m² in summer, compared to 0.05 to 0.10 N m² off Chile in both seasons.

The high-resolution QuikSCAT wind data also resolve patterns of WSC not seen before off Chile, with a general finding that patterns of WSC augment upwelling (negative curl over equatorward wind stress) and downwelling (opposite conditions). The zero contour of WSC defines the maximum in meridional winds, which is located 50–100 km offshore, slightly closer to the coast than reported farther north by Aguirre et al. (2014). These patterns of
wind-driven upwelling along the outer coast are similar to the seasonal changes of wind forcing in the northern CCS and southern Alaska Gyre.

During winter downwelling, poleward advection of warm water next to the coast briefly reverses the normal onshore-offshore gradient of SST in September, creating a band of coastal water slightly warmer than offshore. In October the gradient is near zero (Fig. 7d) and in summer (Fig. 7a) offshore temperatures increase to 15–17 °C, while upwelling creates a cool band of coastal water, approximately 1–2 °C cooler than in the offshore deep ocean to the west. Patterns in the cooler coastal SST suggest that both upwelling and advection are at work. This temperature difference between the coast and the offshore ocean in summer is much less than the 6–7 °C difference found in measured temperatures between the coast of Oregon and regions 100 km to the west (Huyer et al., 2007). Some of this weaker difference may be due to the use of satellite SST in our study, compared to the in situ observations at 10 m depth reported in the climatology of Huyer et al. (2007). The satellite climatology of Venegas et al. (2008) in the northern CCS shows somewhat weaker temperature differences of 3–4 °C between the coast and deep ocean offshore, still greater than in our analysis of the southern HCS.

The seasonal change in SLA at the coast off Chile is of order 4–6 cm (Fig. 9), less than the range of 20 cm or more reported off Oregon and Washington by Venegas et al. (2008). The range in the offshore SLA rise and fall in summer and winter (due to heating and cooling) is more similar (~5–10 cm) off North America to that seen off Chile, but the summer decrease and winter increase due to upwelling and downwelling, respectively, are much less off southern Chile. Off Oregon and Washington, one does not see the initial steric rise at the coast in spring, followed by the drop due to upwelling in summer and then the rise again in autumn and winter due to downwelling. Thus, the SLA and SST results combine to suggest that upwelling has a stronger effect in the northern California Current than in the southern Humboldt Current, cooling the coast to a greater degree and preventing the summer rise in SLA from even starting. The stronger onshore and weaker alongshore components of the wind stress off South America may explain some of the difference. The fact that the Chilean region is more directly inshore of the eastward South Pacific Current than the coast along Oregon and Washington may also decrease the effects of upwelling. Veitch et al. (2010) describe this off South Africa in terms of “geostrophic convergence,” whereby an offshore geostrophic current impinges (converges on) the coast and suppresses upwelling. We hypothesize that this is one of the effects at work off southern Chile. Modeling studies would be useful in further exploring these dynamics.

Maximum geostrophic currents in Fig. 7 are of order 8–10 cm s\(^{-1}\) in summer and 5–6 cm s\(^{-1}\) in winter. These are lower than the velocities of 10–15 cm s\(^{-1}\) found in the modeling study by Aguirre et al. (2014). Some of this difference may be due to a lack of wind-driven Ekman surface velocities in our geostrophic currents, which would augment the summer currents. However, these values are still less than the climatological geostrophic velocities off Oregon, with values of ~20–40 cm s\(^{-1}\) in summer and 5–10 cm s\(^{-1}\) in winter when calculated from in situ density data, as reported by Huyer et al. (2007). Measured velocities off Oregon are slightly less, of order 10–15 cm s\(^{-1}\) (Huyer et al., op. cit.). The finding of weaker geostrophic velocities in the southern Humboldt Current is consistent with the
weaker effects of seasonal upwelling and downwelling forcing, resulting in weaker density and SLA fronts.

Surface CHL concentrations are everywhere higher within approximately 2° of longitude of the coast and within the CIS and other channels between islands. The bands of higher CHL values are strongest and most continuous during summer upwelling, similar to the results of Yuras et al. (2005), but in a wider coastal band than in the 20 km width found off northern Chile. The difference between the northern and southern basins in the CIS agrees with the results of Lara et al. (2016). The higher CHL in the outer coastal band resembles the high satellite-derived CHL off Oregon and Washington reported by Venegas et al. (2008). As with SST, both upwelling and advection affect the patterns of CHL. Additional complications are added by the fact that the satellite sensor receives its signal from within the water column down to one optical depth, which varies in space. Thus, the lower values in regions such as the southern CIS may be due to the deeper vertical mixing of the same amount of phytoplankton that is concentrated in a shallow surface layer in the northern CIS. Other processes, such as the depletion of nutrients in the mixed layer and the concentration of the phytoplankton in a deep chlorophyll maximum layer also affect seasonal changes in the horizontal patterns of CHL as seen by satellites, as do the effects of dissolved organic matter, optically complex (‘Type II’) waters, etc. Intrinsic biological processes (effects of predators, pathogens, etc.) also determine patterns of patchiness in CHL (Lara et al., 2010), consideration of which is beyond the scope of this paper. All of these processes act to reduce the amount of CHL variability that can be represented by the seasonal cycles in CHL.

The only analog to the CIS in the northern California Current is the Salish Sea (Lange, 1999), inshore of Vancouver Island. Like the CIS, tidal forcing is strong and increases as one moves from the main entrance in the south to the narrower regions in the north. It is also used for salmon aquaculture and experiences HAB events. However, it is a less regular basin and a detailed comparison to the Salish Sea is beyond the scope of this study.

Within the CIS, winds follow the offshore pattern, equatorward in summer and poleward in winter. During summer, Ekman transports create upwelling in the east and downwelling in the west, building up higher SLA on the west within the CIS, creating equatorward currents; this is reversed during the poleward winds of winter. The equatorward summer winds also create horizontal shear next to the coasts, creating negative and positive bands of WSC along the eastern and western margins, respectively. Poleward winds in winter produce the opposite pattern of WSC within the CIS. These bands of WSC strengthen coastal upwelling and downwelling in the CIS, as they do along the outer coasts.

The southern half of the CIS is consistently cooler, especially the southeast coast of Chiloé Island, connecting in autumn-winter to similarly cool regions in the Boca del Guafo area and the Canal Moraleda to the south. During most of the year, the northern half of the CIS is warmer. Although the positive and negative values of WSC may contribute to the warmer and cooler northern and southern CIS, the strong topographic mixing and more homogeneous vertical structure around the central islands and in the south is the more important factor in keeping SST cool in the southern basin, while inflow of fresh water and strong stratification keeps SST values warmer in the north.
3.5. **Statistical patterns of interannual variability - EOF analysis**

In this section we first remove monthly (calendar mean) seasonal cycles at each grid point’s monthly time series based on the longest period of data available (1993–2016, except 1998–2016 for Globcolour CHL), followed by a six-month half-power low-pass filter to remove the intraseasonal variability. EOFs are then calculated to characterize interannual variability of the surface wind stress and ocean response. The exception to this process is for meridional wind stress, where we also show the first EOF of the unfiltered data, with seasons included, to demonstrate the effect of removing the seasonal cycles and filtering. Other than this exception, percent variance amounts reported for each mode refer to the filtered, non-seasonal (“interannual”) time series. Each parameter is described briefly, before looking in detail at the 2014–2016 period in Section 3.6. In Section 3.7, interactions between parameters are considered in a more complete discussion of interannual variability over the complete 24-year period, using details of the 2014–2016 event for guidance. Note that the seasonal cycles that are removed are not simple annual harmonics. Their removal, followed by the six-month low-pass filter does not remove quasi-annual signals associated with year-to-year changes in the timing of seasonal transitions.

3.5.1. **Wind stress**—The spatial patterns and full time series for the first mode EOFs of ECMWF meridional wind stress are found in Fig. 10a (seasons retained, unfiltered, 90% of the variance) and Fig. 10b (seasons removed, filtered, 83% of the variance). We do not consider the results in the CIS, since the spatial patterns confirm that ECMWF reanalysis winds do not represent well the winds within the CIS, reducing their strength to values similar to terrestrial winds. When the seasonal cycles are retained, maps of the percent variance explained (not presented) show values of 80–100% north of 44°S, dropping to 50–70% in the southwest corner of the domain. Fig. 10a reveals the large degree of interannual variability in the individual yearly seasonal cycles. Some years have multiple summer or winter peaks (winter 2012 and 2013, summer 1996–1997); others have more continuous seasonal cycles (1995, 1999, and 2014). We note again that these monthly means underestimate the magnitude of the instantaneous winter storm winds, which can be evaluated in Fig. 5. The first mode of the filtered interannual wind stress (Fig. 10b) explains 70–100% of the variance over most of the offshore region, dropping to 50–70% next to the coast north of 39°S and offshore south of 45°S. It has a uniformly positive spatial pattern outside of the CIS and a noisy time series compared with SLA, SST and CHL (signals with 6-months and longer time scales). However, it is useful in assessing whether peaks in the seasonal winds are anomalously strong or weak during specific events identified by the SLA signals. We return to the non-seasonal wind stress below.

3.5.2. **Altimeter SLA and geostrophic velocity**—The first interannual EOF of SLA in Fig. 11 accounts for ~69% of the variance in the filtered interannual SLA fields. The spatial pattern of the first mode is uniformly positive with values of order one (about 0.9–1.2) outside of the CIS. Maps of percent variance explained (not shown) indicate that 70–90% of the variance is represented over most of the region west of the outer coast, 50% in the far offshore region north of 42°S. It has a uniformly positive spatial pattern outside of the CIS and a noisy time series compared with SLA, SST and CHL (signals with 6-months and longer time scales). However, it is useful in assessing whether peaks in the seasonal winds are anomalously strong or weak during specific events identified by the SLA signals. We return to the non-seasonal wind stress below.
~1 cm during the next decade (prior to 2014). This is roughly consistent with global SLR values of ~3 cm/decade. A more quantitative estimate of regional SLR is found by linear fits of the raw SLA time series to time at each grid point. The slopes of the linear trends (not shown) are noisy but have values near 3 cm/decade in the offshore region and 1.5–2.0 cm/decade closer to the coast. The Mode-1 time series also appears to reveal a jump between 2002 and 2003. Below we show that this apparent jump is the result of the super-position of interannual variability (the rise in SLA associated with the 2002–2003 El Niño) and the long-term trend.

