Supplementary Material 3. Literature Review for Exposure Factors Not Included in Downscaled Models

Winds (Upwelling)

Upwelling occurs when equatorward alongshore winds drive offshore Ekman transport, and surface waters are replaced with nutrient-rich deep waters. The input of nutrient-rich waters to the euphotic zone promotes high levels of primary productivity and Eastern Boundary Upwelling Systems (EBUS) such as the California Current System (CCS) support nearly 20% of the global fish catch despite occupying less than 1% of the global ocean (Mote and Mantua, 2002; Pauly and Christensen, 1995).

Upwelling intensity is affected by basin-scale modes of climate variability in the north Pacific (e.g., ENSO, PDO and NPGO). Positive Pacific Decadal Oscillation (PDO) and El Niño Southern Oscillation (ENSO) tend to be correlated with weaker nearshore upwelling (<50 km from shore) and stronger offshore upwelling (>50 km from shore), while the North Pacific Gyre Oscillation (NPGO) is associated with the opposite (Chhak and Di Lorenzo, 2007; Di Lorenzo et al., 2008; Macias et al., 2012; Jacox et al., 2014). The PDO influence is stronger in the northern CCS, while the NPGO influence is stronger in the southern CCS.

On longer timescales, several possibilities have been presented regarding changes in CCS upwelling. Bakun (1990) suggested that global warming would heat the continents faster than the oceans, steepening cross-shore air pressure gradients and intensifying upwelling favorable winds. Several modeling studies have forecast intensified upwelling winds under climate change (Snyder et al., 2003; Auad et al., 2006), and a meta-analysis of the literature found support for increased upwelling in the CCS since 1950 (Sydeman et al., 2014). However, the time series to date are too short to differentiate between natural climate variability from anthropogenic forcing (Sydeman et al., 2014). Basin-scale climate variability (ENSO, PDO, and NPGO) in the CCS may dominate on 30-50 year time scales, making it difficult to distinguish climate change driven upwelling from natural variability (Jacox et al., 2015).

The most recent studies suggest that the Bakun hypothesis is oversimplified and changes in upwelling will be season, region, and latitude dependent, with the CCS experiencing increased upwelling to the north during the spring, decreased summer upwelling to the south, and a decrease in upwelling in the winter (Figure 1; Wang et al., 2015; Rykaczewski et al., 2015). Rykaczewski et al. (2015) predicted an average decrease of 8% (+/- 10% SD) in upwelling favorable winds across the CCS. The Wang et al. (2015) and Rykaczewski et al. (2015) studies are the first to use an ensemble of over 20 coupled atmosphere-ocean general circulation models (AOGCMs), developed in association with the Intergovernmental Panel on Climate Change; therefore, they are currently the most reliable predictions of climate change impacts on upwelling favorable winds in the CCS. However, uncertainty in the predictions still exists because the models included in the ensemble are global models, making it difficult to resolve local dynamics in the CCS.

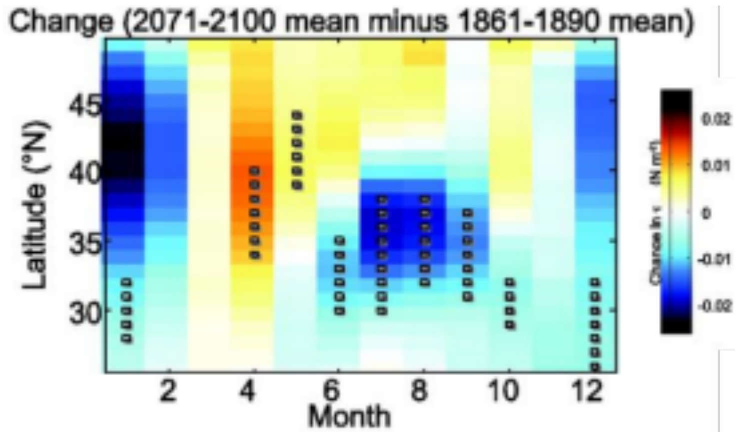

Figure 1: Hovmöller diagram of the change in upwelling favorable winds between the two periods (2071-2100 mean minus 1861-1890 mean) for the CCS (Rykaczewski et al., 2015)

