A regional hindcast model simulating ecosystem dynamics, inorganic carbon chemistry and ocean acidification in the Gulf of Alaska

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Abstract. The coastal ecosystem of the Gulf of Alaska (GOA) is especially vulnerable to the effects of ocean acidification and climate change that can only be understood within the context of the natural variability of physical and chemical conditions. Controlled by its complex bathymetry, iron enriched freshwater discharge, and wind and solar radiation, the GOA is a highly dynamic system that exhibits large inorganic carbon variability from subseasonal to interannual timescales. This variability is

5 poorly understood due to the lack of observations in this expansive and remote region. To improve our conceptual understanding of the system, we developed a new model set-up for the GOA that couples the three-dimensional Regional Oceanic Model System (ROMS), the Carbon, Ocean Biogeochemistry and Lower Trophic (COBALT) ecosystem model, and a high resolution terrestrial hydrological model. Here, we evaluate the model on seasonal to interannual timescales using the best available inorganic carbon observations. The model was particularly successful in reproducing observed aragonite oversaturation and

10 undersaturation of near-bottom water in May and September, respectively. The largest deficiency of the model is perhaps its inability to adequately simulate spring time surface inorganic carbon chemistry, as it overestimates surface dissolved inorganic carbon, which translates into an underestimation of the surface aragonite saturation state at this time. We also use the model to describe the seasonal cycle and drivers of inorganic carbon parameters along the Seward Line transect in under-sampled months. As such, model output suggests that a majority of the near-bottom water along the Seward Line is seasonally under-

15 saturated with regard to aragonite between June and January, as a result of upwelling and remineralization. Such an extensive period of reoccurring aragonite undersaturation may be harmful to CO_2 sensitive organisms. Furthermore, the influence of freshwater not only decreases aragonite saturation state in coastal surface waters in summer and fall, but simultaneously also decreases surface pCO_2, thereby decoupling the aragonite saturation state from pCO_2. The full seasonal cycle and geographic extent of the GOA region is undersampled, and our model results give new and important insights for months of the year and

20 areas that lack *in situ* inorganic carbon observations.

1 Introduction

The Gulf of Alaska Large Marine Ecosystem (GOA-LME) is home to highly productive commercial and subsistence fisheries including salmon, pollock, crab, Pacific cod, halibut, mollusks and other shellfish (Mundy, 2005). The dynamics of these fish-
5 eries and their susceptibility to climate change and ocean acidification remains poorly understood. Large glaciated mountains (Figure 1), complex bathymetry, seasonally varying cycles of winds, iron enriched freshwater discharge, and solar radiation (Stabeno et al., 2004; Weingartner et al., 2005; Janout et al., 2010) set the stage for high temporal and spatial physical, chemical, and biological variability across the GOA continental shelf. At the same time, the GOA-LME is sparsely sampled due to its large spatial extent, harsh environmental conditions, and remote geography. The societal importance of the ecosystem, the
10 large natural variability, and the lack of data make this region a challenge and a priority for understanding ecosystem dynamics, inorganic carbon chemistry, the regional manifestation of climate change and ocean acidification, and its potential impact on the nation's fisheries, economies, and communities (Mathis et al., 2014).

The GOA-LME is especially susceptible to the effects of climate change and ocean acidification. This high latitude region is naturally low in $[CO_3^{2-}]$ due to the increased solubility of CO_2 at low temperatures, seasonal increase in vertical mixing,
15 riverine and glacial inputs, and inner shelf dynamics that tend to retain coastal discharges close to shore (Feely and Chen, 1982; Feely et al., 1988; Byrne et al., 2010; Weingartner et al., 2005). The freshwater fluxes into the coastal zone are increasing (Neal et al., 2010; Hill et al., 2015; Hood et al., 2015; Beamer et al., 2016) as this sub-polar region continues to warm and deglaciate (Arendt et al., 2002; Larsen et al., 2007; O'Neel et al., 2005). Because glacial meltwater in this region is endowed with low total alkalinity (TA) relative to dissolved inorganic carbon (DIC) (Stackpoole et al., 2016, 2017), increasing freshwater dis-
20 charge pushes the system further towards undersaturation with respect to aragonite, and thereby likely exacerbating the effects of ocean acidification (Evans et al., 2014).

The GOA-LME is characterized by high concentrations of biologically available dissolved iron (dFe) and low nitrate (NO_3) concentrations on the shelf, and low dFe and high NO_3 concentrations off-shelf (Aguilar-Islas et al., 2015; Wu et al., 2009; Lippiatt et al., 2010; Martin et al., 1989). These two limiting nutrients lead to a diatom-dominated phytoplankton commu-
25 nity composition in the dFe-rich near-shore and small phytoplankton in dFe-poor off-shelf areas (Strom et al., 2007). Limited inorganic carbon observations suggest that along with primary productivity and remineralization, physical processes such as downwelling, on-shelf intrusions of deep water, tidal mixing, freshwater discharge, and eddies, are important drivers of the inorganic carbon system dynamics. High biological productivity in spring depresses pCO$_2$ in the water column, resulting in shelf waters supersaturated with respect to aragonite, and supporting large fluxes of atmospheric CO_2 into the ocean (Fabry
30 et al., 2009; Evans and Mathis, 2013). In September, bottom waters along the shelf are undersaturated with respect to aragonite. This large-scale pattern of aragonite undersaturation has been speculated to be a result of on-shelf intrusion of DIC-rich Pacific basin water near the seafloor, and remineralization of large quantities of organic matter produced during spring blooms (Fabry et al., 2009; Evans et al., 2014). Recent inorganic carbon surveys near marine tidewater glaciers in Glacier Bay, Alaska and
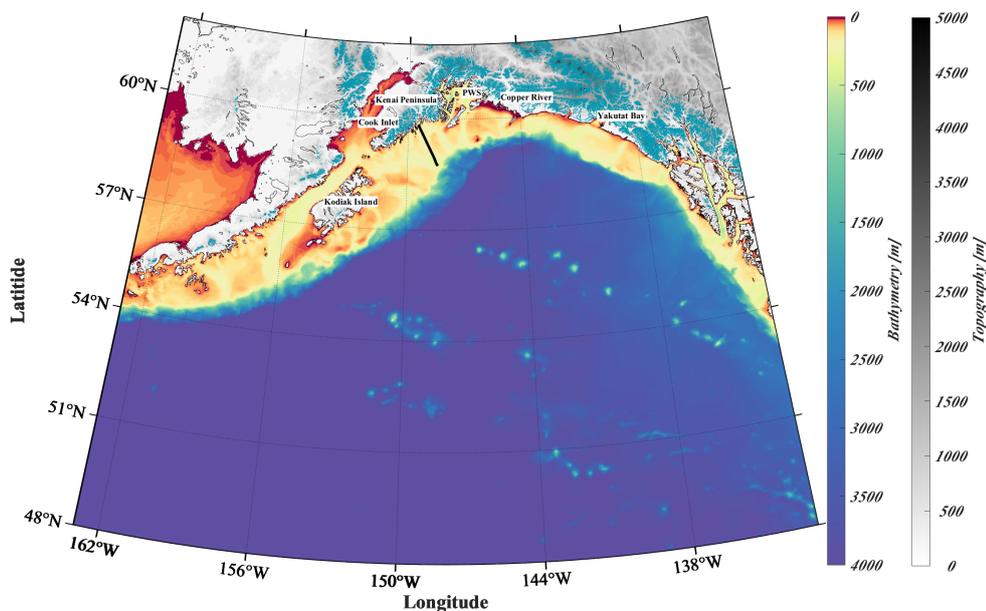
Figure 1. Map of the "GOA-COBALT" model domain showing depth [m] of the ocean in color and altitude of the mountains [m] in grey. Glaciated areas are indicated by turquoise coloring of the mountains. The Seward Line is shown by the black line. PWS = Prince William Sound.

Prince William Sound also suggest that glacial melt water, which is endowed with naturally low total alkalinity (TA), induces seasonal aragonite undersaturation in surface waters (Evans et al., 2014; Reisdorph and Mathis, 2014) and forms corrosive mode water that may subsequently be advected and potentially subducted to distant shelf regions (Evans et al., 2013, 2014). Insufficient inorganic carbon data coverage impede our ability to understand the interplay between drivers in different seasons and to determine the relative importance of the various controls of the inorganic carbon system at different locations. Ongoing and historic inorganic carbon system observations in the GOA consist of a limited set of shipboard oceanographic observations, underway measurements from research vessels and scientific sampling systems installed on ferries, and a small number of sensors deployed at fixed nearshore coastal locations (Fabry et al., 2009; Evans and Mathis, 2013; Evans et al., 2014, 2015; Reisdorph and Mathis, 2014; Miller et al., 2018). These draw an incomplete picture of the spatial and temporal variability. Also, because variability is large from seasonal to decadal scales in these waters, the timeseries from deployed sensors at fixed nearshore locations are not long enough yet to detect an anthropogenic trend in seawater $p\mathrm{CO}_2$ and pH (Sutton et al., 2018). In other shelf systems, regional biogeochemical models have successfully been used to better understand past, present, and future inorganic carbon dynamics and impacts of climate change and ocean acidification on these ecosystems (Hauri et al., 2013; Turi et al., 2016; Gruber et al., 2012; Franco et al., 2018). Previous regional GOA-LME modeling studies focused on refining large scale and mesoscale circulation features (Dobbins et al., 2009; Xiu et al., 2012), iron limitation (Fiechter and Moore,

2009), and the influence of Ekman pumping (Fiechter et al., 2009), and eddies (Coyle et al., 2012) on the timing and magnitude of the spring bloom. There are only two regional models that simulate the oceanic carbon cycle in the GOA (Siedlecki et al., 2017; Xiu and Chai, 2014). However, neither of these models simulate the influence of spatially and temporally highly variable freshwater input along the coast. Siedlecki et al. (2017) used the monthly riverine input climatology by Royer (1982) and applied it equally to the top most vertical cell along the land mask, whereas Xiu and Chai (2014) did not include riverine input at all. Therefore, to fill gaps in the understanding of the inorganic carbon cycle in the GOA we need a moderately high resolution model that includes 1) important physical processes such as downwelling, eddies, and cross-shelf fluxes, 2) the carbon cycle, 3) the large spatial and temporal variability of freshwater input, 4) historical simulation long enough to be able to distinguish natural variability from the anthopogenic longterm trend, 5) multiple phytoplankton groups, and 6) iron limitation to reproduce the highly productive nearshore and high-nutrient low-chlorophyll offshore regions.