Recognized equatorial El Niños occurred in 1997–1998 (strong), 2002–2003 (moderate), 2009–2010 (moderate) and 2015–2016 (strong). These can be seen in the Mode 1 time series, with mid-year and year-end pulses of high sea level during the 1997–1998, 2009–2010 and 2015–2016 events. The double-pulse signature is much clearer when the six-month filter is not applied. Although the two moderate El Niños can be identified as positive signals in the time series, other interannual signals have similar magnitudes. Some of these, such as 2004–2005, 2006–2007 have been identified as marginal warm events using indexes such as the Multivariate El Niño Index (MEI (Wolter and Timlin, 1993, 1998)). Knowing that these signals are there, one can identify most of them in the second mode as well (2015 is the exception), although they are not easily differentiated from other similar signals. Note that anomalous equatorward winds identified in Fig. 10b oppose the high SLA signals during the autumn-winter peaks in the strong 1997–1998 and 2015–2016 events. The high SLA in the second mode is clearly concentrated in the 75–100 km next to the coast and within the CIS, whereas the first mode is positive everywhere with moderate coastal and offshore maxima. During each of the El Niños and many of the other maxima, the peak in the more coastal second mode precedes that of the more offshore first mode, consistent with an offshore spreading of the high SLA coastal signal. Maps of percent variance explained for the second mode show that it primarily represents 20–30% of the variance in the coastal region of positive values in Fig. 11b, slightly more (20–40%) in the CIS. The offshore region is not represented by the second mode.

The first two interannual modes of meridional geostrophic currents are presented in Fig. 12. The time series of the first interannual V_y mode (Fig. 12a) is similar to that of the negative of the second interannual SLA mode (Fig. 11b). In the spatial pattern for the first non-seasonal mode of V_y (13% of the variance) a band of equatorward (positive) flow is slightly separated from the coast when values of the time series are positive. This band overlaps the zonal gradient of the Mode-2 SLA pattern (negatively correlated as shown below), demonstrating their geostrophic relation. Geographically, the percent variance explained (not presented) is 30–60% only in the coastal band of positive values and in the wester CIS, nearly zero elsewhere. The negative first mode time series indicates the strongest poleward motion in this band during the year-end periods of 2002, the beginning of 1993 and during the double pulse of 1997. There are many other years, however, with positive and negative signals that are nearly as strong as the moderate El Niños. There are positive signals (equatorward flow) in the first mode following the El Niños (1998, 2003, 2010 and 2016). During the entire period of 2007–2013, there is a tendency for more equatorward currents. The second mode appears more random and the map of explained variance reveals that it
explains almost zero variance except in offshore patches. We do not interpret the second mode further.

### 3.5.3. Satellite SST and CHL

The first EOF mode of filtered interannual SST (Fig. 13a) explains ~70% of the variance. As with the first SLA mode, there is no zero crossing in the spatial pattern, which has values of order 1.0 ± 0.2 over most of the region. The reader is reminded that the SST fields used for the interannual analysis are from the Reynolds OI climatology with 0.25° gridding, rather than the MODIS 4 km data. These fields provide an overall indication of differences in the first mode signals in the central CIS compared to offshore, with little spatial differentiation within the CIS. Time series extremes of ± 1 °C produce a range of approximately ± 1 °C over much of the region west of the coast, lower ranges within the CIS. The percent of variance explained (not shown) is 60–90% over most of the offshore domain between 39°–45°S, 45–60% in the CIS. The non-seasonal Mode-2 (Fig. 13b) depicts a wedge of warmer water in the north, concentrated along the coast from 38° to 44°S, with little signal in the south and in the CIS. Between 25% and 50% of the variance is represented by this mode in the wedge of positive coastal values in the north, almost zero percent elsewhere except in the far offshore region between 41° and 45°S, where 20–30% of the variance is explained. Spatial extremes of ± 2 combine with time series extremes of ± 0.5 °C to produce another 1 °C rise and fall in this coastal wedge. When the time series is positive, this distribution of temperature is similar to that expected during an El Niño. Although the MODIS data begin in 2003 and miss the 1997–98 El Niño, interannual EOFs computed from the MODIS data (not shown) produce similar time series and spatial patterns during 2003–2016.

In the Mode-1 time series, the two strongest El Niño events produce relatively strong warm anomalies over much of the region during 1997–1998 and 2015–2016. Some, but not all, of the other identified ‘warm periods’ also correspond to positive values of the Mode-1 time series, indicating that these warm events covered a broad offshore area. The Mode-2 time series displays moderate peaks during most of the El Niño and other warm events, as expected given the coastal wedge shape of the spatial pattern. But these peaks in the time series are not distinguishable from other peaks (e.g., 1996, 2000, 2001, 2002, 2011). In the Mode-1 time series, there are two periods during which SST drops quickly (1999 and 2009) and then rises slowly over the several years. In the Mode-2 time series, there is a noticeable trend to more negative values, which would correspond to a gradual rise in offshore temperatures between 40° and 44°S (~0.2–0.4 °C) and a gradual decrease in SST within the wedge of higher spatial values along the coast in the north (also ~0.2–0.4 °C).

For the non-seasonal CHL EOF modes (Fig. 14), the spatial patterns are concentrated next to the coast and within the CIS, with a wider region of positive values in the south in Mode 1, stronger and narrower in the north in Mode 2. Together, the first two modes explain only ~26% of the non-seasonal variability. For the first mode, the percent of variance accounted for is 40–50% in the offshore positive patch between 38°–40°S and 40–60% in the offshore patch between 42°–44°S. During many of the years between 1998 and 2009, the Mode-1 time series can be thought of as representing year-to-year changes in the timing and magnitude of the seasonally productive spring-summer period, producing positive peaks between October and March (especially 1998–1999, 1999–2000, 2002–2003, 2005–2006).
This pattern changes after 2009, when positive peaks appear in the fall-winter periods in the middle of the years and negative troughs are seen during spring-summer. This produces a damping of the seasonal cycle after 2010, consistent with the results of Lara et al. (2016).

There is also a gradual rise from low values in 2000–2001 to higher values between 2007 and 2013, decreasing to lower values in 2016. The second mode has the appearance of a narrower coastal mode and the percent of variance explained is only significant (15–30%) over this coastal band between 39°–42°S and within the CIS. The pattern we note in the time series is a tendency for positive coastal anomalies in CHL at the end of the warm periods in 1998, 2003, 2005, 2007 and 2016, with several other peaks of similar magnitude (2011 and 2013).

3.6. 2014–2016 anomalies

The altimeter SLA and geostrophic velocity EOF time series identify strong interannual anomalies during the El Niños of 1997–98 and 2015–16. In this section we look at the 2014–16 period in detail, which includes an additional type of anomaly, an extension to the north of a dinoflagellate bloom of *A. catenella* during the first half of 2016. Although we cannot establish rates for *A. catenella*’s northward expansion, we can document the physical conditions during that event and discuss whether the environmental factors favored this extension.

3.6.1. El Niño conditions during 2014–2016—The oceanic El Niño pathway consists of the wave guide along the equator, connecting to wave guides along the eastern margins of the basins in the northern and southern hemispheres. In Fig. 15, we calculate the SLA in a region within 75 km of the coast along each of the standard altimeter tracks next to the coast of South America from TOPEX (1995–1998) and Jason-2 (2013–2016), removing the seasonal cycles (1993–2016) from all time series. A trend of 1.5 cm/decade is also removed. The actual trend over this large region is not uniform and the trend we remove is a compromise. Fig. 15’s Hovmöller diagram shows the time-latitude progression of the non-seasonal SLA over these two four-year periods. During 1997, there is a clear progression of high SLA southward from the equator during late-April through July and again in October through April 1998, with signals restricted to lower latitudes after January 1998. The maximum anomalous SLA values are 20–25 cm during both periods. A somewhat similar (but reduced) pattern is found in 2015–2016. At the lower latitudes, the signals are approximately half the value as during 1997.

Also evident in Fig. 15 is the April-May 2014 period of high SLA at the equator, which can be seen to move down the coast relatively strongly to approximately 32°S in May, with weaker signals perhaps reaching south of 40°S. The SLA signals decrease after June. There is a period of slightly higher SLA values at low latitudes in August-October, but not a second strong pulse as found in 1997 and 2015. Greater values of SLA are found in September 2014 at mid- and higher-latitudes.

In our domain of interest, the time series of the non-seasonal EOFs of SLA (Fig. 11) indicate two pulses of high SLA next to the coast in both 2014 and (weaker) 2015 (Fig. 11, Mode-2), which lead to an increase in sea level in a broader coastal region starting in mid-2014 (Fig.
11, Mode-1). These are accompanied by brief (several month) spikes of poleward currents in a band next to the coast during 2014 and (weaker) 2015 (Fig. 12, Mode-1) and also by higher SST values during most of 2015 (Fig. 13, both modes), continuing in the more offshore SST Mode-1 during the second half of 2016. Time series of the Mode-1 (unfiltered and seasonal) alongshore winds (Fig. 10) suggest that the wind forcing was poleward and downwelling favorable (supporting the high SLA) during the middle (winter) of 2014 and 2015 but equatorward, opposing the high SLA in spring-summer 2015–2016.