Impacts of upwelling on the ecosystem will depend on their interaction with other changes in the water column. If upwelling intensifies, increased nutrient input could stimulate added productivity to a point, however too much wind will increase turbulence and offshore transport of surface waters and reduce productivity nearshore. Conversely, either reduced upwelling or upper ocean warming (and resultant increased water column stratification) would reduce the efficacy of coastal upwelling for delivering nutrients to the euphotic zone. Such a scenario has been invoked to explain ecosystem changes in the latter half of the 20th century, where an abrupt increase in sea surface temperature and water column stratification around 1977 reduced the input of nutrients to the surface layer and was linked to widespread changes in the zooplankton community, decreased abundance of larval fish and seabirds, and reduced commercial fish landings.(Roemmich and McGowan, 1995; Palacios et al. ,2004; Di Lorenzo et al., 2005). Finally, changes in basin-scale circulation may result in dramatic increases in the nutrient content of upwelling source waters (Rykaczewski and Dunne, 2010), in which case the coastal zone is likely to be highly productive regardless of how the wind changes.

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Subsurface Oxygen

Dissolved oxygen (DO) is fundamental for all aerobic life and it plays a direct role in biogeochemical cycling of nutrients in the ocean (Keeling et al., 2010). Once DO concentrations drop below a certain level organisms suffer from a variety of stresses; this oxygen deficiency is termed hypoxia. Waters are considered hypoxic when O_2 concentrations are below 60 μmol kg^{-1} and the depth of hypoxia is the shallowest depth at which waters become hypoxic (Bograd et al., 2008). DO concentrations are influenced by air-sea oxygen exchange, circulation, ventilation, production, and respiration (Bograd et al., 2008). DO levels are predicted to decline with climate change because oxygen is less soluble in warm water and because increased stratification reduces the exchange of surface oxygen into the ocean interior (Keeling et al., 2010).

Hypoxic conditions occur at mid-depths (250-400 m) across the North Pacific. In fact, the CCS has an acute and extensive oxygen minimum zone (OMZ), the zone in which oxygen saturation is at its lowest (Keeling et al., 2010; Moffitt et al., 2015). Therefore, even small declines in DO may create hypoxic conditions in species' critical habitats. For example, in 2002 and 2006 the central Oregon Coast was exposed to severe hypoxia because upwelling transported DO depleted deep water onto the inner shelf (<100 m), leading to massive fish and crab die-offs (Grantham et al., 2004; Chan et al., 2008).

Over the past 50 years there has been an expansion in the area and volume of OMZs in the open North Pacific (Keeling et al.2010). Similar declines have been observed in the CCS, with studies focusing on the central Oregon coast and the Southern California Bight. In the North Pacific the depth of hypoxia has shoaled from 400 to 300m from the 1950s to present (Whitney et al., 2007). The hypoxic boundary in Southern California shoaled by 45 (offshore) to 90 (inshore) m from 1985 to 2007 (McClatchie et al., 2010; Bograd et al., 2008). On the Oregon Coast, there was no evidence of severe inner-shelf hypoxia until 2000 (Chan et al., 2008). The results from studies in the open North Pacific and the coastal CCS are listed in Table 1.

Reference	**Location**	**Latitude (degrees)**	**Longitude (degrees)**	**Depth (m)**	**Dates**	**O2 change (μmol kg−1)**
Open North Pacific						
Whitney et al. (2007)	Station P	50N	145W	100-400	1956-2006	-20 to -35
Mecking et al. (2008)	Eastern North Pacific	27-55N	152W	150-550	1984-2006	-10 to -25
Coastal CCS						
Whitney et al. (2007)	Station P4	49N	127W	150	1987-2006	-23
Chan et al. (2008)	Oregon Coast	42-46N		<100	1950-2006	~ -40
Bograd et al. (2015)	Southern CCS	30-33N	118-124W	200-300	1984-2012	-11 to -39
McClatchie (2010)	Southern CCS	32-34N	117-119W	100-350	1950-2007	-13 to 6.5

Table 1: Studies documenting ocean dissolved O_2 changes

The observed declines in CCS DO concentrations are caused by both basin-scale dynamics and local factors. The sampling locations with the largest declines were at the core of the southward flowing California Current and at the core of the deeper, northward flowing California Undercurrent (Whitney et al., 2007, Bograd et al., 2008, Chan et al., 2008). Therefore, declining DO concentrations in source waters (the North Pacific Subtropical Gyre and the eastern tropical Pacific) that are advected into the two currents influence local DO trends (Bograd et al., 2015). The primary cause of reduced DO concentrations in source waters is because of a global reduction in vertical mixing with increasing stratification, a trend that is also occurring within the CCS (Bograd et al., 2008; Keeling et al., 2010; Bakun et al., 2015).