Here, we introduce a new GOA model that includes a three-dimensional regional ocean circulation model, a complex ecosystem model, and a moderately high-resolution terrestrial hydrological model. This model system will allow us to better simulate the influence of variations in past freshwater runoff, full suite of atmospheric forcing and physical air-sea fluxes of heat, water vapor and momentum, and atmospheric CO_2 on the physical, chemical, and biological oceanography. We make a thorough model evaluation to test the model's capability of simulating seasonal and interannual inorganic carbon patterns, and use it to study the seasonal variability and drivers of the inorganic carbon system along the historic Seward Line (Figure 1).

2 Methods

2.1 Model set-up

We used a GOA configuration of the three-dimensional physical Regional Oceanic Model System (ROMS, Shchepetkin and McWilliams (2005)). ROMS is a free-surface, hydrostatic primitive equation and finite volume (Arakawa C-grid) ocean circulation model. The vertical discretization is based on a terrain-following coordinate system (50 depth levels), with increased resolution towards the surface and the bottom of the ocean. Due to shallower bathymetry, the shelf and coastal areas have higher vertical resolution. The horizontal resolution is eddy resolving at 4.5 km. The grid covers a large coastal area from the Canadian - US border in the Southeast to west of Sandpoint in the middle of the Aleutian Islands (Figure 1). The current model configuration is based on Danielson et al. (2016), although with a larger grid extent and a lower horizontal resolution to accommodate the computationally intense ecosystem model. We attempt to improve on the GOA domain used by Coyle et al. (2012) in that we include the Alexander Archipelago. Prior experiments in the GOA have shown that the model had insufficient near-surface vertical mixing, leading to overly fresh water at the surface which is challenging to mix down. In order to improve on our surface mixing, we added the parameterization of Li and Fox-Kemper (2017). They looked to Large-eddy simulations (LESs) to study the effects of unresolved Stokes drift driven by surface waves and the resulting Langmuir circulation on the vertical mixing in a variety of stratification regimes. Their parameterization is now being routinely used in global climate simulations (e.g. Adcroft et al. (2019)).

The hydrological model is based on a suite of weather, energy-balance snow ice/melt, soil water balance, and runoff routing

Table 1. Summary of the definitions, abbreviations, values and units of the model parameters adjusted from Van Oostende et al. (2018) to better represent the ecosystem in higher northern latitudes (Strom et al., 2010)

Parameter	Symbol	Value	Units
Iron scavenging coefficient onto sinking detritus	alpha fescav	10/a	sec^{-1}
Chl-a specific initial slope of the photosynthesis-light curve			
Large chain-forming phytoplankton	alpha Lg	3.0e-6 * b	$(g\ C)\ (g\ Chl^{-1})\ (W\ m^{-2})^{-1}$
Medium Phytoplankton	alpha Md	3.33e-6 * b	$(g\ C)\ (g\ Chl^{-1})\ (W\ m^{-2})^{-1}$
Small Phytoplankton	alpha Sm	6.0e-6 * b	$(g\ C)\ (g\ Chl^{-1})\ (W\ m^{-2})^{-1}$

a = 365.25*24*3600, b = 4.5998

models forced with Climate Forecast System Reanalysis data products (Saha et al., 2010).

The biogeochemical model used for this study is a modified version of the Carbon, Ocean Biogeochemistry and Lower Trophic (COBALT) marine ecosystem model, which has been applied on a global scale in conjunction with the Geophysical Fluid Dynamics Laboratory's (GFDL's) earth system model (Stock et al., 2014). Only recently, COBALT was coupled to ROMS and

5   modified with an additional coastal chain-forming diatom to better represent biogeochemical processes and properties in highly productive coastal regions (3PS-COBALT, Van Oostende et al. (2018)). 3PS-COBALT resolves the cycles of nitrogen, carbon, phosphate, silicate, iron, calcium carbonate, oxygen, and lithogenic material with 36 state variables. The model contains small phytoplankton that are grazed by microzooplankton, a medium phytoplankton that can be consumed by small copepods and a larger, chain forming diatom, which can only be grazed by large copepods and krill. Light, temperature, the most limiting nu-

10   trient and metabolic costs are used to calculate primary productivity for each phytoplankton group. The chlorophyll-to-carbon ratio is dynamic and based on light (Manizza et al., 2005) and nutrient limitations. Iron limitation depends on an internal cell quota and nitrate/phosphate limitations are simply dependent on their ambient concentrations in the seawater. The currency of biomass and productivity in the model is nitrogen. Organic nitrogen is converted into organic carbon following the Redfield Ratio of 106:16 (Redfield et al., 1963). Dissolved inorganic carbon (DIC) and TA are state variables and dictate the inorganic

15   carbon system. Like all other tracer concentrations in the model, these two variables are affected by diffusion, horizontal and vertical advection, and sources minus sink terms that include net primary production, $CaCO_3$ formation and dissolution both in the water column and sediments, detritus remineralization in the water-column and sediments, zooplankton respiration, and atmosphere-ocean CO_2 fluxes. The Chl-*a* specific initial slope of the photosynthesis-light curve and iron scavenging coefficient onto sinking detritus were both adjusted to better represent the GOA biogeochemistry and ecosystem (Table 1). All other

20   constants are based on Van Oostende et al. (2018).

2.2 Initial Conditions, Boundary Conditions and Forcing

Physical initial and boundary conditions for currents, ocean temperature, salinity, and sea surface height were taken from the Simple Ocean Data Assimilation ocean/sea ice reanalysis 3.3.1 [SODA, Carton et al. (2018), available as five-day averages]. The model was forced at the surface with three-hourly winds, surface air temperature, pressure, humidity, precipitation and radiation from the Japanese 55-year Reanalysis (JRA55-do) 1.3 project (Tsujino et al., 2018). The atmospheric fields are used to compute surface stresses and fluxes using a bulk flux algorithm (Large and Yeager, 2008). Precipitation is solely counted as a negative salt flux and does not change any volume or dilute any other tracers. Along its coastal boundary, freshwater was brought in from numerous rivers and tidewater glaciers. This was done with a point-source river input through exchange of mass, momentum and tracers through the coastal wall at all depths (Danielson et al., in review). A hydrological model provided riverine input at a 1 km resolution and at a daily timestep (Beamer et al., 2016). The hydrological model is based on a suite of weather, energy-balance snow ice/melt, soil water balance, and runoff routing models forced with Climate Forecast system Reanalysis data (Saha et al., 2010). Nutrient, DIC, and TA concentrations in the freshwater are based on available observations and are summarized in (Table 2). As found in other freshwater data (Rheuban et al., 2019), DIC > TA showing that the freshwater discharging into the GOA is quite acidic (Stackpoole et al., 2016, 2017).

DIC and TA initial conditions for the hindcast simulation were extracted from the mapped version 2 of the Global Ocean Data Analysis Project dataset (GLODAPv2.2016b, Lauvset et al. (2016)). GLODAP DIC data, which is referenced to 2002, was normalized to 1980 using the anthropogenic CO_2 estimates for the GOA by Carter et al. (2017). A seasonal cycle was added to the surface based on Takahashi et al. (2014). Following Hauri et al. (2013), DIC and TA are assumed to vary throughout the upper 200 m of the water column, but are attenuated proportional to the seasonal variations of temperature across depth. Each year, DIC boundary conditions increase as a result of anthropogenic CO_2 increase, following estimates for the region made by Carter et al. (2017). Nitrate, phosphate, oxygen, and silicate initial conditions were taken from the World Ocean Atlas 2013 (Garcia et al., 2013). All other variables were initialized based on output from a global model simulation (Stock et al., 2014). Atmospheric pCO_2 was forced with monthly means derived from the Mauna Loa CO_2 record (https://www.esrl.noaa.gov/gmd/ccgg/trends/data.html).

3   Model Evaluation

3.1 Physics

The model domain is strongly influenced by freshwater coming from hundreds of glacier-fed rivers and tidewater glaciers that ring the GOA. Our approach of using a hindcast simulation from a highly resolved land hydrography model (Beamer et al., 2016) to force the freshwater input was recently evaluated through comparison to salinity, temperature, velocity, and dynamic height observations (Danielson et al., in review). The influence of the springtime freshet is well reproduced by the model, with low salinities (∼26 PSU) in the nearshore upper 10 to 20 m of the water column (r ∼ 0.5-0.6, p ∼ 0.05). While the correlation between salinity observations and model output is only slightly weaker in summertime, the model has difficulty in reproducing

Table 2. Table listing riverine dissolved inorganic carbon, total alkalinity, nitrate, dissolved iron, dissolved oxygen, and temperature values. When seasonal observations were available, the variables followed a seasonal cycle. These variables are listed with their mean, maximum, and minimum levels. All other variables were initialized with a very small number.

Parameter	Value	Units	Reference
Dissolved inorganic carbon	397, 446, 365	umol kg^{-1}	Kenai river (Stackpoole et al., 2016, 2017)
Total alkalinity	366, 410, 333	umol kg^{-1}	Kenai river (Stackpoole et al., 2016, 2017)
Iron	30	nmol kg^{-1}	(Aguilar-Islas et al., 2015)*
Nitrate	4.86, 8.91, 1.5	umol kg^{-1}	Herbert River (Hood and Berner, 2009)
Temperature	5.06, 11.88, 0.16	°C	Danielson, personal communication
Dissolved oxygen	386.5, 405.2, 355.1	umol kg^{-1}	Cowee Creek (Fellman et al., 2015)

*Aguilar-Islas et al. (2015) observed a maximum of 10 nmol kg^{-1} dFe in nearshore waters with a salinity of 25. In order to model high nearshore dFe values (see Figure 2), riverine dFe was set to 30 nmol kg^{-1}.

small-amplitude salinity variations between October and March, when riverine intput is lowest and the signal to noise ratio of salinity is small. Temperature is particularly well modelled in the first half of the year, with highest correlations (r $\sim$ 0.9, p < 0.001) in the middle of the water column. The model is also able to skilfully reproduce monthly dynamic height anomalies in summer, fall, and winter. Overall, freshwater fluxes used in this study are more realistic than in previous studies because they are derived from a temporally and spatially highly resolved hindcast simulation forced with meteorological reanalysis data sets. Compared to previous studies (Royer, 1982; Wang et al., 2004; Hill et al., 2015; Siedlecki et al., 2017), this current model configuration is better able to reproduce low salinity levels in coastal areas across the GOA and can thereby more adequately reproduce the Alaska Coastal Current (Danielson et al., in review).