To look at the forcing and response in more detail over a larger latitudinal extent, we show the non-seasonal alongtrack SLA (Fig. 16a), using the median values from the 75 km of the altimeter tracks closest to the coast, averaged over multiple tracks in latitudinal bands stretching from 5°N to 45°S. The time series in Fig. 16a (and Fig. 16b) are not smoothed with the 6-month filter, retaining the intraseasonal details important during these events. These are compared to the non-seasonal meridional wind stress (ECMWF), averaged over the same locations as the tracks’ coastal crossings for the bands between 15°S and 45°S. To better compare winds and SLA, poleward, negative values of $\tau$ are plotted toward the top of Fig. 16a, to coincide with the positive SLA values. Two SST El Niño indexes (Niño 1 + 2 and Niño 3.4) are also included in the top panels for comparison (the wind stress anomalies are negligible north of 20°S, as seen in the middle panel). In Fig. 16b we include altimeter cross-track geostrophic velocities calculated from the difference in height between the first two 75 km sections of altimeter track next to the coast, again plotting negative (poleward) values of the cross-track velocity toward the top of the page. Meridional wind stress anomalies in Fig. 16b are not identical to those in Fig. 16a, since we average the meridional wind stress in Fig. 16b over the locations where velocities are calculated, using a smaller number of tracks than for SLA, due to the orientation of the tracks. As a consequence, we average over a larger low-latitude band (0°–20°) in Fig. 16b than used in Fig. 16a.

During 2014, high SLA non-seasonal signals (Fig. 16a) with magnitudes of 9–18 cm occur next to the coast from mid-April through mid-June from 5°N to approximately 15°S (top two panels). As one moves farther south, the period of high coastal SLA becomes shorter, centering on late-April and May. The magnitude also decreases to ~5 cm south of 35°S. Anomalous wind forcing, as indicated by the non-seasonal meridional wind stress, is weak at the low- and mid-latitudes, but becomes more vigorous between 35° and 45°S. During September-October 2014, relatively weak (5 cm) positive SLA values occur at the lowest latitudes south of the equator (0°–15°S), with slightly higher values between 20° and 45°S. Between 35° and 45°S, there are downwelling-favorable poleward wind stress anomalies during May-September 2014 but the higher SLA values in the south are only coincident with moderate downwelling-favorable anomalies of wind stress in August-September. Otherwise, periods of poleward wind stress anomalies prior to September 2014 and after October do not correspond well to the periods of high SLA, which appear more connected to the lower latitude signals.

Altimeter-derived cross-track velocities within 150 km of the coast in Fig. 16b confirm poleward velocity anomalies between 0° and 35°S during March-April and September-October 2014. We note that negative values of both $\tau_y$ and $V_y$ are again plotted toward the
top of the figure to remain consistent with high SLA in Fig. 16a. These signals do not appear within our region of interest (here represented by 35°–45°S). Instead, poleward velocity anomalies between 35° and 45°S occur during June-September 2014. Wind stress anomalies are also predominantly poleward during June-September, but there is weak agreement between the wind and velocities on the event scale. At the bottom of Fig. 16b, the same meridional wind stress and surface velocity data are presented with the seasons retained, for later reference.

El Niño indexes in Figs. 16a and 16b only vaguely relate to the anomalous SLA and meridional velocity signals during 2014. The Niño 3.4 index, representing SST anomalies along the equator far to the west of South America, has minor positive peaks during April-July and September-December 2014, the episodes of higher-latitude SLA peaks. The Niño 1 + 2 index reflects the SST anomaly between 0° and 10°S next to the west coast of the continent, with a stronger signal in May-June 2014, following the first peak in SLA. The decrease in this index during the second half of 2014 is due to the lack of an eastern equatorial SST signal, i.e., the failure of a full El Niño (as defined by SST) to develop during the last half of 2014.

During summer 2014–2015 between 35°–45°S, winds are upwelling-favorable and anomalously strong, as demonstrated by the positive wind stress and wind stress anomalies in Fig. 16b. Coastal SLA values in Fig. 16a, however, remain moderately high along the entire South American coast, in agreement with Fig. 11, in which the time series of the Mode-1 non-seasonal SLA remains high in early 2015. The first period of high sea level during the strong 2015 El Niño commences at lower latitudes in early-March, followed by a rapid rise in the Niño 1 + 2 SST index in late-March. The SLA signal spreads rapidly to 45°S, where it lasts to mid-June, aided by winds (only between 35° and 45°S) in May. Alongshore (cross-track) velocities in Fig. 16b are predominantly poleward, in separate month-long pulses, between 0° and 20°S during the first period of high sea level in 2015 (March-June). Between 20° and 35°S, the poleward velocities are reduced to month-long pulses during March-April and July.

The second period of high SLA associated with the 2015 El Niño begins in early-August north of 15°S (Fig. 16a) and in late-August farther south. Several pulses of high SLA move down the coast with approximately monthly intervals during August-February, reversing in April 2016 north of 15°S. Between 35°–45°S, periods of more continuous high SLA along the coast occur during September-October and late-November through mid-March. In February-March 2016, the high SLA separates from the coast and moves to the west (see also Fig. 17c). Alongshore velocities in Fig. 16b south of 35°S during the second 2015 period are around 6–8 cm s⁻¹ in August-September and again in late 2015 and early 2016. In this southern region the northern signals are less evident in Fig. 16b. At times, the alongshore velocities appear more related to the alongshore wind stress, sometimes coincident, sometimes with lags of ± a month.

With this larger-scale development in mind, the progression of conditions in our region of interest can be better understood. Fig. 17 presents a more direct depiction of the sea level and geostrophic velocity fields for specific months in 2014–2016, focusing on the domain
between 39° and 44°S, including Chiloé Island and the mainland to the north. Here we include the seasons, to gain a better understanding of the direction of the true currents (without the long-term mean, which is approximately 2 cm s\(^{-1}\), toward the north), since these are the currents which will be of interest in discussing the transport of *A. catenella* toxic populations to the mainland in the next section. In Fig. 17a–c the first four months of each year are found; Figures d–f continue the sequences for each year with the months of May, July, September, and November. Looking at the first part of the year, one can compare the seasonal mean currents in Fig. 7 for January and April with the same months in these three years, demonstrating a large degree of year-to-year variability. The generally equatorward velocities of 2–8 cm s\(^{-1}\) in the seasonal monthly means for these two months are somewhat similar to the fields in the Januaries of 2014 and 2015, with stronger and more coherent equatorward velocities in the Februaries of 2014 and 2015 and in March of 2016. Eddies appear in March and April in the fields approximately 100 km offshore in the coastal transition zone (Hormazabal et al., 2004) and continue into May in Fig. 17d–f.

In 2014, the normal equatorward currents in January-March appear somewhat disrupted in April and strongly disrupted in May, although the high sea level next to the coast in May 2014 is not as strong nor as continuous as in May 2015. The connection to the lower latitude signals during May 2014 is established by the coincident high SLA values along the coast in Fig. 16a. The SLA and velocity signal during September 2014 is even more continuous in Fig. 17d between 39°–44°S and in Fig. 16a SLA, connecting back to the equator. In full maps of SLA (not shown), September 2014 shows a continuous, narrow band of non-seasonal coastal SLA extending from 0° to 50°S. It is not present in August and breaks up during October and November. Sea levels also increase in the larger-scale eastern equatorial Pacific in September, after decreasing in July and August. At the same time, the Niño 1 + 2 Index does not show anything other than a small blip at the beginning of October. The implication is that there was a brief second pulse of sea level at the equator during September 2014 that did not have a significant SST signal. We hypothesize that such a signal was generated at the lower latitudes and travelled down the coast to our region, aided by poleward winds at the mid- and higher-latitudes (20°–45°S). The cross-track velocities in Fig. 16b are poleward in September 2014, supporting the hypothesis that such a signal did occur. It may have had little significance at the equator and little effect on the atmosphere, but it affected the SLA and circulation off southern Chile.

The height fields in the early months of 2015 and 2016 (Fig. 17b and c) show the result of the increase in height that began at the end of 2014, seen in the non-seasonal Mode-1 for SLA in Fig. 11. The peaks of the 2015 El Niño occur at low latitudes (0°–15°S) in March-June and September-December (Fig. 16a). The mapped fields during these months (Fig. 17b and c) show the high SLA next to the coast in May, July and September 2015. High offshore sea levels persist during January-March 2016, while low SLA appears next to the coast in March 2016 after a period of prolonged, strong upwelling (Fig. 16b). The low SLA spreads offshore in April. In March-April 2016, the high offshore sea levels combine with the low coastal values to create strong jets of equatorward currents, occurring a month later in the year than during the previous two years. Thus, the 2015–2016 El Niño signal in SLA appears to be gone at the coast by March and early April, but plays a role in creating stronger equatorward currents during those months.
3.6.2. The 2016 Harmful algal bloom of A. catenella—As described in the Introduction, the presence in the phytoplankton of the toxic dinoflagellate species A. catenella has been detected at ever more northern locations along the coast of southern Chile during the past several decades. Since 1998, it has been regularly observed during sampling within the CIS in summer. Table 3 summarizes observations of the presence of A. catenella in different regions, providing one aggregate observation during summer-autumn of 2013 and 2014, monthly summaries between October 2015 and March 2016, with more frequent (extraordinary) sampling in April-May 2016. The years 2013 and 2014 were typical of recent years, with observed presence in the northern CIS in 2013 but only in the southern CIS in 2014. Starting in October 2015, observations of A. catenella spread northward from the southern CIS to the northern CIS in November. In February 2016, the species was observed along the outer coasts of the islands west of Aysén and Canal Moraleda (~45°S, not included in the Table 3). In early April 2016, observations of A. catenella stretched to the west coast of Chiloe Island and the Chacao Channel north of the Island, continuing to spread northward along the mainland in April and May. Sampling in the Los Ríos Region on May 6 found the species as far north as 39.4°S (Buschmann et al., 2016). By mid-May (May 17), there was only one coastal location where the HAB was observed along the mainland, near 39.9°N. Note that these observations were only made at locations on the coast and that no sampling was carried out at sea. Thus, the observed distributions may underestimate the actual spread of the species, along with its decline in mid-May, north of the CIS.