Climate models predict a global decrease in DO concentrations of 1-7% by 2100 (2-12 μmol kg^{-1}; Keeling et al., 2010; Oschlies et al., 2008). An earth system model predicted an 18% decline in CCS DO concentrations by 2100 (Rykaczewski and Dunne, 2010). It is expected that the declining DO trends in the CCS will continue over the next few decades with climate change, leading to increased occurrences of hypoxia (Rykaczewski and Dunne, 2010; Stramma et al., 2008; Bakun et al., 2015).

As oxygen poor areas expand, favorable habitat areas for many pelagic and benthic fish species will be compressed (Stramma et al., 2010; Bograd et al., 2008). For example, during the 2002 hypoxic event on Oregon's inner-shelf there were unusually large aggregations of fish in very shallow waters, indicating that they were seeking refuge in atypical habitats (Grantham et al., 2004). This expansion of oxygen poor areas could lead to cascading effects including shifts in community organization and animal distribution (Stramma et al., 2010; Bograd et al., 2008).

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Sea Level Rise

Sea level rise occurs with climate change for two principal reasons: 1) ocean warming causes the expansion of seawater and 2) the melting of land ice serves as a freshwater influx to the ocean (NRC, 2012). The two main ice sheets that have the potential to cause significant sea level rise are the Greenland Ice Sheet, which has the potential to raise the world's oceans by 6m, and the Antarctic Ice Sheet, which could raise the global ocean level by 58m (Viñas and Rasmussen, 2016). From 1900 to 1930 the rate of sea level rise was 0.6 mm/yr, the rate increased to 1.4 mm/yr between 1930 and 1992, and currently the rate of sea level rise is 3.3mm/yr (1992-2015; Hansen et al., 2016).

The Intergovernmental Panel on Climate Change's (IPCC) fifth assessment report projected that global sea levels will rise between 0.28 and 0.97 meters by 2100 (IPCC, 2013). Recently, Hansen et al. (2016) projected an eventual sea level rise of 5-9m, with a near certainty of a multimeter sea level rise by 2100. It is clear that researchers agree there will be global sea level rise, but there is uncertainty in how rapidly it will occur, with predictions ranging from 0.28 to 9 meters by 2100. The majority of this uncertainty stems from whether there will be nonlinear ice mass loss in Greenland and the Western Antarctic (NRC, 2012).

Projections of sea level rise along the West Coast of the United States have also been estimated. In 2012 the National Research Council (NRC) predicted future sea level rise in California, Oregon, and Washington for the years 2030, 2050, and 2100 (NRC, 2012). The NRC included in their models how sea level rise is affected by regional factors including: 1) ENSO events (El Niño leads to higher sea levels and La Niña to lower), 2) the rising and sinking of land along the coast, and 3) the proximity of Alaska's glaciers which exert a gravitational pull on sea water. The results are listed in Table 1. Sea level rise south of Cape Mendocino are similar to global projections, where the coast is sinking 1 mm/yr; while north of Cape Mendocino the projections are lower because much of the coast is rising (around 1.5-3.0 mm/yr), causing seismic strain. If there was a large earthquake north of Cape Mendocino to reduce the strain, sea level could rise an additional 1-2m above these projections.

	2030	**2050**	**2100**
South of Cape Mendocino	.04 – .3 meters	.12 – .61 meters	.42 – 1.67 meters
North of Cape Mendocino	-.04 – 23 meters	-.03 – .48 meters	.10 – 1.43 meters

Table 1: Sea level rise projections for the West Coast of the U.S. relative to the year 2000.

An increasing rate of sea level rise will worsen the impacts of high tides, storms, and floods (Cayan et al., 2008). Furthermore, intense storms are expected to become more frequent with climate change, leading to more frequent sea level extremes (Cayan et al., 2008; Hansen et al., 2016). For example, the incidence of extreme water heights in San Francisco Bay is predicted to increase from 9 hours per decade to hundreds of hours per decade by 2050 (NRC, 2012). Sea level rise will result in salinity intrusion into delta areas and will damage marginal ecosystems (e.g., seagrass beds and salt marshes) that many juvenile fish rely on (Cayan et al., 2008).