3.2 Nutrients and Chl-α

High concentrations of biologically available dFe prevail across the inner to mid GOA shelf due to the dFe-rich input of glacially fed rivers (Aguilar-Islas et al., 2015; Wu et al., 2009; Lippiatt et al., 2010). While these shelf areas are low in NO$_3$, they border the NO$_3$-rich and dFe-poor waters off-shelf (Martin et al., 1989). The spatial variability of these two limiting nutrients orchestrate the phytoplankton community composition, with diatom-dominated areas in the dFe-rich near-shore environment and small phytoplankton in iron depleted, offshore areas (Strom et al., 2010). Here, we first evaluate the modelled spatial variability of surface dFe, NO$_3$, and Chl-α by comparing it to *in situ* and satellite observations (Aguilar-Islas et al., 2015; Lippiatt et al., 2010; Crusius et al., 2017; MODIS-Aqua Ocean Color Data; NASA Goddard Space Flight Center, 2014, Accessed 08/17/2019)

dFe observations along a transect near the Copper River estuary suggest that coastal surface concentrations can reach up to 9.4 nmol kg^{-1} in mid to late summer west of Prince William Sound (Figure 2a), when glacial melt and riverine input are highest. These surface concentrations decrease rapidly with distance from shore, with levels of <1 nmol kg^{-1} near the shelf break and
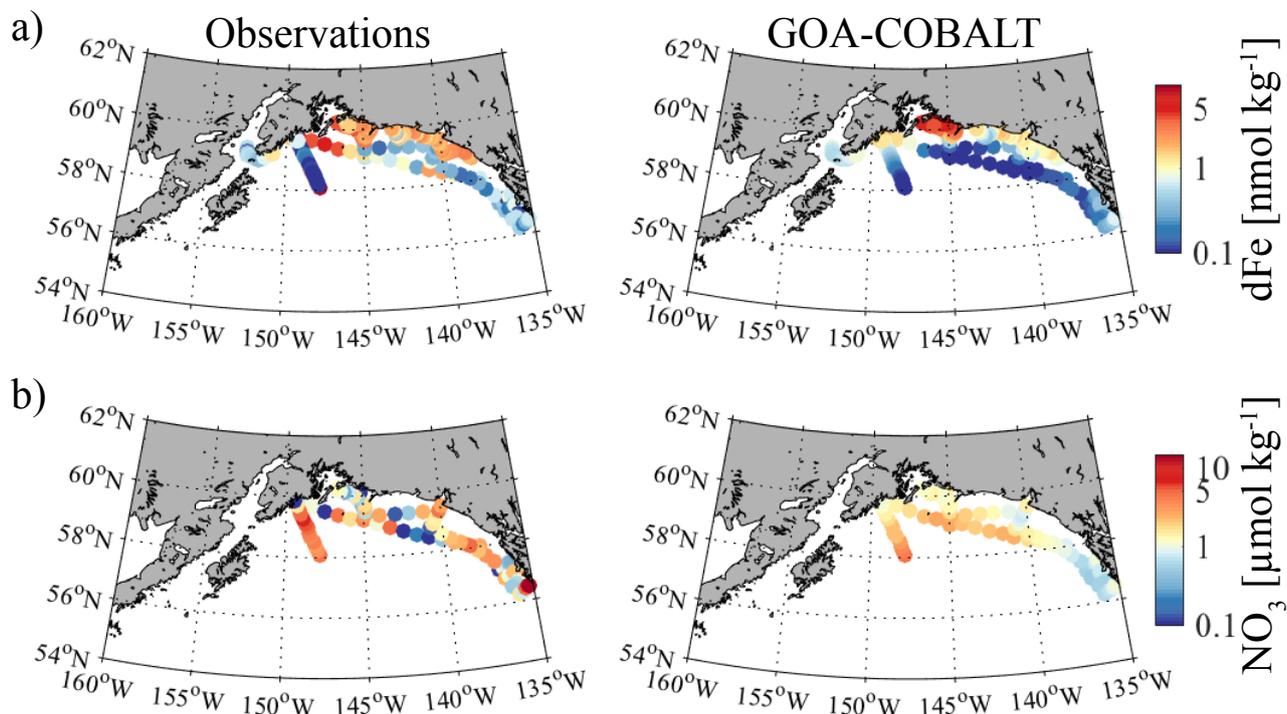
Figure 2. Observed (left) and modeled (right) surface dissolved iron (a, dFe [nmol kg^{-1}]) and nitrate (b, NO$_3$ [μmol kg^{-1}]) concentrations. The observations were taken in July, August, and September (Aguilar-Islas et al., 2015; Lippiatt et al., 2010; Crusius et al., 2017). The model output is based on a mean of modelled July, August, and September values from a 1980 to 2013 climatology. Dissolved iron and nitrate are plotted on a log scale.

beyond. The horizontal surface dFe gradient across the shelf, with high dFe concentrations in coastal areas and dFe levels near depletion off-shore, was simulated well by the model. For example, the model shows dFe values of up to 8 nmol kg^{-1} (Figure 2a) in some coastal areas and low concentrations (< 0.5 nmol kg^{-1}) at the shelf break. On average, the model underestimates summertime surface iron concentrations by $\sim$ 2 nmol kg^{-1} despite the relatively high dFe riverine boundary condition (Table

5   2).

The low observed surface summertime values of NO$_3$ in coastal areas are simulated well by the model. Observed NO$_3$ values range up to 5.4 μmol kg^{-1} (with exception of a few outliers in southeast Alaska) and modelled surface NO$_3$ concentrations also only reach a maximum of 4.3 μmol kg^{-1} (Figure 2b).

The modelled spring phytoplankton bloom starts in April and therefore coarsely matches with the timing of the satellite ob-

10   served spring bloom (Figure 3). Then, simulated surface Chl-α ranges between 3 to 4 μg kg^{-1} across the entire shelf, with some nearshore areas reaching surface Chl-α levels of 5 μg kg^{-1}. However, satellite images suggest that in April, Chl-α concentrations are more patchy than shown by the model, with lower concentrations across the shelf (1-2 μg kg^{-1}). While peak observed Chl-α concentrations between 7 and 9 μg kg^{-1} are localized in southeast Alaska in May, simulated Chl-α levels of 5
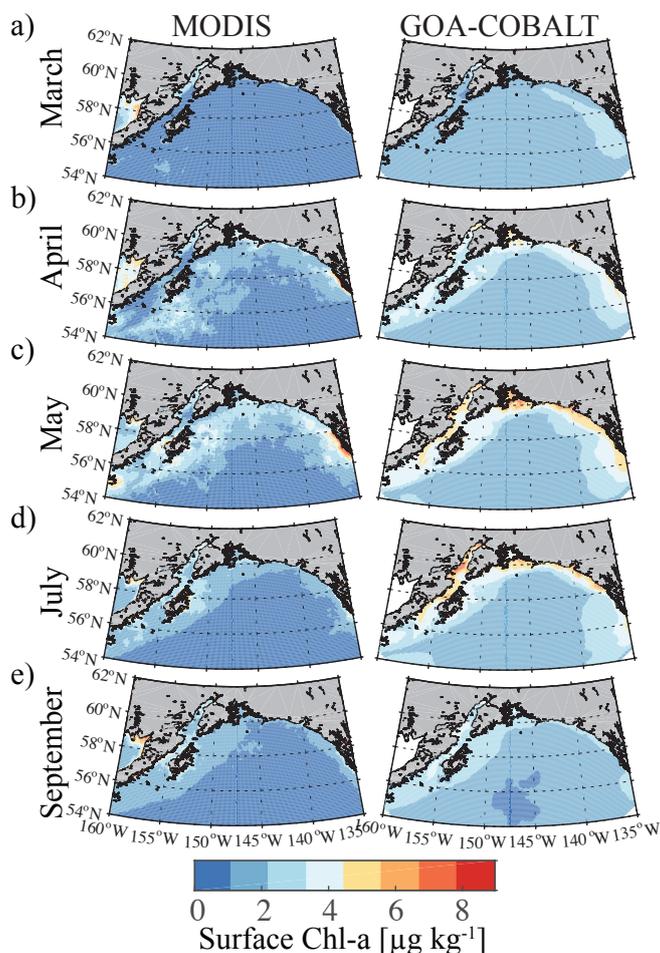
Figure 3. Monthly climatology for satellite based (left, MODIS) and simulated (right) surface Chl-α concentrations for (a) March, (b) April, (c) May, (d) July, and (e) September. Source for MODIS data: (MODIS-Aqua Ocean Color Data; NASA Goddard Space Flight Center, 2014, Accessed 08/17/2019)