What conditions accompanied this northward spread of A. catenella? This period includes the last months of the strong 2015–2016 El Niño, as seen by the rapid fall of non-seasonal SLA at low latitudes at the end of March 2016 in Fig. 16a. The Niño 1 + 2 SST index also drops to negative values at the same time. Although periods of moderately high SLA persist between 15° and 35°S, coastal SLA in the southern band between 35° and 45°S drops in mid-March and stays low for several months. Non-seasonal meridional wind stress anomalies in Fig. 16a become strongly upwelling-favorable in the same band during January-March, helping to lower sea level at the coast, while high sea levels caused by the El Niño persist in the offshore regions during January-March 2016. Fig. 17c demonstrates how these combine with the lower coastal sea levels in March-April to create strong equatorward currents, forming a pathway that extends from south of Chiloé Island to the region north of 40°S. However, this pathway is not unique to 2016. Similar currents are observed during most summers, as seen in general in Figs. 6 and 7 and during January-February 2014 and 2015 in Fig. 17a and b. Whether A. catenella populations were present on the west coast of Chiloé Island or at other locations outside the CIS and islands during the other years and available to be transported north is unknown, since those locations are not usually monitored.

Nor is the strong upwelling during March unique to 2016. In Fig. 16b, the bottom panel repeats the time series of meridional wind stress and currents, retaining the seasonal cycles (still removing the long-term mean). Upwelling during March 2014 and January-February 2015 appear as strong or stronger than in March 2016. More generally, Montecinos and Aceituno (2003) describe the typical strengthening of the anti-cyclonic atmospheric circulation in austral summer during El Niños, which would enhance equatorward wind...
stress. During summer 2015–2016, Garreaud (2018) specifically cites the strengthened high pressure system associated with atmospheric blocking as the source of the stronger equatorward winds observed in January–March 2016.

What appears to be more unusual, however, is the lack of strong downwelling in May and June 2016. Fig. 18 presents a comparison of daily (ECMWF) meridional wind stress during the first six months of 2015 and 2016, averaged over 2° boxes next to the coast in our domain of interest between 38° and 46°S. Here we follow the standard notation and plot positive (equatorward) wind stress toward the top of the page. Along the outer west coast between 38° and 46°S, strong upwelling-favorable winds in January and February decrease in March 2015 as storms increase in the south. In 2016, equatorward winds continue through March and into early April, punctuated by only 2–3 storms in April. Between late-April and early June, however, the winds in 2016 are unusually calm, lacking the nearly continuous poleward winds found during storms in May 2015. How unique is this calm period in autumn 2016?

The strength of high-frequency (several-day period) synoptic storm winds over a longer period can be evaluated in Fig. 5, where the ECMWF winds are high-pass filtered to keep only periods less than 7 days, creating monthly means of the absolute magnitude of the strength of synoptic-scale storm winds during 1993–2016. The first EOF of this high-frequency (HF) wind stress magnitude (Fig. 5b), explains ~90% of the variance in the monthly time series. The spatial pattern (Fig. 5a) is maximum south of 42°S but is still strong near 40°S. The seasonal cycle is maximum in winter and minimum in summer, as expected for the synoptic storm winds. The normal seasonal magnitude in May is nearly half of the maximum value (red cross in Fig. 5a). Note that negative values of the time series are found for this positive quantity because the EOF calculation removes the temporal mean field. The time series in Fig. 5b (seasonal cycles retained) and Fig. 5c (seasonal cycles removed) both indicate that May-June 2016 experience the most anomalously low values of synoptic storm wind stress magnitude for the 24-year time series (several months in 2004 come close to this minimum). The unusual seasonal cycle during 2016 is especially obvious in Fig. 5b.

Garreaud (2018) attributes the absence of storms and the resulting drought off southern Chile during the 2015–2016 summer to the phenomenon of atmospheric blocking. To explore this process, we form a Blocking Index (BI), as described in Section 2.4.3. This quantifies the upper level easterly wind anomaly near 45°S, which forms between the cyclonic and anticyclonic atmospheric circulation anomalies found to the north and south of this latitude, respectively. These upper level wind anomalies oppose the eastward movement of storms, deflecting them either to the north or south. Just as this reduces the clouds and precipitation associated with the storms (Garreaud et al., 2013; Garreaud, 2018), it also suppresses their strong winds. Fig. 19 presents two views of the BI temporal and spatial development during 2015 and 2016, first as a time-longitude Hovmöller diagram (Fig. 19a) and then as line plots of BI values at 90°W and 80°W (270°E and 280°E, Fig. 19b). Note that the coast is at approximately 74°W.
There are two periods of strongly positive values of the BI: late-August to mid-October 2015 and late April to early-June 2016 (our calm period). During these periods, the 5-day values of the BI exceed magnitudes of 20 m s$^{-1}$, a value only reached by 10% of the 5-day BI values in the 39-year time series. Moreover, when we form three-month running means of the 5-day values at 80°W, only five periods have values exceeding 12 m s$^{-1}$, demonstrating persistent strong blocking. These are the periods centered on September and October 2015 and the periods centered on April, May, and June 2016. Thus, the degree to which storms were blocked during these periods is exceptional during the 1979–2016 record. In discussing the drought during 2015–2016, Garreaud (2018) notes the unusual positive values of both the Niño 3.4 SST Index (associated with deep convection that triggers PSA atmospheric teleconnections) and the Southern Annular Mode (SAM). Both of these modes are statistically associated with blocking by high pressure along the southern Chile coast; but the two climate modes are usually out of phase, occurring at different times (Montecinos and Aceituno, 2003). Garreaud (2018) attributes the extreme strength of blocking (and the lack of precipitation) in summer 2016 to the coincidence of positive values of the two modes.

When the high pressure system that forms part of the dipole described above moves offshore of the coast, it creates strong equatorward coastal wind stress and upwelling, as seen during November 2015 through March 2016 (Fig. 16b), following the low winds during the blocking event in September-October 2015. It also reduces cloud cover, resulting in anomalously high values of solar radiation, including the photosynthetically active radiation (PAR) needed to stimulate phytoplankton growth. Satellite-derived estimates of surface CHL cannot distinguish between species of phytoplankton but do show the presence of general blooms in the upper ocean. Monthly composites of satellite CHL for February-May 2016 (Fig. 20, in this case using the sensor on the MODIS-Aqua satellite for monthly snapshots) record the extensive bloom of phytoplankton during February-March 2016 along the west coast of Chiloé Island and the mainland at least as far north as 38°S. Moderately high values remain next to the coast between 38° and 41°S in April, weakening further in May. Strong mixing and upwelling of nutrients would favor diatoms as dominant species in the bloom during January-March. As winds decrease in April and solar radiation remains relatively strong, diatoms sink out of the stratified upper water column, which is rapidly depleted of nutrients. Under these conditions, dinoflagellates such as *A. catenella* are favored, since these can swim between the upper illuminated layer and the nutrients at depth. Deep chlorophyll maxima are also formed, which result in lower values of surface CHL as seen by satellite sensors.

Satellite SST fields also show the effects of upwelling in creating a band of cold SST anomalies over the same region in March, with values of −1°C to −2°C (Fig. 21a and c). After the cessation of winds in May, the offshore front moves back to the coast (while temperatures cool overall in autumn), producing warm anomalies of +1°C to +2°C next to the coast (Fig. 21b and d). We hypothesize that phytoplankton, including *A. catenella*, were advected to the coast with the warm water after the cessation of strong upwelling-favorable winds in April.
The poleward and onshore movement of fronts and water to the coast from the deep ocean is well documented during ‘upwelling relaxations’ (Roughgarden et al., 1991), although the effect on the transport of biological populations depends on the location of the populations within the water column and their ability to move vertically (Shanks and Brink, 2005; Dudas et al., 2009).

In summary, we can describe a sequence of physical conditions which would have increased the chance of an exceptional bloom of *A. catenella* along the coast of Chile north of Chiloé Island in 2016. These include: (1) Observations of *A. catenella* throughout the CIS and to the south in Canal Moraleda and Aysén, and along the outer coasts of the islands south of Chiloé Island by February 2016, extending to the west coast of Chiloé Island by April; (2) Strong equatorward winds due to the enhanced high pressure system, driving increased upwelling and mixing during January-March 2016, with an observed widespread phytoplankton bloom (most likely dominated by diatoms); (3) A transport pathway of equatorward currents from south of Chiloé Island to north of 40°S during early 2016, especially strong during equatorward jets in March and April that result from a combination of high offshore SLA (El Niño related through the ocean) and vigorous wind-driven upwelling (El Niño related through the atmosphere, as the high pressure moves offshore (Garreaud, 2018)); (4) an extreme lack of winds from late-April through early-June 2016 due to an exceptional period of atmospheric blocking of storms (also connected to the El Niño through the onshore movement of the high pressure), resulting in clear skies and excess insolation. Given observations of *A. catenella* along the west coast of Chiloé Island and at coastal locations north of 41.5°S in early May, we hypothesize that *A. catenella* had arrived in the offshore water north of Chiloé Island by April and were advected onshore during the wind relaxation at the end of April, benefiting from the high levels of PAR and strongly restratified water that developed in the coastal waters.

### 3.7. Discussion of 1993–2016 interannual physical variability

We have presented the interannual variability in the coastal ocean between 38°–46°S, using 24-year records of monthly anomalies of wind stress, SST, SLA and geostrophic velocity, with a shorter record for CHL. We have also looked in detail at anomalous conditions during 2014–2016, including the relation to ENSO indexes and low-latitude signals of SLA and $V_y$.

Since a primary mode of interannual variability in the Pacific Ocean is related to tropical ENSO variability, we look now at the relation of the longer record of interannual variability and ENSO, using the details during 2014–2016 to inform our interpretations.