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Phenology of Upwelling

A change to the phenology of upwelling refers to potential shifts in the seasonal cycle of upwelling (Bograd et al., 2009). In the CCS there are two distinct seasonal ‘modes’ of upwelling; the more intense spring-summer upwelling (April through September) and winter upwelling (January through March; Black et al., 2011). The northern CCS (North of Cape Mendocino) is characterized by seasonal upwelling; in the central and southern CCS (Cape Mendocino to Point Conception and South of Point Conception respectively) upwelling occurs in all seasons (Snyder et al., 2003). It is important to note that in the northern region there is also winter downwelling, which is affected by changes in upwelling favorable winds (Rykaczewski et al., 2015).

From 1967 to 2007 there has been a trend toward a later spring transition date (i.e. the date of the onset of spring-summer coastal upwelling) and shorter upwelling season in the northern CCS and a longer upwelling season in the southern CCS, which may be a sign of the direction of future change (Bograd et al., 2009). Additionally, upwelling is delayed and weakened in the central CCS during El Niño years (Bograd et al., 2009); and the frequency of extreme El Niño events is predicted to double with climate change (Cai et al., 2014). Furthermore, positive phases of the NPGO appear to be associated with an earlier start to the upwelling season (spring transition) in the central CCS (Chenillat et al., 2012). However, the current set of climate models are not able to predict changes to the NPGO, so it is difficult to predict how the onset of the spring-summer upwelling period will be affected.

There is no consensus among climate change models on how climate change will affect the timing of the spring transition date. The first study to model potential shifts in the upwelling season projected no change (Mote and Mantua, 2002). A study published the following year predicted that upwelling favorable winds would shift to later in the year, leading to a later spring transition date (Synder et al., 2003). Most recently, Wang et al. (2015) predicted an expansion of the upwelling season by 1-2 days per decade north of 40° latitude. Using a suite of global climate models forced by the RCP8.5 emissions scenario, Rykaczewski et al. (2015) found projected changes in upwelling favorable winds over the 21st century to be latitude and season dependent. In the northern CCS, modest increases in upwelling are projected throughout the upwelling season. In the central CCS, upwelling in April is projected to increase, while June-September upwelling is projected to decrease in both the central and southern CCS.

Both the summer and winter 'modes' of upwelling are biologically important, by supporting high levels of phytoplankton and zooplankton and subsequently higher trophic levels (Bakun et al., 2015). Many organisms in the CCS have life histories traits (foraging, reproduction, migration) adapted to the current phasing of upwelling, changes to the phenology may significantly affect the marine ecosystem (Bograd et al., 2009). For example, a delayed or early start to winter upwelling or downwelling and spring-summer upwelling may result in a temporal mismatch between predator needs and prey availability (Bograd et al.2009; Bakun et al., 2015). In 2005 there was a one month delay in the spring transition date, leading to a late spring bloom in phytoplankton and zooplankton. This change disrupted some higher trophic level species by decoupling food availability from consumer demands (Barth et al., 2007). Early season mussel and barnacle recruitment was reduced by 83% and 66% respectively and there was reproductive failure of the Cassin's auklet (*Ptychoramphus aleuticus*) in both central California and southern British Columbia (Barth et al., 2007; Sydeman et al., 2006). However, mussel and barnacle recruitment rebounded (returned to normal and 38% less than normal, respectively) due to increased upwelling-favorable winds later in the season.

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Ocean Currents

There are three principal circulation features in the CCS: the shallow equatorward flowing California Current (CC), the deep poleward flowing California Undercurrent (CU), and the winter-time poleward flowing Davidson Current (DC) (King et al., 2011). The CC is part of the clockwise flowing North Pacific Gyre, the largest of the oceanic gyres, which is formed by four major currents, the North Pacific Current to the north, the California Current to the east, the North Equatorial Current to the south, and the Kuroshio Current to the west. The eastward flowing North Pacific Current feeds the CC with cool, low-salinity, and nutrient rich waters and the slow, broad CC moves the waters equatorward year-round (Figure 1; Checkley and Barth, 2009). The CC flows about 50-1000 km offshore in the upper 50 meters (King et al., 2011). Directly south of Point Conception a portion of the CC turns north and becomes the Southern California Countercurrent, which in summer does not make it completely north and becomes the biologically rich, self-contained Southern California Eddy (Meredith and Hogg, 2006; King et al., 2011). The narrow and deep (100-300 meters) CU also flows poleward bringing high salinity, warm, low oxygen water from Baja California to Vancouver Island (King et al., 2011). Finally, the DC also flows poleward from Point Conception to Vancouver Island in the winter.