μg kg^{-1} are more wide spread along the coast, with peak concentrations of 6 to 7 μg kg^{-1} in some select areas. In July through August, simulated Chl-α levels slowly taper off, though some nearshore areas still reach peak concentrations of 6 μg kg^{-1}. In September, on-shelf modelled Chl-α are < 3 μg kg^{-1}. To sum up, observed Chl-α concentrations are lower than suggested by the model across the shelf and the domain throughout the summer. With exception of November through February, when modelled Chl-α concentrations are < 1 μg kg^{-1}, standing stock Chl-α concentrations are also overestimated by a factor of two by the model throughout the year.
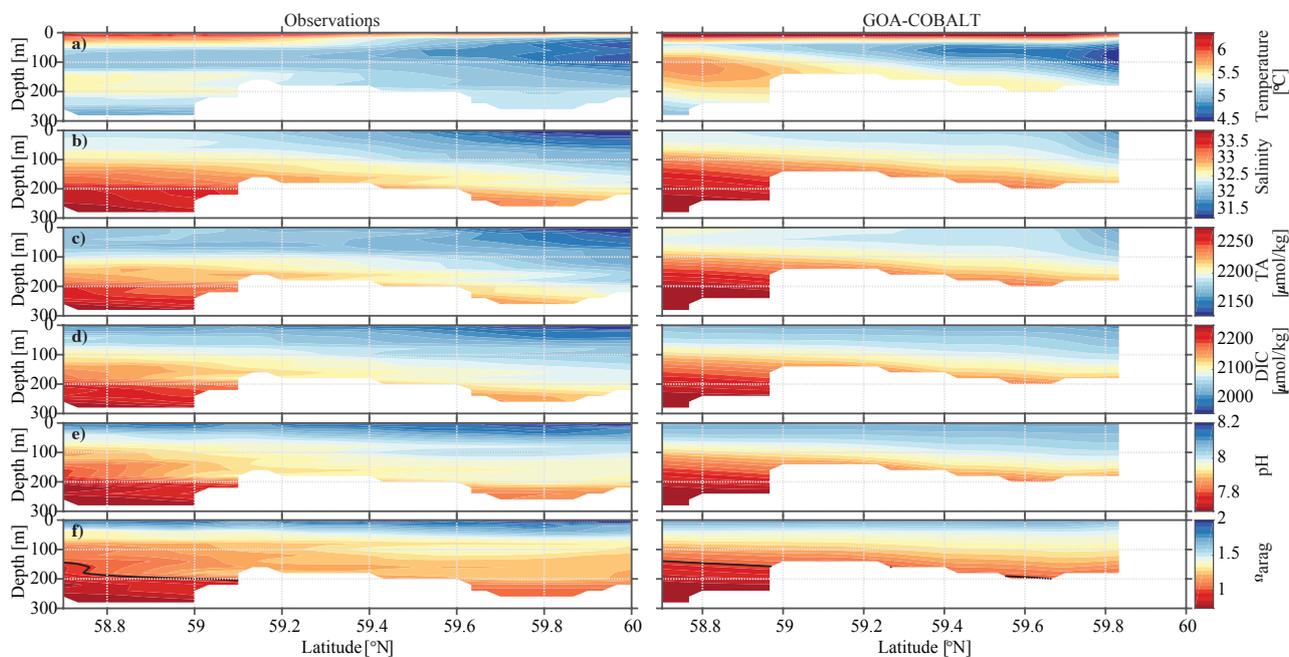
Figure 4. Visual comparison of vertical sections of (a) temperature ($^\circ$C), (b) salinity, (c) total alkalinity (TA, μmol kg^{-1}), and (d) dissolved inorganic carbon (DIC, μmol kg^{-1}), (e) pH, and (f) aragonite saturation state (Ω_{arag}) of observations taken along the Seward Line in May (left, Evans et al. (2013)) and corresponding model output (right). The model output was sampled at locations where observations were taken. Observations and model points were then vertically and horizontally interpolated onto the same grid and averaged across years 2008 - 2012. The white area depicts observed sampling depth in the left panel and seafloor or coast in the model output.

3.3 Model skill to simulate spring and fall oceanographic conditions

Inorganic carbon observations in the GOA have been collected every spring and fall along the historic Seward Line since 2008 (Evans et al., 2013). Here, we compare May and September monthly means of our 2008 - 2012 model output to the publicly available ship board observations taken in May and September of the corresponding years. To do so, the model output was sampled at the closest grid point to every Seward Line station. The model output and observations were then interpolated onto the same grid and averaged across all years for May and September.

In May, observations and model output clearly show the influence of freshwater on the coastal water properties, including inorganic carbon chemistry (Figure 4). In nearshore surface areas, waters are cold, fresh, and low in TA and DIC concentrations. The influence of freshwater is visible throughout the transect, but with slowly increasing values of salinity, DIC, and TA with distance from the coast. The model generally reproduces these patterns well, although with slightly higher values for all three parameters across the transect. For example, observed surface salinity ranges between 31.4 at 59.8 $^\circ$N and 32.2 at the end of the transect. TA varies between 2143 μmol kg^{-1} and 2170 μmol kg^{-1}, and DIC between 1965 μmol kg^{-1} and 2017 μmol kg^{-1}. Simulated ranges of these parameters are similar (salinity: 31.7 - 32.4, TA: 2163 - 2203 μmol kg^{-1}, DIC: 2009 - 2038
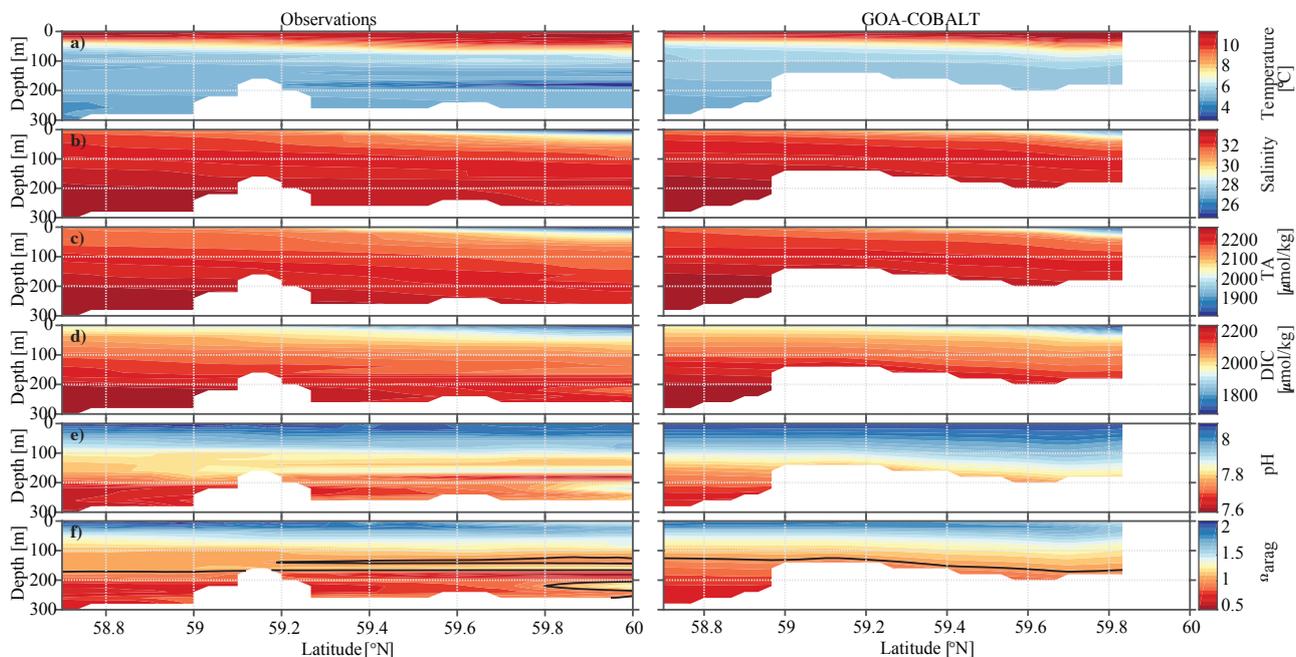
Figure 5. Visual comparison of vertical sections of (a) temperature ($^{\circ}$C), (b) salinity, (c) total alkalinity (TA, μmol kg^{-1}), and (d) dissolved inorganic carbon (DIC, μmol kg^{-1}), (e) pH, and (f) aragonite saturation state (Ω_{arag}) of observations taken along the Seward Line in September (left, Evans et al. (2013)) and corresponding model output (right). The model output was sampled at locations where observations were taken. Observations and model points were then vertically and horizontally interpolated onto the same grid and averaged across years 2008 - 2012. The white area depicts observed sampling depth in the left panel and seafloor or coast in the model output.
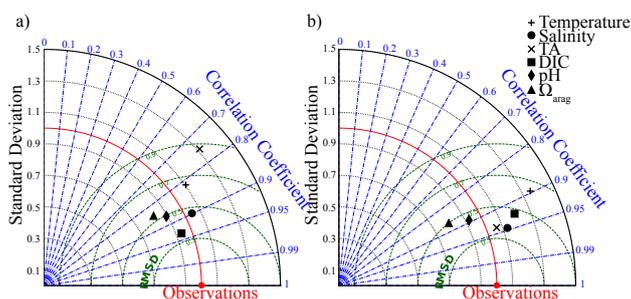
Figure 6. Taylor diagrams (Taylor, 2001) of model simulated temperature, salinity, total alkalinity (TA, μmol kg^{-1}), dissolved inorganic carbon (DIC, μmol kg^{-1}), pH, and aragonite saturation state (Ω_{arag}) compared to observations taken along the Seward Line in a) May and b) September. Averaged modelled monthly means of May and September are compared to data collected during spring and fall cruises in years 2008 through 2012. The distance from the origin is the normalized standard deviation of the modeled parameters. The azimuth angle shows the correlation between the observations and the modelled output, whereas the distance between the model point and the red observation point shows the normalized root mean square misfit between modeled and observed data.
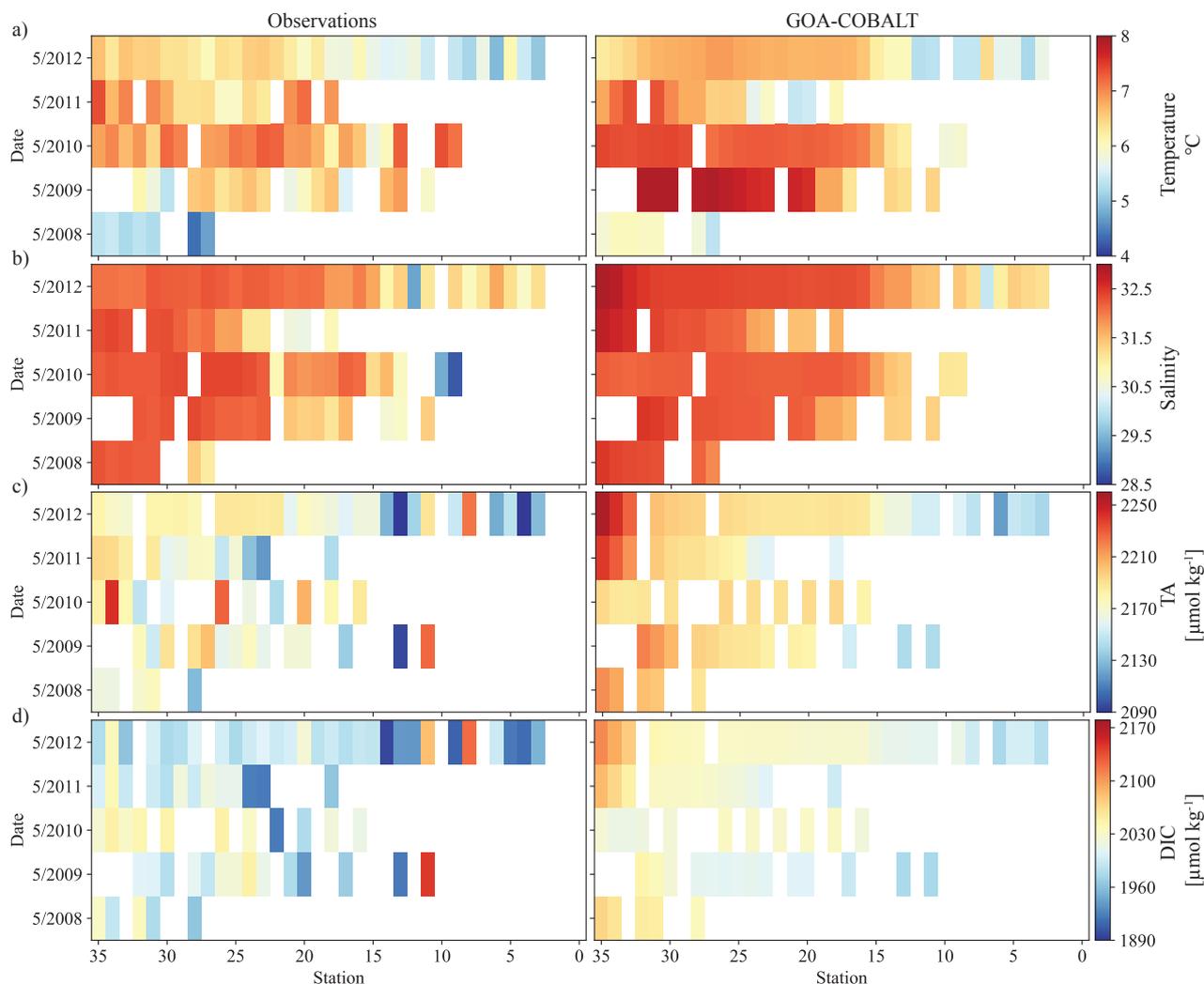
Figure 7. Hovmoeller plots showing observed (left) and modelled (right) monthly mean surface (a) temperature ($^\circ$C), (b) salinity, (c) total alkalinity (TA, μmol kg^{-1}), and (d) dissolved inorganic carbon (DIC, μmol kg^{-1}) in May along the Seward Line between 2008 - 2012.