When discussing the effects of El Niño signals on eastern Pacific mid-latitude coastal regions, the usual question is whether the signals travel from the tropics via oceanic or atmospheric pathways. The more complete question regards the role played by each pathway, working together simultaneously or at different times. In the results presented above, the primary oceanic El Niño signals include increased sea level and poleward transport, along with higher SST values. Signals travelling through the atmosphere include changes in sea level pressure, upper level winds and storm tracks, altering surface winds, cloudiness, precipitation and incoming solar radiation (through cloudiness). Only surface
winds are included in the observations considered in our region, except in constructing the Blocking Index from upper level (500 hPa) winds.

As a summary, we present three ENSO indexes in Fig. 22, along with interannual variability represented by EOF-1 for SLA, $\tau_y$ and $V_y$. We also include the non-seasonal alongtrack SLA signal (the most basic form of altimeter data) from the inner 75 km of the tracks between $0^\circ$–$15^\circ$S and $35^\circ$–$45^\circ$S. These monthly time series have been smoothed with a 3-month half-span loess filter. The shorter half-span retains more structure in the ocean’s response to El Niño anomalies, while still removing much of the intraseasonal variability. The altimeter data have also been detrended with a uniform trend of 1.5 cm decade$^{-1}$.

The Niño-3.4 and MEI appear in panel 22a, identifying the well-known events (1997–1998, 2002–2003, 2009–2010, 2015–2016), along with other moderate warm periods (early 1993, 1994–1995, 2004–2005, 2006–2007). Since Niño 3.4 is simply the temperature anomaly in an equatorial region between $120^\circ$ and $170^\circ$W, it is often used to represent a connection between the equatorial ocean and the East Pacific mid-latitude ocean through the atmospheric processes of convection and Rossby waves. However, it is also connected with westerly wind bursts, equatorial Kelvin waves and sea level signals that travel through the oceanic pathway. Since the Niño 1 + 2 index is a coastal SST index, it might be thought of as more closely tied to the oceanic pathway, assuming that equatorial signals need to go through this region in order to travel to higher latitudes along the South American coast.

In Fig. 22b, the time series from the first EOF of detrended SLA is strikingly similar to both Niño 3.4 and MEI indexes in panel 22a. Removing the trend in the SLA time series eliminates the appearance of a rapid ‘jump’ in 2003 noted in Fig. 11a. Rather it is simply the transition from several years of negative ENSO index values to several years of positive values. In Fig. 22c and d, there is a correspondence between the first EOF time series for meridional wind stress (panel 22c) and meridional geostrophic velocity (panel 22d), both of which have higher frequency natures. Finally, the time series of the Niño 1 + 2 Index displays many of the same features as the coastal alongtrack SLA between $0^\circ$–$15^\circ$S and $35^\circ$–$45^\circ$S.

Correlations between the various signals in Table 4 quantify these relationships. The correlations need to be interpreted with caution, since the altimeter signals are not all independent. The EOF’s of SLA and $V_y$ in the region between $38^\circ$ and $46^\circ$S are related, since $V_y$ is calculated from gradients of SLA; both are also related to the raw SLA signal along tracks between $35^\circ$ and $45^\circ$S. However the raw coastal SLA signal along tracks between $0^\circ$ and $15^\circ$S, the wind stress, SST and CHL signals are independent of the other signals. Below we report the highest correlation for lags of $-1$, 0, or $+1$ month, allowing for the smoothing of the time series. Autocorrelations of each variable allow us to estimate the decorrelation scales as follows: 6 months, for values of 6 months or less ($\tau_y$, $V_y$ and CHL-1), 9 months, for 6–9 months (SLA-2, SST-1, SST-2, CHL-2 and SLA 35–45) and 12 months (SLA-1). Dividing the total number of months by these scales results in numbers of independent samples of 48, 32 and 24, respectively. The corresponding correlation values for 95% and 99% significance are given in the last column of Table 4. Correlations at or above these values are shown in italics and bold italics, respectively.
The results of the correlations are similar for the SLA EOF-1 time series (SLA-1, top row) and the alongtrack SLA between 35°–45°S (SLA 35–45, bottom row). Both are strongly correlated with the three ENSO indexes and with each other, representing the strong connection of the large-scale SLA off southern Chile and the ENSO events. The first EOF of the meridional velocity ($V_y - 1$), is also correlated with the alongtrack SLA 35–45, but even more strongly correlated with the second EOF of SLA (SLA-2). The velocity is not independent of the southern SLA variables, since the nearshore SLA is used to calculate the velocity. This is clearly shown by the even stronger correlation of the second EOF of SLA (SLA-2) and $V_y$, since SLA-2 represents the coastal SLA better than SLA-1.

Is sea level variability off southern Chile communicated through the atmosphere or ocean? Considering the atmosphere, correlation values between the meridional wind stress and the ENSO indexes are not significant. This is initially a surprise, since previous studies have found a connection between the Niño 3.4 index and the high pressure system off central and southern Chile (Montecinos and Aceituno, 2003; Garreaud, 2018). However, the examination of the blocking events during 2015–2016 (above) demonstrates that the anomalously high pressure off southern Chile can produce both strong upwelling winds when the anti-cyclonic system remains offshore, and periods of low wind speeds when it moves onshore (Garreaud, 2018). The higher frequency nature of the winds in Fig. 22, compared to the ENSO indexes and the SLA variables, reflects this dependence on the higher frequency nature of the movement of the pressure systems on synoptic and intraseasonal time scales. Examples of the former include the cyclonic systems that move through the area from west to east (Rahn and Garreaud, 2013) and low pressure signals that propagate poleward along the coast (Garreaud et al., 2002; Garreaud and Rutllant, 2003). Examples of the latter include the 1–2 month blocking events described above. Thus, although it is difficult to connect the alongshore winds statistically to the ENSO indexes, they are affected by the pressure systems and ENSO variability through the atmospheric pathway.

What about the oceanic pathway? The correlation between the alongtrack coastal SLA in our (approximate) domain of interest (SLA 35–45) and the similar alongtrack SLA between 0° and 15°S (SLA 0–15), is also among the greatest values and slightly stronger than the correlations between SLA 35–45 and the ENSO indexes. The first EOF of SLA in our region of interest is also strongly correlated with coastal SLA in the tropics, SLA 0–15. These correlation can be seen in Fig. 22, applying to both positive and negative anomalies of SLA. The strong correlation between SLA 35–45 and the Niño 1 + 2 index strengthens the conclusion that the oceanic pathway is strong: interannual anomalies in the coastal ocean sea level between 35°–45°S are closely connected to coastal sea level and temperature in the tropics in a continuous fashion, not simply during specific ENSO ‘events’.

The oceanic pathway is much simpler along South America than along the Central and North America west coast. Signals travelling north from the equator must pass through the Panama Bight, the ITCZ, North Equatorial Countercurrent, Coast Rica Dome, the strong gap-winds along Central America, the mouth of the Gulf of California, large capes along Baja California and the Southern California Bight. Signals travelling south from the equator have no equivalent series of complex currents, winds or coastal geometry to negotiate. The
change in coastal direction along southern Perú is characterized by very large scales of curvature that pose no problem for the propagation of sea level signals. Winds are relatively weak along northern Chile, giving traveling signals less interference as they move to the higher latitudes. Along the South American coast, tropical coastal signals of moderate to large magnitude, with positive and negative signs, can reach and have an impact on coastal sea levels and currents along southern Chile (36°–48°S), as demonstrated by the time series in Fig. 22.

Returning to the wind and velocity, the high frequency variability in the meridional wind \( \tau_y \) at least partially drives the meridional coastal velocity, \( V_y - 1 \) (\( r = 0.50 \), the highest correlation for \( \tau_y \)) and its coastal sea level signature, SLA-2. The surface velocity is also correlated with the raw coastal SLA (SLA 35–45) necessarily, since the coastal SLA is half of the signal used to calculate the velocity. Because of this mechanical connection to the coastal sea level, the velocity is also moderately well correlated with the ENSO indexes and SLA-1. The velocity is also well correlated with the coastal sea level at lower latitudes (SLA 0–15), which we interpret to represent the connection of coastal sea levels through the oceanic pathway.

There is another set of relationships between coastal sea level, meridional velocity and ENSO that can be seen in the individual time series and which contributes to the correlation coefficients between \( V_y \), ENSO indexes and \( \tau_y \). After the sea level along southern Chile is raised by signals arriving along the coast from lower latitudes, those signals eventually move offshore in the form of Rossby waves. This creates a gradient between lower sea level at the coast and higher offshore values, reversing the coastal sea level and velocity signs, as the ENSO indexes are also reversing. At the end of some events, there are additional equatorward winds, perhaps associated with the offshore movement of an enhanced high pressure system. This occurred in January-April 2016, seen most clearly in Fig. 16b. At the end of many of the ‘warm/high-SLA’ periods identified by the red underlines in Fig. 22d, one can often find examples of poleward velocities changing to equatorward velocities coincident with or following equatorward winds (Fig. 22c).

In Fig. 22, we have emphasized sea level, meridional velocities and wind stresses, rather than SST and CHL. Correlations between the first EOF of SST in the 38°–46°S region (SST-1) and the ENSO (MEI and Niño 1 + 2) indexes are significant at the 95% level, connecting to SST at lower latitudes through the oceanic pathway. The highest correlation for SST-1 is for SLA-1, followed by SLA 35–45, indicating warm SST anomalies during the high SLA periods. This is the expected pattern for ENSO warm events and Fig. 13 confirms a pattern of warm SST-1 and/or SST-2 patterns during the identified warm/high-SLA periods. The highest correlations of SST-2 are with \( \tau_y \) and \( V_y \) (both negative), showing typical upwelling relationships between SST in the coastal wedge north of 43°S (Fig. 13), and alongshore wind and currents.