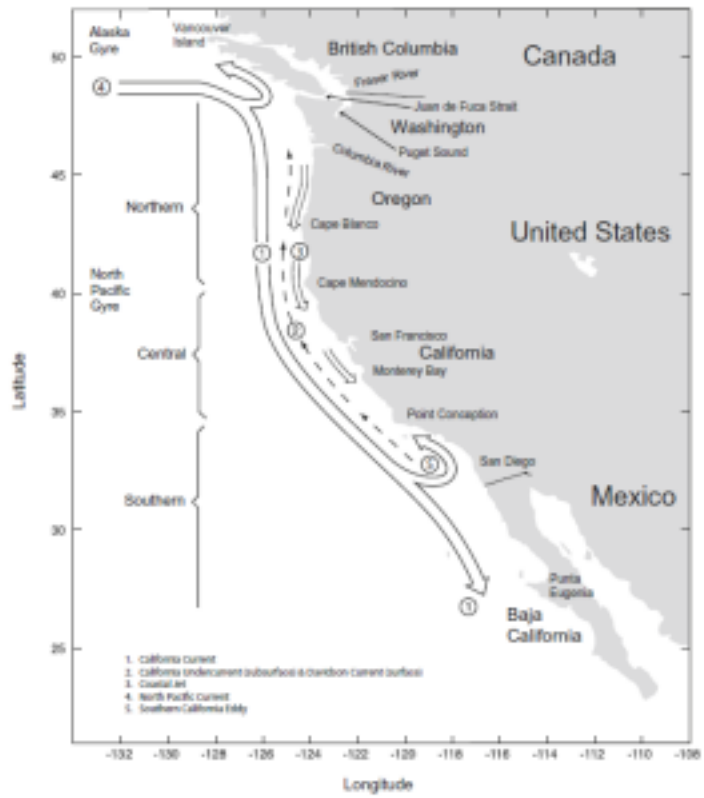

Fig. 1. Map of the California Current System. Major regions, currents, and geographic features are shown. The California Current derives from both the North Pacific Current, to the north, and the coastal jet, to the east (after Checkley et al., 2009b).

Figure 1. Map of the California Current System. Major regions, currents, and geographic features are shown. The California Current derives from both the North Pacific Current to the north, and the coastal jet, to the east (after Checkley and Barth, 2009).

Equatorward flow in the CC is driven by wind forcing; there is a positive relationship between the strength of CC flow and negative wind stress curl (Checkley and Barth, 2009). Therefore, changes in flow are driven by fluctuations in the strength of wind forcing (Cummins and Freeland, 2007). Increases in wind speed could increase current strength and subsequently increase eddy activity (Meredith and Hogg, 2006). The strength of the CC is also driven by fluctuations in the strength of the NPC, which affects the amount of water transported into the Alaskan Gyre and into the CC (Cummins and Freeland, 2007). Furthermore, CC strength is affected by El Niño Southern Oscillation (ENSO) and Pacific decadal modes. The positive (negative) phases of ENSO and the PDO (NPGO) are associated with a weaker CC, and vice versa (Chelton et al., 1982; King et al., 2011). From the late 1970s to the early 1990s as temperature increased the CC weakened and moved shoreward and the CU also weakened; a temperature decrease in the late 1990s was accompanied by a stronger and broader CC as well as a stronger CU (Chavez et al., 2003). However, there has not been a clear trend of a strengthening or weakening of currents in the CCS. There has been a small but statistically significant increase in eddy kinetic energy in the Eastern North Pacific from 1992 to 2010 (O'Donnoll, 2015).

The combination of substantial natural variability and high uncertainty of predicted nearshore zonal winds in global climate models makes it difficult to determine how the North Pacific Gyre, CCS currents, and eddies will be impacted by climate change (King et al., 2011). Currents in the CCS can change location and intensity; for example the Northern Hemisphere Hadley Cell is predicted to expand poleward, which may result in a poleward displacement of zonal currents

(Bakun et al., 2015; Francis et al., 1998). However, predictions indicate that there will not be dramatic change in the structuring of the major oceanic gyres and currents in the CCS (Stouffer et al., 2006; King et al., 2011).

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