μmol kg^{-1}). The largest bias in salinity (S$_{bias}$ = 0.6), alkalinity (TA$_{bias}$ = 35 μmol kg^{-1}), and DIC (DIC$_{bias}$ = 46 μmol kg^{-1}) are found around 59.6 $^\circ$N. The overestimation of salinity suggests that the freshwater influence is overly weak, leading to the modelled overestimation of TA and DIC. However, the bias in DIC relative to the bias in TA is largest in the grid cell closest to the coast, which leads to an underestimation of modelled pH and Ω_{arag} of up to 0.09 and 0.29, respectively. The bias of surface pH and Ω_{arag} vanish with distance from the coast. At the offshore end of the transect, the model overestimates Ω_{arag} by 0.07 (pH overestimation is insignificant). The larger difference in DIC relative to TA near the coast is likely due to an underestimation of biological carbon draw-down, leading to the underestimation of pH and Ω_{arag} in nearshore surface waters.
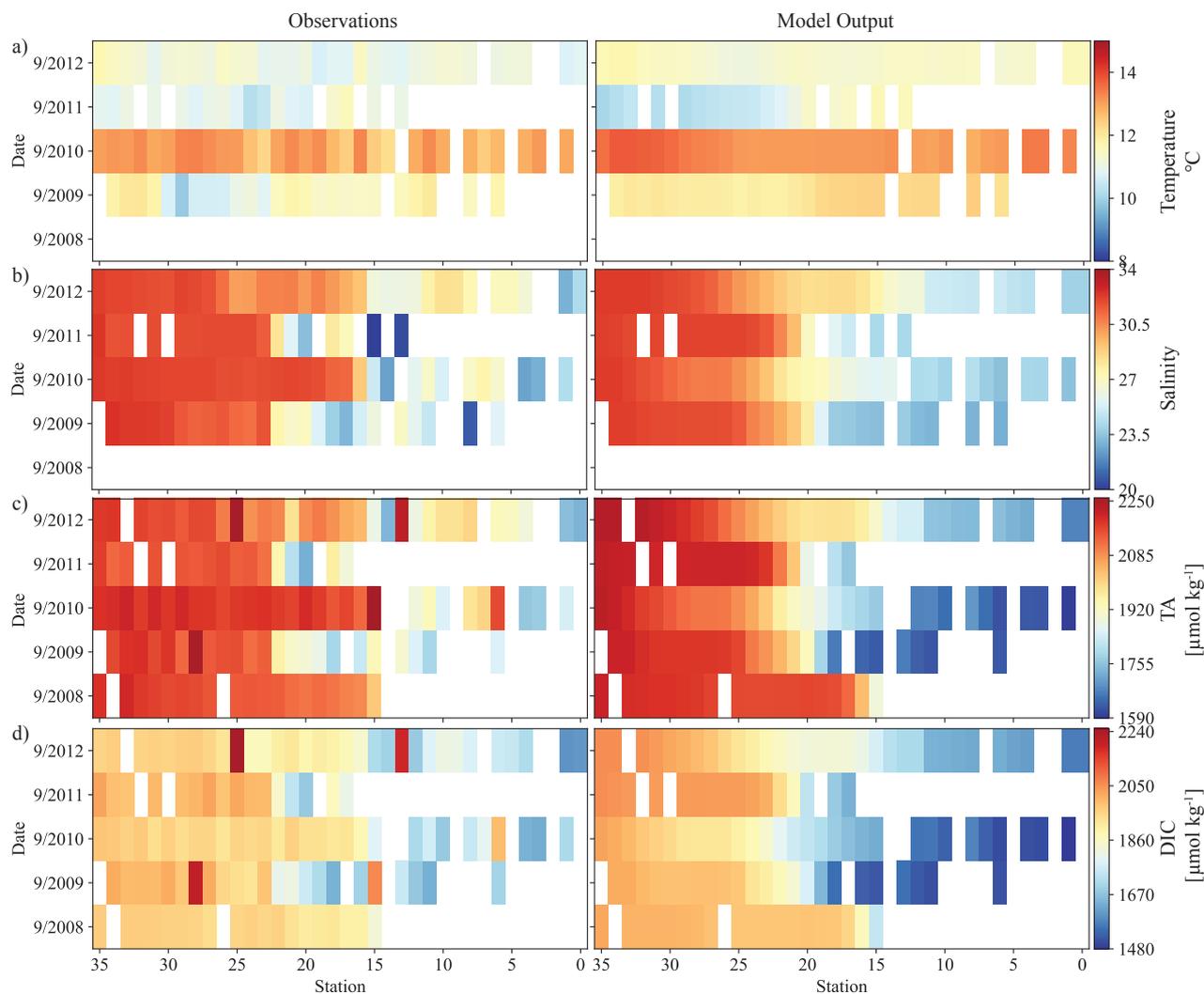
Figure 8. Hovmoeller plots showing observed (left) and modelled (right) monthly mean surface (a) temperature ($^\circ$C), (b) salinity, (c) total alkalinity (TA, μmol kg^{-1}), and (d) dissolved inorganic carbon (DIC, μmol kg^{-1}) in September along the Seward Line between 2008 - 2012.

At depth farther offshore, salinity, DIC and TA enriched waters are visible on the shelf. As a result, the springtime *in situ* aragonite saturation horizon (depth where Ω_{arag} = 1) is shallower offshore than nearshore. Observations suggest that with a depth of the aragonite saturation horizon of about 200 m, the seafloor on the shelf up to 59.1 $^\circ$N is undersaturated with regard to aragonite. The model underestimates the aragonite saturation horizon depth slightly. However, because the model's bathymetry is too shallow (white areas in Figure 4), modelled bottom water masses across the shelf are not understaturated with regard to Ω_{arag} north of 59 $^\circ$N.

In September, nearshore surface waters are warm ($\sim$ 11°C) and much fresher than in spring. Observed surface salinity is as low

as 27.6 at 59.8 °C and increases to 32 salinity units at the end of the transect. Modelled surface salinity ranges between 26.3 and 31.8. It is important to note that observed surface salinity at the station closest to the coast is also as low as 26.8, but is masked out in this point by point comparison. As a result of this bias in salinity, modelled TA and DIC are underestimated by 160 and 60 μmol kg^{-1} at 59.8 °N. Because the model is likely underestimating the magnitude of late season phytoplankton blooms,

5 the bias in DIC is not as large as in TA. Modelled surface Ω_{arag} is 1.48 at the coast and thereby underestimates observations by 0.27. At depth, model output aligns well with observations. Both observations and model output suggest that bottom waters across the Seward Line transect are undersaturated with regard to aragonite in fall. Observations also show an intermediate maximum in DIC at about 150 m depth, which leads to an even shallower aragonite saturation horizon. This small scale feature may be due to remineralization at mid depth and is not simulated by the model. Overall, the model does a reasonable job in

10 simulating salinity, temperature, TA, DIC, pH, and Ω_{arag} in spring and in fall, with Pearson Correlation Coefficients between 0.75 and 0.95. All standard deviations and Correlation Coefficients between observed and modelled variables are summarized in Taylor Diagrams in Figure 6.

3.4 Model skill to simulate interannual variability

The GOA oceanography undergoes large interannual to decadal variability as a result of the El Nino Southern Oscillation

15 (ENSO) and Pacific Decadal Oscillation (PDO) and other climate modes of variability that drive changes in freshwater flux, water temperature and winds (Whitney and Freeland, 1999; Hare and Mantua, 2000). These physical changes are likely translated into the inorganic carbon chemistry. As a first approach, we use Hovmoeller plots to investigate the model's skill to simulate interannual features (Figures 7 and 8). In the following, we will describe observations of years that stand out within this five year record and determine whether these features were captured by the model.