Surface CHL fields produce even noisier time series than SST, resulting in EOF modes that explain only 19% and 7% of their interannual variance. The first EOF of CHL (CHL-1) is only correlated with coastal SLA in the 35°–45°S region. The negative correlations with the coastal SLA are in the expected direction for high sea level events, associating high sea
levels with deeper pycnoclines and nutriclines, restricting the upwelled water to come from above the nutricline, as described off Perú by Barber and Chavez (1983) and Chavez (1996) and over the large-scale Pacific by Thomas et al. (2012). If the nutricline is not too deep, however, the expected increase in equatorward winds noted above during austral summer of the El Niño event should lead to positive signals in the coastal CHL EOF time series. Fig. 14 indicates positive signals in the CHL time series following warm events in 1998, 2003, 2005, 2007, and 2016.

From the above results, using 1997–98 and 2015–2016 as primary examples, we propose a hypothetical set of relationships during and following El Niño conditions in our region of interest: One or more pulses (~month-long) of poleward currents and extended periods of high SLA and SST occur during the event, arriving mostly through the oceanic pathway, possibly divided into mid-year and end-of-year periods (a ‘double-pulse’). Toward the end of the event during austral summer (early in the next year), the high sea level moves offshore, leading to equatorward coastal currents. At around the same time, there may be periods of strong equatorward winds, associated with a strengthened anti-cyclonic SLP system, as occurred in early 2016 (Fig. 18b). The last feature is communicated through the atmospheric pathway. If the sea level has reduced sufficiently (ending the period of depressed pycnoclines), a phytoplankton bloom may accompany the equatorward winds and upwelling (Fig. 14b). The enhanced anti-cyclonic system also has the potential to move over the coast and produce periods of calm and clear weather due to atmospheric blocking (Fig. 19), as it did in April-June 2016 (Fig. 18d). However, 2016 was unusual, even for El Niño periods, due to the superposition of positive phases of both El Niño and SAM climate modes (Garreaud, 2018).

3.8. Discussion of the extensive A. catenella HAB bloom of 2016

In both the open-ocean outer coastal ocean and within the CIS, some of the observed changes in 2016 may represent typical physical conditions and biochemical interactions that influence the overall seasonal succession of phytoplankton species (Montecino et al., 2006; Iriarte et al., 2007; Ulloa and Grob, 2009, Marrari et al., 2017). This succession includes the replacement of a diatom bloom by dinoflagellates, along with the subsequent disappearance of the dinoflagellates during the decreasing radiation and temperatures of autumn. However, the anomalous extension of the HAB in April-May 2016 to the north of Chiloé Island was not previously reported, leading us to consider the anomalous environmental conditions during 2015–2016.

At the end of Section 3.6.2, a sequence of events surrounding the northward extension of the A. catenella bloom during 2016 is presented. These form a hypothesis, that the observations of A. catenella along the mainland coast north of 41.5°S are due to a combination of a more southern bloom of A. catenella (not unusual) that is advected to the north in January-April offshore of the coast, then comes ashore during the sudden wind relaxation near the end of April, growing during the unusually calm period in late-April and early-May (Fig. 18d). Garreaud (2018) reports high levels of solar radiation during this same period of drought, which would favor growth of phytoplankton assemblages and further stratify the upper water column. The 2015–2016 El Niño played several roles during this sequence. It
provided the high coastal sea levels through the oceanic pathway during 2015, which moved offshore in early 2016. It strengthened the anti-cyclonic atmospheric circulation through the atmospheric pathway, enhancing the normal equatorward winds and upwelling, also in early 2016. The upwelling further lowered sea levels at the coast in March, combining with the offshore high sea level to augment the strong equatorward ocean currents, as it brought nutrients to the surface next to the coast to fuel the initial phytoplankton bloom in January-March. The same high pressure system moved over the coast and blocked storms in April-June to create the unusually calm period. The cessation of upwelling and mixing reduced the supply of nutrients, including the silicon needed to sustain diatoms, further favoring the success of dinoflagellates (Olsen et al., 2014).

Considerations of stability and vertical mixing lead us to the effects of the drought associated with the 2015–2016 El Niño. The flow of fresh glacial meltwater water into the fjords increases the stability of the water column and plays a crucial role in controlling productivity on seasonal scales (Aracena et al., 2015). The severe drought and reduced fresh water inflow to the CIS during summer 2015–2016, as documented by Garreaud (2018), decreased the upper water column stratification in the CIS and would have allowed nutrients to be mixed to the surface in greater amounts. Iriarte et al. (2016) suggest that reduced streamflows might result in a dominance of flagellate species in coastal waters of southern Chile. During this time, observations reported by León-Muñoz et al. (2018) document an extreme bloom of a different ichthyotoxic algae (the raphidophyta, Pseudochattonella) within the CIS. If populations of *A. catenella* also increased within the CIS, normal estuarine circulation would provide a path to the outer coastal ocean.

Heterotrophy may also play a role in the succession between diatoms and *A. catenella*. After a phytoplankton bloom, diatoms are often infected by fungi (Gutiérrez et al., 2016), producing episodes of widespread fungal parasitism and massive mortality of diatoms. Dissolved and particulate inorganic material are released during these episodes, providing a substrate for heterotrophic activity. Although *Alexandrium* species have historically been treated as phototrophic, studies have shown them to be mixotrophic, capable of both autotrophic and heterotropic behavior, with the ability to take up high molecular weight organic molecules (Legrand and Carlson, 1998; Burkholder et al., 2008; Blossom et al., 2017). This may have occurred after the widespread phytoplankton bloom observed during strong upwelling in January-March 2016.

Acidity levels and CO₂ concentrations affect other biochemical interactions, with impacts on the phytoplankton and microbial community. These operate on scales ranging from the molecular to entire ecosystems. On the smallest scale are processes related to carbon acquisition - i.e., carbon concentration mechanisms with different efficiencies that change with CO₂ concentrations (Reinfelder, 2011; Eberlein et al., 2014) and ocean acidity. Vargas et al. (2017) report high CO₂ concentrations (values of ~400–450 μatm) in river plumes of south-central Chile. Schultz et al. (2013) and Mardones et al. (2016) suggest that increases in ocean acidity and CO₂ concentrations favor the growth of success of dinoflagellates, such as *A. catenella*, in our domain of interest. On the larger ecosystem scales, the effects due to interactions with predators, parasites and pathogens (Smetacek, 2012; Gutiérrez et al., 2016), are not well known for *A. catenella*. Given future expectations, the effects of
increases in CO₂ concentrations and acidity on blooms of specific toxic species need to be examined on all scales, to help understand and predict extended blooms such as observed during 2016.

4. Summary and conclusions

A new Mean Dynamic Topography field provides modifications and additional details to the traditional steady state description of mean poleward flow next to southern Chile, inshore of the SPC bifurcation. In the new mean circulation (Section 3.2, Figs. 1 and 2), equatorward coastal currents extend to 46°S.

When seasonal variability is added (Section 3.3, Fig. 6), the SPC in summer connects to the coast much farther south, near 50°S, flowing into a coastal current that is equatorward north of 48°S and has equatorward currents in some regions as far south as 51°S. In winter, the SPC bends to the north as it approaches the coast near 40°S, joining a coastal current that is poleward south of 37°S. Thus, throughout our domain of interest (38°–46°S), coastal currents alternate between equatorward flow in spring and summer (November-April, maximum in February-March) to poleward in autumn and winter (May-October, maximum in June-July). A primary difference between this depiction and the traditional description of the mean and seasonal surface circulation along southern Chile is a more prominent role of equatorward transport in the mean and seasonal coastal ocean, providing a mechanism to move both water properties and biological populations from the fjord region in the south to the mainland north of Chiloé Island in this transition region.

Scatterometer data provide details of forcing by the mean wind stress (Section 3.2, Fig. 3), which has a strong equatorward component in the north of our domain, changing to a weak poleward component in the south, where the strongest component is toward the east. Seasonal variability (Section 3.3, Figs. 4, 5, and 8) changes the direction of wind stress from equatorward everywhere in our domain during summer to poleward in winter, with a strong eastward component in the south throughout the year. Scatterometer winds within the CIS follow those offshore, with similar strength but with a stronger poleward direction. Global atmospheric reanalysis wind fields such as ECMWF do not capture the strength of the winds nor define the wind stress curl within the CIS and immediately next to the outer coast, although they agree with scatterometer winds to the west of the outer coast. The net result is upwelling in summer along the outer coastal boundary; within the CIS in summer there is upwelling (downwelling) along its eastern (western) margin. These conditions reverse directions in the winter. Wind stress curl enhances the upwelling (with cyclonic curl) and downwelling (with anticyclonic curl) in most coastal regions.

Although the domain between 38° and 46°S has similar characteristics to those in the northern California Current and southern Alaska Gyre, the magnitudes of the seasonal signals (both forcing and response) are less than in the Northern Hemisphere counterparts (Section 3.4). This applies to onshore-offshore differences in SST and SLA, also resulting in weaker changes in the seasonal currents. This is partly due to the weaker seasonal changes in alongshore wind forcing. We hypothesize that it may also be partly due to the impingement of the SPC on the coast, suppressing the upwelling system somewhat through
the phenomenon of ‘geostrophic convergence’, as found in the Benguela Current off South Africa (Veitch et al., 2010).

Interannual variability in sea level between 38° and 46°S is strongly correlated with indexes of ENSO, which represent equatorial SST in the middle of the Pacific and also along the eastern equatorial boundary (Section 3.7, Fig. 22). SLA off southern Chile is directly connected through an oceanic pathway to signals off Ecuador and Perú. This is true in a continuous manner, not just during isolated ENSO ‘events’.

Although alongshore wind stress in the transition zone is not strongly correlated with the ENSO SST indexes, it is directly forced by the atmospheric pressure systems in the mid- to high-latitude southeast Pacific. That system, primarily the anti-cyclonic high pressure cell, has been statistically linked to ENSO and other climate signals (Montecinos and Aceituno, 2003; Garreaud et al., 2008; Rahn, 2012). The effect of the strengthened anti-cyclonic pressure and winds can vary from increases in the strength of equatorward coastal wind stress to the blocking of mid-latitude storms and an overall decrease in wind-forcing. Both effects are demonstrated at different times during the 2014–2016 period by Garreaud (2018) and by the results presented here (Section 3.6, Figs. 5, 16 and 18). The local meridional winds then force the alongshore currents, as affirmed by correlations over the 24-year period (Section 3.7, Fig. 22).