20 Observed and modelled springtime surface temperature show large interannual variability. Within the 5 year long record, 2008 was a particularly cold spring, with surface temperatures around 5 °C. The model also simulated May 2008 as the coldest spring within this five year period, however, it overestimated the temperature by about 1 °C. In May 2010, observed sea surface temperatures were particularly high (> 6 °C) across the shelf, which was well reflected by the model. While the model also suggests high temperatures in the previous spring, the observations do not show a particularly warm spring in 2009. In May

25 2012, colder surface temperatures (< 6 °C) prevailed near the coast and through station 15 in both the observed and modelled data.

The interannual variability in salinity is visible by how far the freshwater penetrates into the open ocean. For example, observations and model output suggest that freshwater penetrated farther offshore in years 2008, 2009, and 2011. A similar pattern is visible in observed and modelled TA, with low TA values (<2170 μmol kg^{-1}) advecting across the shelf and beyond in

30 2009 and 2010. As already described in Section 3.3, the model persistently overestimates DIC in spring and summer. While the influence of low salinity on DIC becomes visible in 2009 and 2010, the observed low DIC in 2012 was not reproduced by the model, likely because it was driven by strong biological carbon drawdown rather than freshwater.

The anomalous warm sea surface temperatures observed in spring 2010 remained persistent into fall, with temperatures around 13 °C across the whole transect (Figure 8). This anomalously warm fall was well simulated by the model. Interestingly, in

the same year, observations suggest anomalously high salinities penetrating closer to the nearshore than in other years, which was not reflected by the model. However, observed lower salinities mid transect in 2009 and 2011 were also simulated by the model, which directly translated into TA and DIC. Overall, fall DIC levels were better simulated by the model than spring conditions.

5 4   Seasonal inorganic carbon variability along the Seward Line

The historic Seward Line timeseries gives insights into the inorganic carbon dynamics in May and September. Here, we use our model output to explore other months of the year that are not covered by inorganic carbon observations (Figure 9). Lowest surface temperatures of $< 3°$C are found nearshore in February and March. In spring, surface water slowly warm, reaching an annual maximum in July/August, when surface temperatures are $> 13°$C. Highest nearshore surface salinities (S^{max}=31) are

10 observed in late winter, which decrease to 24 salinity units between August and October, when the influence of freshwater is strongest. The freshwater also decreases surface TA and DIC to their lowest levels of 1625 μmol kg^{-1} and 1500 μmol kg^{-1} in August. The additional biologically driven decrease of DIC between April and June, when surface Chl-α concentrations (up to 7 μg kg^{-1}) are highest, leads to relatively high pH (8.13) and Ω_{arag} (2.05), despite the freshwater influence and its diluting character. Once phytoplankton blooms begin to taper off in July and August, DIC slowly increases relative to TA,

15 leading to a decrease in Ω_{arag} to a low of 1.21 and pH of 8.05 in October. Ω_{arag} remains < 1.5 across the whole water column between January and March. In April, the incoming light and nutrient concentrations are sufficient again to trigger phytoplankton blooms that slowly decrease DIC and thereby increase Ω_{arag} and pH.

Starting in May, downwelling relaxes and more saline and DIC rich waters intrude onto the shelf, leading to maximum bottom salinities and DIC of S^{max}=33.6 and DIC^{max}=2186 μmol kg^{-1}, respectively, on the shelf. Relaxation of downwelling results

20 in aragonite undersaturation of bottom waters across the transect between June and January. Destruction of organic matter and remineralization additionally increase DIC between June and September and thereby further enhance aragonite undersaturation (Figure 10 b). Only nearshore areas north of 59.8°N remain oversaturated throughout the year. The onset of downwelling in September typically starts the annual decrease of near-bottom shelf salinity and DIC (Ω_{arag} and pH) levels, which reach their respective minimums in late winter or spring.

25 5   Influence of glacial freshwater on surface Ω_{arag}, pH, and $p\text{CO}_2$

Glacial freshwater is the most important driver of the near-shore inorganic carbon dynamics of the GOA in summer and fall. In the following, we further investigate the influence of coastal dilution from the non-zero DIC and TA and rather acidic freshwater end-member (Table 2) on surface Ω_{arag}, pH, and $p\text{CO}_2$ (Figure 11a). To do so, we used mixing diagrams of surface Ω_{arag}, pH, and $p\text{CO}_2$ versus salinity from an area starting at the Kenai Peninsula to south of Yakutat Bay. Note that the plots

30 show the effects of salinity changes, and biogeochemistry on surface Ω_{arag}, pH, and $p\text{CO}_2$, as well as seasonal differences in temperature and in the off-shore and freshwater end-members DIC and TA. Following Rheuban et al. (2019), we used a
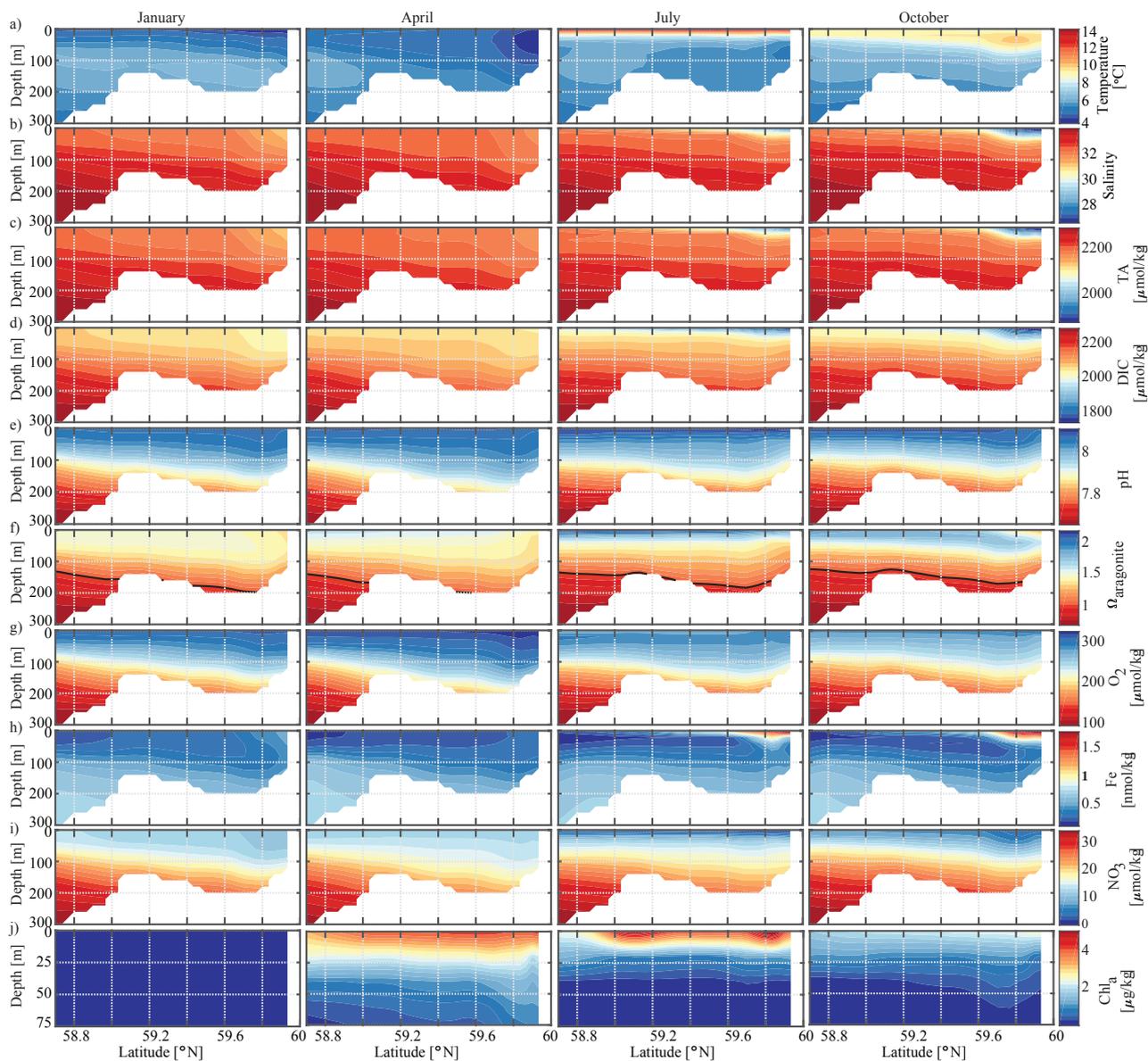
Figure 9. Seward Line transect of modelled (a) temperature ($^\circ$ C), (b) salinity, (c) total alkalinity (TA, μmol kg^{-1}), (d) dissolved inorganic carbon (DIC, μmol kg^{-1}), (e) pH, (f) aragonite saturation state (Ω_{arag}), (g) dissolved oxygen (DO, μmol kg^{-1}), (h) dissolved iron (FE, nmol kg^{-1}), (i) nitrate (NO$_3$ μmol kg^{-1}), and (j) Chl-α (μg kg^{-1}) in January, April, July, and October (averaged across 2008 - 2012). Note that Chl-α is only plotted across the upper 75 m of the water column. The black solid line in row f indicates $\Omega_{arag} = 1$.

linear Taylor decomposition to evaluate the importance of these different drivers in controlling the variations in surface Ω_{arag}, pH, and pCO$_2$ (Figure 12). The shaded areas in Figures 11c and d show the additive effect of salinity and dilution with seasonally varying DIC and TA freshwater end-members. Deviations from the shaded area are driven by biogeochemistry and
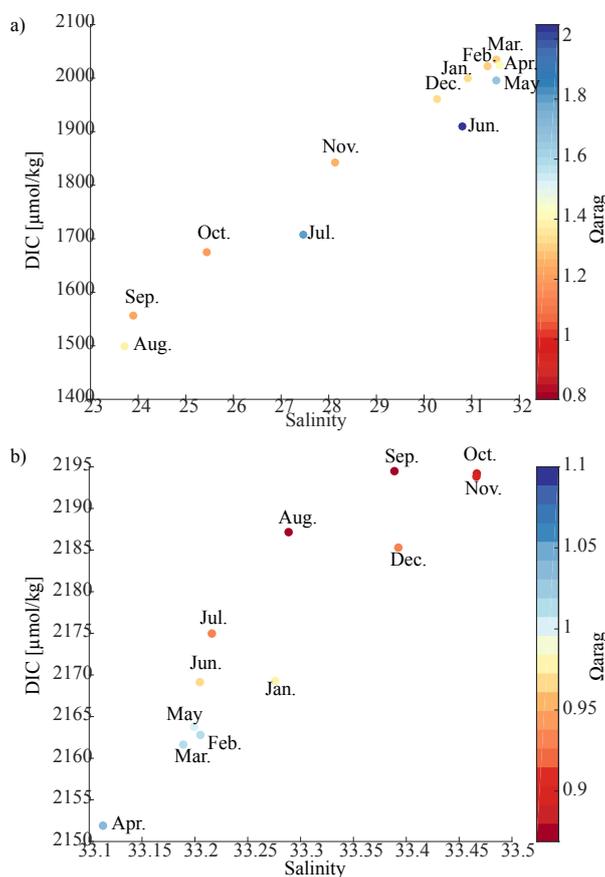
Figure 10. Salinity versus dissolved inorganic carbon (DIC, μmol kg^{-1}) as a function of Ω_{arag} and month a) at the surface and grid cell closest to the coast and b) on the seafloor at 59.2 °N along the Seward Line. Note that colorbars are on different scales.

temperature. We focus our analysis on April and August because these two months are, respectively, the least and most affected by freshwater.

In April, surface salinity is > 30, with exception of a few near-shore spots in Prince William Sound and Yakutat Bay, where surface salinity can decrease to 28 salinity units (Figure 11 a). Surface Ω_{arag} ranges between 1.5 and 1.85, pH varies between

5   8.08 and 8.10, and pCO$_2$ between 280 μatm and 340 μatm (Figures 11 b and c).

In August, surface salinity can be as low as 18 in Prince William Sound and near Copper River, with a strong salinity gradient increasing towards offshore, reaching salinities > 32 there. Surface Ω_{arag} ranges between 1.14 and 2.5 and pCO$_2$ between 205 μatm and 333 μatm, both reflecting the strong influence of freshwater. In contrast, the range of pH in August is similar to the one in April (8.07 to 8.16), suggesting that freshwater affects pH to a much smaller degree than Ω_{arag} and pCO$_2$. Decreasing

10   salinity exerts a positive effect on both, Ω_{arag} and pH, while mixing with low TA and DIC freshwater end-members has a negative effect. For pH, the positive effect of decreasing salinity is much stronger than for Ω_{arag}, thereby cancelling or even counteracting, depending on the freshwater end-member, the negative effect of mixing. In contrary, Ω_{arag} is more strongly

affected by mixing than by salinity, resulting in a decrease of Ω_{arag} by about 1 unit (Figures 12a and b). In contrary to Ω_{arag} and pH, decreasing salinity exerts a strong negative effect on $p\text{CO}_2$, which is also supported by, but to a lesser extent, a negative effect of mixing with low TA and DIC end-members (Figure 12c). Deviations from the shaded areas in Figure 11d are mainly driven by biogeochemistry, decreasing Ω_{arag} and pH, and increasing $p\text{CO}_2$. Temperature effects are small for Ω_{arag} and pH,

5   whereas for $p\text{CO}_2$, the negative effect of decreasing temperature is similar to the negative effect of mixing with low TA and DIC freshwater end-members, enhancing a decrease in $p\text{CO}_2$.

6   Summary and Conclusions

Here, we introduced a new regional biogeochemical model GOA-COBALT and evaluated the model's skill in simulating seasonal and interannual inorganic carbon patterns. Our model set-up is unique because it includes a moderately-high resolution,

10   three-dimensional regional ocean circulation model, a complex ecosystem model with an ocean carbon cycle, a high-resolution terrestrial hydrological model, and is forced with reanalysis products to simulate interannual variability and long term changes over the past 30+ years. In addition, we used available TA and DIC observations to parameterize seasonal concentrations of DIC and TA in the freshwater forcing.

Comparison with a limited amount of *in situ* inorganic carbon observations showed that the model is able to reproduce the

15   general hydrographic and inorganic carbon patterns in spring and fall. The GOA-COBALT was particularly successful in reproducing the depth of the aragonite saturation horizon, showing oversaturated conditions across large parts of the shelf in May and undersaturation across the shelf in September. The largest deficiency of GOA-COBALT may be that it generally overestimates Chl-α concentrations and that the Chl-α peak is simulated in June rather than in May. Interestingly, results of the inorganic carbon data - model analysis suggest, however, that DIC is not drawn down enough by spring biological production.

20   This contradiction may be due to an underestimation of modelled C:Chl-α ratios, which are dependent on light and nutrient limitation (Geider et al., 1997; Stock et al., 2014). Modelled C:Chl-α of large and small phytoplankton on the shelf are < 30 g/g and < 40 g/g respectively. Measurements of C:Chl-α of small phytoplankton in the GOA suggest a median C:Chl-α of 41 g/g during intense and early blooms and 76 g/g in low Chl-α conditions (Strom et al., 2016). Equivalent data for large phytoplankton does not exist.

25   GOA-COBALT is more successful in reproducing surface inorganic carbon patterns in September, when freshwater is the dominant driver of the system. In addition to its capacity of resolving prominent seasonal inorganic carbon variability, the model also resolved some of the observed interannual differences present in the publicly available five year-long inorganic carbon Seward Line record. However, the GOA-LME is largely undersampled despite its large spatial and temporal variability. The limited amount of inorganic carbon observations that cover one transect in May and September, and include a few newer

30   nearshore time-series are not sufficient to properly evaluate the model. More observations are needed to test the model's skills in areas other than the Seward Line and during other times of the year. While some of the year to year differences seen in Figures 7 and 8 may be a result of aliasing the spatial and temporal variability that exists during the cruises, the most prominent anomalies have been described in the past. For example, as a result of anomalies in coastal runoff, winter cooling, stratification,
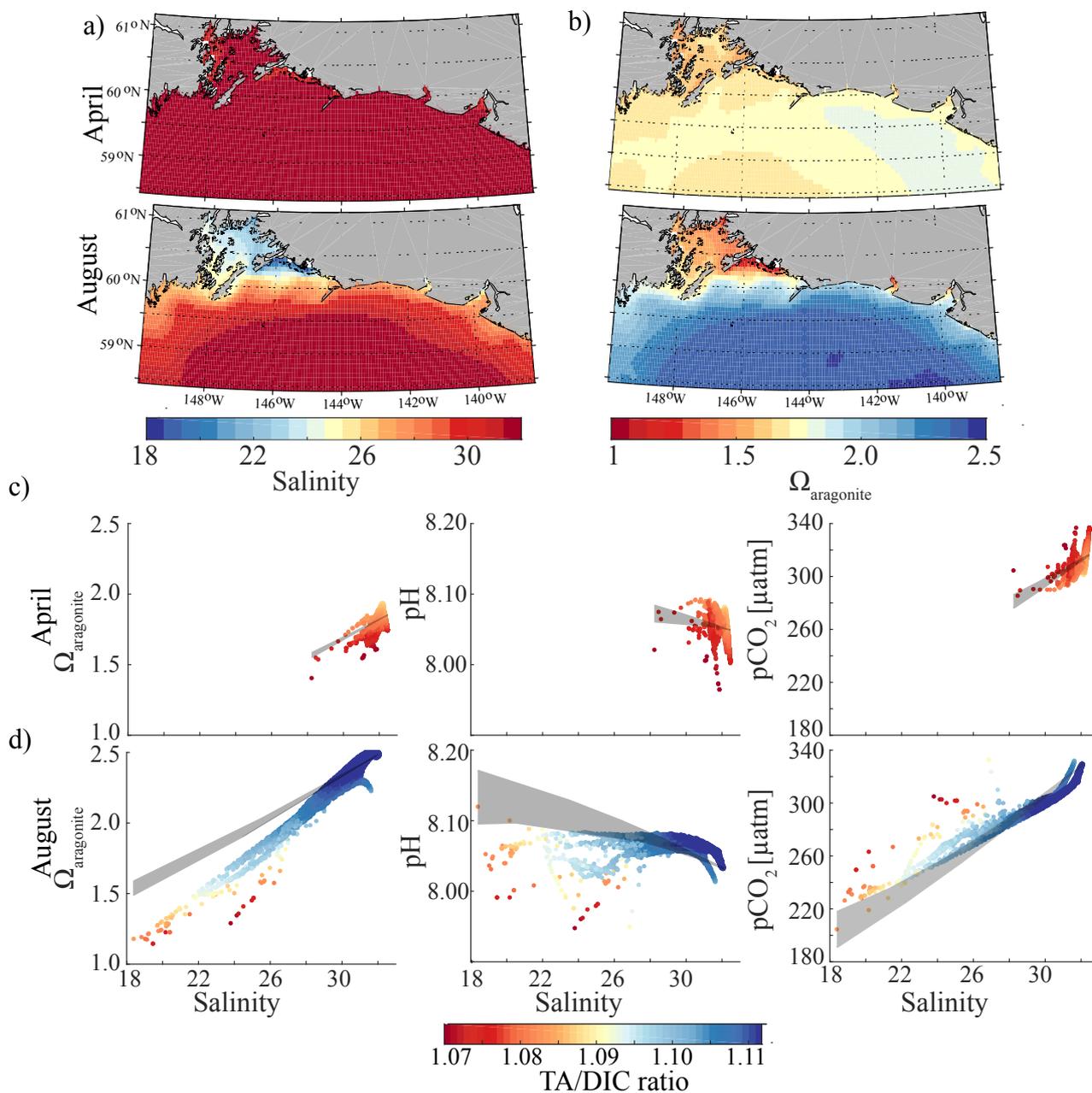
Figure 11. Map of climatological mean (1980 -2013) of a) surface salinity and b) surface Ω_{arag} in April (top) and August (bottom). Plots of surface Ω_{arag} (left), pH (middle), and $p\text{CO}_2$ [μatm](right) vs. salinity as a function of TA/DIC ratio in c) April and d) August. The shaded area indicates the range of Ω_{arag}, pH, and $p\text{CO}_2$, respectively, if changes of the respective variable would only arise from variations in salinity and from the dilution effect of freshwater on DIC and TA.
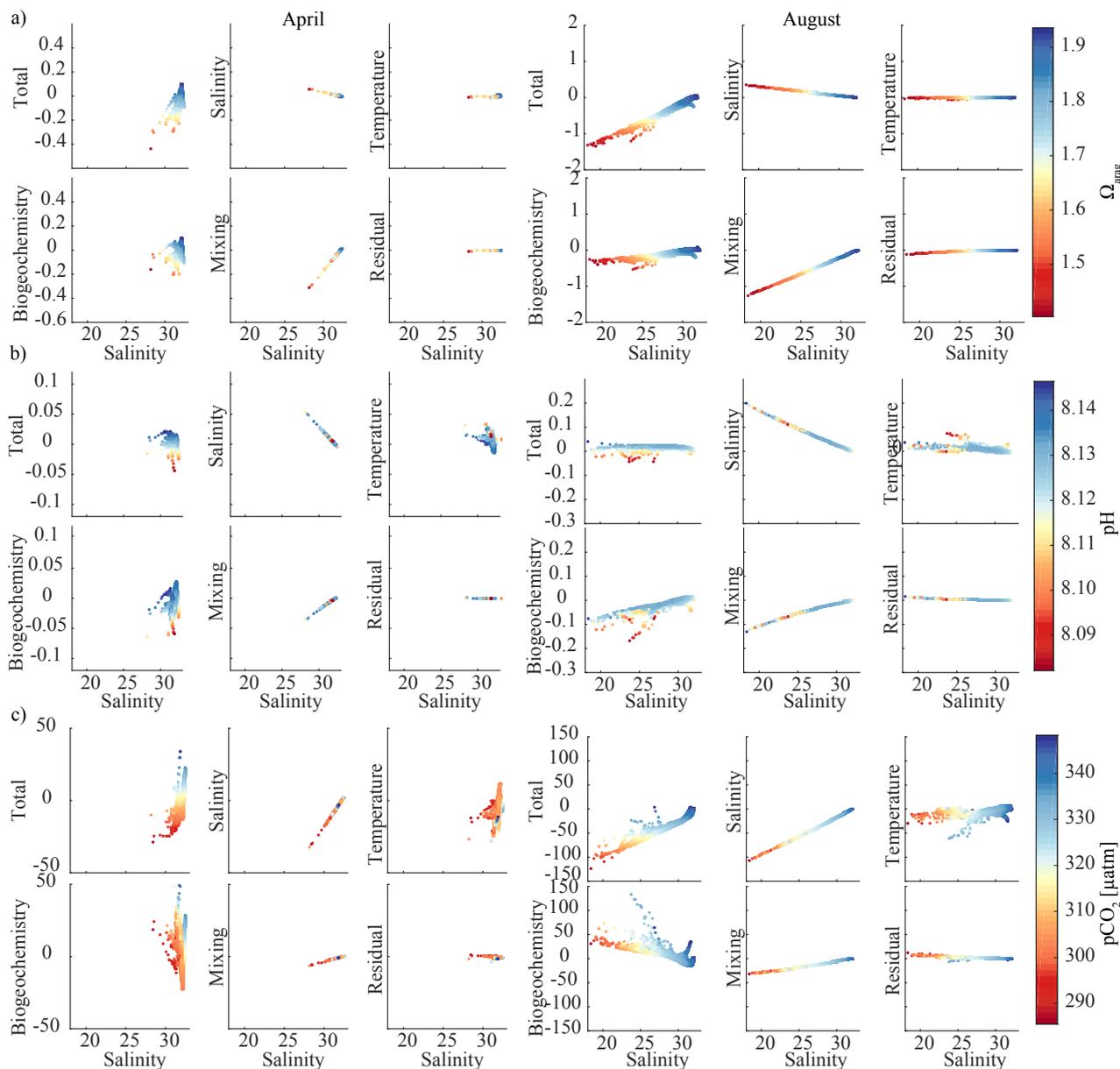
Figure 12. Component distributions of the linear Taylor decomposition of surface a) Ω_{arag}, b) pH, and c) pCO$_2$ versus salinity in April (left) and August (right). Following Rheuban et al. (2019) components are total perturbation from the oceanic end-member (Salinity > 32), perturbations due to salinity, temperature, biogeochemistry, freshwater mixing, and an estimated residual term.