The strongest relationships for regional SST off southern Chile are with the first mode of SLA in the same region and with the MEI and Niño 1 + 2 index. Regional CHL patterns along southern Chile are more difficult to connect to climate indexes and other signals in the same region on the interannual time scales considered here. CHL variability is characterized by much shorter spatial and temporal scales, with phytoplankton doubling times of several days, making it difficult to correlate to lower frequency climate signals. The strongest statistical connection is to sea level: high sea level corresponds to low CHL, a general relationship found elsewhere in the Pacific Ocean.

During 2015–2016, the data document a sequence of conditions that would have favored a bloom of the dinoflagellate A. catenella, such as was observed in April-May 2016 farther north of Chiloé Island than prior to 2016 (Section 3.6, Figs. 17–21). This sequence includes a bloom of A. catenella within the CIS and along the outer coast of Chiloé Island and islands south of the CIS, equatorward currents during January-April that could advect the population of A. catenella north, while keeping it offshore during upwelling, a sudden wind relaxation near the end of April that would have allowed the offshore species to move onshore, extremely calm winds and high insolation during late-April and early-May that would favor growth of dinoflagellates, such as A. catenella. The winds and currents during this sequence are affected by the 2015–2016 El Niño through both the atmosphere and ocean, but neither the presence of A. catenella in the south nor the northward currents are the most unusual aspects of conditions during 2016. The sequence of strong wind-driven upwelling, followed by near calm for an extended period during April-June 2016 (Figs. 5, 16a and 18d) is the most anomalous condition. The wind patterns, in turn, are due to an extremely strong blocking event (Fig. 19) that reached levels during September-October 2015 and April-June 2016 greater than any other periods in the 39-year record. Thus, the
atmosphere provides the most extreme anomalies in the sequence of events surrounding
the northward extension of the toxic algal bloom. Although the interaction of the physical
conditions with the phytoplankton is only hypothesized (lacking in situ data), the results of
the extensive analysis of physical conditions can serve as a hypothetical set of conditions
that might be used to guide more intense sampling in the future.

Supplementary Material

Refer to Web version on PubMed Central for supplementary material.

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gridded (0.25°) SLA fields are from the European Copernicus Marine Environment Monitoring Service (http://
marine.copernicus.eu). Altimeter alongtrack SLA data and environmental corrections are from the Radar Altimeter
Database System (RADS), http://rads.tudelft.nl/). The CNES MDT fields come from the AVISO site at https://
www.aviso.altimetry.fr/.

QuikSCAT data vers 3.0 and 3.1 data come from the JPL PODAAC web site (https://podaac.jpl.nasa.gov).
QuikSCAT data vers 4.0 are available from Bryan Stiles (bryan.w.stiles@jpl.nasa.gov); ECMWF wind and sea
level pressure fields are obtained from https://rda.ucar.edu/; they are also available from the ECMWF web site
(http://apps.ecmwf.int/datasets/).

MODIS Aqua SST data are from JPL PODAAC at (ftp://podaac-ftp.jpl.nasa.gov/allData/midis/L3/aqua/11um/
v2014.0/4km/monthly.

MODIS Aqua Chlorophyll-a pigment concentrations are from https://oceandata.sci.gsfc.nasa.gov/, doi:
10.5067/AQUA/MODIS/L3M/CHL/2014.

NOAA Reynolds OI SST are from the JPL PODAAC web site: https://podaac.jpl.nasa.gov/dataset/AVHRR_OI-
NCEI-L4-GLOB-v2.0.

NCEP Data for upper level winds: https://doi.org/10.1175/jcli-d-12-00823.1, Saha, Suranjana, and Coauthors, 2014:
The NCEP Climate Forecast System Version 2. J. Climate, 27, 2185-2208.

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Fig. 1.
Mean Dynamic Topography from the CNES 2013 processing, derived from altimeter SLA, satellite gravity measurements, and in situ data. Heights are shown as colors on a 0.25 degree grid; geostrophic velocities calculated from the heights are shown on the same grid (black vectors, every other grid point). White vectors are two degree averages of all velocities (note the difference in scale).
Fig. 2.
Mean Dynamic Topography and geostrophic velocities (from the CNES 2013 processing), as in Fig. 1. White vectors are one degree average of the velocities (note the difference in scale). Fields within the CIS have greater uncertainty.
Fig. 3.
(a) Mean (1999–2009) QuikSCAT wind stress (vectors), showing every fourth grid point and wind stress curl at each grid point (color); (b) Mean MODIS Aqua SST (2003–2016, color) and mean (1993–2016) absolute altimeter geostrophic velocities (MDT included), as in Fig. 2; (c) Mean (1998–2016) Globcolour chlorophyll-a pigment concentrations (color) and mean (1993–2016) ECMWF wind stress.
Fig. 4.
Average monthly wind stress from QuikSCAT (black vectors) over wind stress curl (colors). (a) January; (b) April; (c) July; (d) October. The long-term mean is subtracted for (e) January and (f) July.
Fig. 5.
(a) Seasonal spatial pattern and mean monthly time series (two cycles) for the first EOF mode of the monthly mean magnitude of the high-frequency (HF) filtered total wind stress (periods less than 7 days); (b) Monthly time series for the first mode of the mean monthly mean magnitude of the HF total wind stress from ECMWF winds with the seasonal cycles included. The spatial pattern is identical to that in the top panel. (c) The same as in (b) after first removing seasonal cycles from the monthly mean magnitude of the HF total wind stress from ECMWF winds. The spatial pattern (not shown) is nearly identical to that shown in the top panel, except that the maximum value on the color bar is 1.40. The complete 1993–2016 period is used in all calculations. Red crosses are placed on the May 2016 values.
Fig. 6a.
Mean monthly SLA + MDT and geostrophic velocities for (left) January and (right) July. Adding the MDT creates absolute geostrophic velocities. During austral summer (left) velocities next to the coast are consistently equatorward as far south as ~48°S. During austral winter (right), poleward velocities are found next to the coast as far north as 37°S.
Fig. 6b.
Mean monthly SLA and geostrophic velocity anomalies (long-term mean removed) for (left) January and (right) July. The most coherent changes in heights and velocities occur within several hundred kilometers of the coast between ~34°S. and 54°S, as the sea level drops (rises) in austral summer (winter).
Fig. 7.
Mean monthly MODIS SST (2003–2016, color) and mean monthly altimeter absolute
geostrophic velocity vectors (also 2003–2016). (a-d) The long-term mean has not been
removed. (e and f) The long-term mean has been removed.
Fig. 8.
Mean monthly Globcolour CHL (1998–2016, color) and ECMWF (1998–2016) wind stress vectors. (a-d) The long-term mean has not been removed. (e and f) The long-term mean has been removed.
Fig. 9. (a-c). Time series of seasonal cycles (mean values removed). Locations for the “Offshore” and “Inshore” and “CIS” are shown in Fig. 9g. For each location, the left panels show SLA, SST and Chlorophyll-a concentrations; the right panels show altimeter north-south geostrophic surface velocity, and north-south wind stress (ECMWF and QuikSCAT). Northward wind stress and currents are shown as positive (standard notation). (e-g) Time series of seasonal cycles (mean values removed) for the two locations within the CIS. Note the different scales for ECMWF and QuikSCAT wind stress. (g) Locations of the data points.
Fig. 10.
Spatial patterns and monthly time series of ECMWF Mode 1 meridional wind stress ($\tau_y$), with (a) seasonal cycles retained (no filter), and (b) seasonal cycles removed and low-pass filtered (6-month half-power). The red lines correspond to periods associated with anomalous warm periods (Fig. 22) (even if too weak to be formally classified as an El Niño). Most are also ‘high-SLA events,’ as discussed in the text.
Fig. 11.
The first two modes of the altimeter SLA, from the gridded AVISO data set, during 1993–2016, with the seasons removed and low-pass filter applied. Red lines along the time line identify ‘warm/high-SLA’ periods in Fig. 22.
Fig. 12.
The first two modes of the altimeter meridional velocity, from the gridded AVISO data set, during 1993–2016, with the seasonal cycles removed and low-pass filter applied. Red lines along the time line identify ‘warm/high-SLA’ periods in Fig. 22.
Fig. 13.
The first two modes of SST, from the Reynolds OI 0.25° gridded data set, with the seasonal cycles are removed and the low-pass filter applied. Red lines along the time line identify ‘warm/high-SLA’ periods in Fig. 22.
Fig. 14.
The first two modes of surface chlorophyll concentration, from the Globcolour combined
sensor data set during mid-1997 through 2016, with the seasons removed and low-pass filter
applied. Red lines along the time line identify ‘warm/high-SLA’ periods in Fig. 22.
Fig. 15.
Hovmöller plots (time-latitude) for the non-seasonal sea level anomaly in the 75 km next to the western coast of South America from the alongtrack altimeter data, using just the reference mission tracks. The periods 2013–2016 and 1995–1998 are shown after removing the long-term mean, seasonal cycles and a trend of 1.5 cm/decade at each point.
Coastal SLA, meridional wind stress and El Niño indexes. (Black with Yellow/Blue fill): Averaged median coastal SLA values along tracks crossing the coast in latitude bands. (Blue) Nino 3.4 and Nino 1 + 2 indexes are shown in the top two panels. (Red) ECMWF Era-interim meridional wind stress values from the same locations as SLA are averaged within the higher latitude bands. Negative values (downwelling favorable) are plotted toward the top of the figure to match positive SLA anomalies. The 1993–2016 seasonal cycles are removed from the SLA and wind stress.
Fig. 16b.
Cross-track velocity next to the coast, meridional wind stress and El Niño indexes. (Black with Yellow/Blue fill): Geostrophic surface velocities in the 150 km next to the coast. (Blue): Nino 3.4 and Nino 1 + 2 indexes are shown in the top two panels. (Red) ECMWF meridional wind stress values from the same locations as velocities within each latitude band. Negative values (poleward) are plotted toward the top of the figure. The 1993–2016 seasonal cycles are removed from the data except in the bottom panel, where seasonal cycles are retained.
Fig. 17.
(a-c). Maps of SLA (color) and geostrophic surface velocity anomalies (long term mean removed, seasonal cycle included) between 39° and 44°S during January-April 2014–2016.
(d-f). Maps of SLA and geostrophic surface velocity anomalies (long term mean removed, seasonal cycle included) during the peaks of the 2015 El Niño (May, September) and months before and after (July, November), with the same months in 2014 and 2016.
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Fig. 18.
(a and b) Time series of ECMWF meridional wind stress during January-March 2015 (top) and 2016 (bottom). Positive wind stress (equatorward) is plotted toward the top of the page (standard notation). (c and d) Time series of ECMWF meridional wind stress during April-June 2015 (top) and 2016 (bottom). As in Fig. 18a and b.
Fig. 19. The Blocking Index during 2015–2016. (a) Time longitude plot of the Blocking Index between 120° and 50°W using five-day averages of the index. Time proceeds downward during each year/panel. (b) Time series of the Blocking Index at 90°W (blue) and 80°W (red). Note the coast is at approximately 74°W in our region. Values above 20 m s\(^{-1}\) during late-August to mid-October 2015 and late-April to early-June 2016 are in the top 10% of the values during the 1979–2016 period.
Fig. 20.
(a-d) Monthly maps of satellite (MODIS-aqua) surface chlorophyll during February-May 2016. Patterns in individual months are clearer in MODIS fields, although values for MODIS processing are higher than for the Globcolour processing.
Fig. 21.
(a-d) Maps of SST (top, a and b) and SST anomaly (bottom, c and d, seasonal cycle removed) during March and May 2016.
Fig. 22. Monthly time series for ENSO indexes, time series from SLA, $\tau_y$, and $V_y$ EOFs, along with time series from coastal alongtrack SLA in two latitude bands. All time series except MEI are smoothed with a loess filter with a 3-month half-span. Red lines along the time lines identify ‘warm/high-SLA’ periods based on ENSO indexes and the SLA EOF 1. Note that negative velocities and stresses (poleward) are plotted toward the top of the page in panels c and d.
Table 1

Characteristics of the primary data sets used in the analysis.

| Variable                        | Period     | Grid resolution | Temporal resolution | Notes                                      |
|---------------------------------|------------|-----------------|---------------------|--------------------------------------------|
| Gridded AVISO Altimeter SLA     | 1993–2016  | 0.25°           | Daily/Monthly       | Anomaly relative to 1993–2012              |
| MDT - Mean Dynamic Topography   |            | 0.25°           | Long-Term Mean      |                                            |
| RADS Alongtrack SLA             | 1993–2016  | ~6 km alongtrack| ~10-day             | Anomaly relative to 1993–2012              |
| ECMWF Vector Winds              | 1993–2016  | ~0.67°          | 6-Hr, Monthly       | ERA Interim Reanalyses                     |
| QuikSCAT Vector Winds           | 11/99–10/09| 0.1°            | 1–2 Day Swath/Monthly| Time Average of Gridded Swaths            |
| QuikSCAT Curl τ                 | 11/99–10/09| 0.1°            | 1–2 Day Swath/Monthly| Time Average of Gridded Swaths            |
| Blocking Index                  | 1979–2017  | Large-scale Index| 5-day: 3 month running means | 25°–60°S, 500mb winds |
| MODIS Aqua SST                  | 2003–2016  | 4 km            | Monthly             | Monthly Composite                          |
| Reynolds SST                    | 1993–2016  | 0.25°           | Monthly             | OI fields from IR, Microwave, In Situ      |
| MODIS Aqua CHL                  | 2003–2016  | 4 km            | Monthly             | Monthly Composites                        |
| Globcolour CHL                  | 9/97–2016  | 4 km            | Monthly             | Multi-Satellite                            |
Table 2: Monthly Variances of Each Time Scale

| Parameter                     | 100% Total | Seasonal Variance | Interannual Variance | Intra-Seasonal Variance |
|-------------------------------|------------|-------------------|-----------------------|-------------------------|
| Gridded Altimeter SLA         | 13.18      | 2.98              | 7.46                  | 2.74                    |
| Gridded Altimeter V_y        | 28.90      | 5.50              | 19.3                 | 6.12                    |
| ECMWF F timetable            | 2.17E-03   | 3.2E-03           | 8.0E-04               | 3.0E-04                 |
| ECMWF F y                    | 1.34E-03   | 7.93E-04          | 5.2E-06               | 8.5E-06                 |
| QuikSCAT F x                  | 1.49E-03   | 4.2E-04           | 9.2E-06               | 5.0E-06                 |
| QuikSCAT F y                 | 1.93E-03   | 2.4E-05           | 1.8E-06               | 3.4E-06                 |
| Reynolds SST                  | 4.33       | 3.04              | 2.7E-07               | 1.1E-07                 |
| Globocor CHL                  | 1.08       | 0.45              | 4.1E-06               | 9.1E-06                 |

Variances and % variance: Variance units are cm$^2$ and cm$^2$ s$^{-2}$ for SLA and V_y, respectively; N$^{-2}$ m$^{-4}$ and N$^{-2}$ m$^{-6}$ for wind stress and wind stress curl, respectively; (°C)$^2$ for SST and (mg m$^{-3}$)$^2$ for chlorophyll-a concentrations.
Table 3

Latitudes (°S) and region names from north to south (rows) showing the absence/presence of *A. catenella* in 2013–2016 (Presence is 1–2 cells). White cells indicate no periodic monitoring sampling sites that shown either pooled by year during summer-autumn periods in 2013–2014 (columns), or pooled by month in late 2015, early 2016 and dates of sampling days* in April and May 2016 (0 = not present in blue; 1 = only observed at one location in the region in yellow; M = observed at multiple locations in the region in red). *these sampling dates correspond to extraordinary samplings sites undertaken because of the exceptional bloom reported by animal deaths.

| Year | 2013 | 2014 | 2015 | 2016 | April | May |
|------|------|------|------|------|-------|-----|
|      | 10   | 11   | 12   | 1    | 2     | 3   | 8   | 19 | 21 | 2 | 6 | 17 |
| Month |      |      |      |      |       |     |     |    |    |    |    |    |
| Dates |      |      |      |      |       |     |     |    |    |    |    |    |
| Latitude | Regions |      |      |      |       |     |     |    |    |    |    |    |
| 39–40° | Los Rios       | 0 | 0 | 1 | M | 0 |
| 40–41.5° | Los Lagos     | 0 | M | M | M | M | 1 |
| 41.5–42° | Chacao Ch Chiloé | 0 | 0 | 0 | 0 | 0 | 0 | M | M | M | M | M |
| 42.5° | West Coast Chiloé  | 1 |     |     |     | 1 | 1 | 1 | 1 |
| 42° | North CIS    | 1 | 0 | 0 | M | M | M | M | M | M | M | M |
| 43° | South CIS    | M | M | M | M | M | M | M | M | M | M | M |
| 43.5° | South Tip Chiloé | M | M | M | M | M | M | M | M | M | M | M |
| 44° | South of CIS-Aysen | M | M | M | M | M | M | M | M | M | M | M |
Correlations between selected variables. The rows correspond to time series for the first or second EOF of the variables shown, except for the last row, which is the coastal alongtrack SLA between 35° and 45°S. The columns correspond to similar variables and three El Niño Indexes: Niño 3.4, the MEI, and Niño 1 + 2. All monthly time series except MEI are smoothed with a loess filter with a 3-month half-span. Estimates of decorrelation times (6, 9, or 12 mo) from autocorrelations are used to specify approximate 95% and 99% correlation levels, shown in the last column. Values above the 95% and 99% levels are shown in italics and bold italics, respectively. The time series extend from 1993 to 2016, except for CHL (1998–2016).

|       | N 3.4 | MEI  | N 1+2 | SLA-1 | SLA 0–15 | SLA 35–45 | $\tau_{y\cdot -1}$ | $V_{y\cdot -1}$ | 95%, 99% |
|-------|-------|------|-------|-------|----------|-----------|-------------------|----------------|----------|
| SLA-1 | 0.71  | 0.75 | 0.67  | 1.00  | 0.72     | 0.74      | −0.19            | −0.33          | 0.39, 0.50 |
| SLA-2 | 0.33  | 0.27 | 0.32  | 0.01  | 0.45     | 0.65      | −0.44            | −0.92          | 0.32, 0.44 |
| $\tau_{y\cdot -1}$ | −0.10 | −0.10| −0.14 | −0.19 | −0.10   | −0.27     | 1.00             | 0.50           | 0.28, 0.36 |
| $V_{y\cdot -1}$    | −0.37 | −0.33| −0.33 | −0.45 | −0.66   | 0.50      | 1.00             | 0.28, 0.36    |           |
| SST-1  | 0.25  | 0.32 | 0.38  | 0.57  | 0.31     | 0.40      | −0.23            | −0.11          | 0.32, 0.44 |
| SST-2  | 0.24  | 0.23 | 0.13  | 0.16  | 0.16     | 0.32      | −0.46            | −0.41          | 0.32, 0.44 |
| CHL-1  | −0.20 | −0.24| −0.14 | −0.25 | −0.23   | −0.39     | 0.20             | 0.27           | 0.28, 0.36 |
| CHL-2  | −0.04 | 0.07 | −0.11 | 0.02  | −0.11   | −0.17     | 0.09             | 0.19           | 0.32, 0.44 |
| SLA 35–45 | 0.67 | 0.67 | 0.65  | 0.74  | 0.76     | 1.00      | −0.27            | −0.66          | 0.32, 0.44 |