and winds, sea surface temperature was by 1.5 °C colder in spring 2006, 2007, and 2008 than average spring temperatures, and the lowest since early 1970 (Janout et al., 2010). In contrast, as a result of warm air temperatures, GOA sea surface temperatures in 2009 were anomalously high (Danielson et al., 2019).

The relationship between Ω_{arag}, pH, and pCO$_2$ in freshwater influenced coastal waters is different than what is found elsewhere. Similarly to what Evans et al. (2014) observed *in situ* near a tidewater glacier in Prince William Sound, our freshwater-inorganic carbon analysis showed that low Ω_{arag} values are not always accompanied by low pH and high pCO$_2$ values in this glacially influenced environment. This decoupling of the three inorganic carbon parameters is driven by the additive effects of

5   salinity and mixing with low TA and DIC end-members. Whether the freshwater induced low surface pCO$_2$ caused enhanced CO$_2$ uptake in coastal areas, as speculated by Evans et al. (2014), could not be confirmed here and will need additional investigation.

The strong influence of freshwater on the inorganic carbon system emphasizes the importance of choosing the right DIC and TA freshwater endmembers in order to correctly model the inorganic carbon dynamics in this area. However, there is a lack of

10   biogeochemical data that describes the composition of different freshwater sources on seasonal timescales. The biogeochemical composition of freshwater is influenced by its exposure to basal rock and therefore by the pathways and duration it takes, and processes it undergoes until it is introduced to the ocean. Current sources of freshwater in the GOA are tidewater glaciers, proglacial streams, and non-glacial streams, each exposing the water to basal rock, soils and the atmosphere differently. As this region continuous to rapidly deglaciate (Arendt et al., 2002; Larsen et al., 2007; O'Neel et al., 2005), not only the amount

15   of freshwater discharge will increase (Beamer et al., 2017), but most of the glaciers surrounding the GOA will recede away from the ocean and into higher elevations (Huss and Hock, 2015), resulting in a change of the biogeochemical composition of freshwater. This and previous studies have shown that freshwater discharge is the largest driver of the nearshore inorganic carbon dynamics in summer and fall, and is known to exacerbate the effects of ocean acidification (Evans et al., 2014; Siedlecki et al., 2017; Pilcher et al., 2018). Understanding the composition of freshwater sources is, therefore, particularly important for

20   the study of long-term trends of ocean acidification in the GOA-LME. As a first best approach, we used seasonal observations of DIC and TA from Kenai River (proglacial river) and applied it to every freshwater point source across the domain, however this approach is likely masking out large differences in the biogeochemical composition of freshwater input that could have implications on the coastal inorganic carbon system.

The only other regional biogeochemical model with an oceanic carbon cycle (Siedlecki et al., 2017; Pilcher et al., 2018) used

25   Royer (1982)'s monthly riverine input time-series and applied it equally to the top cell along the coast, masking out the large spatial and interannual variability of freshwater input along the GOA coast. Furthermore, Siedlecki et al. (2017) simulated year 2009, with only one year of spinup and with DIC and TA boundary conditions based on salinity relationships. Their static freshwater TA concentration to represent a tidewater glacier was 650 μmol kg^{-1}, while freshwater runoff did not affect coastal DIC at all. In comparison, our GOA-COBALT model and the Siedlecki et al. (2017) both simulate the lowest surface Ω_{arag} in

30   winter and highest in summer (Figure 13). Similarly, the aragonite saturation horizon is shallowest in summer and fall in both models. In our model, the aragonite saturation horizon is shallower throughout the year, thereby causing aragonite undersaturation across wider areas on the shelf. A thorough model inter-comparison study would need to be done to understand how the new features of our model set-up have affected and potentially improved the modeling results over Siedlecki et al. (2017) and Pilcher et al. (2018).

35   Our simulation results give new and important insights for months of the year that lack *in situ* inorganic carbon observations.
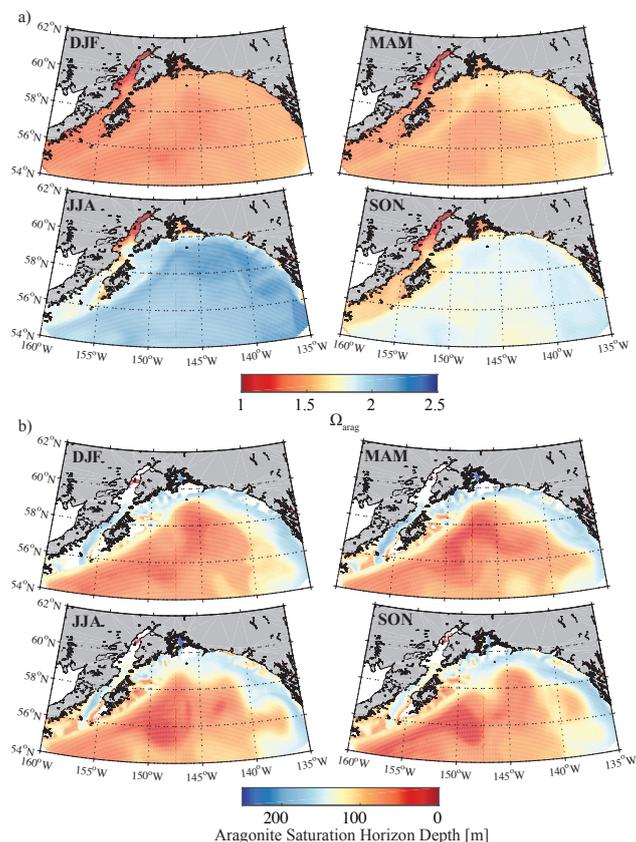
Figure 13. Climatological mean a) surface aragonite saturation state (Ω_{arag}) and b) aragonite saturation horizon (m) December, January, and February (DJF), March, April, and May (MAM), June, July, and August (JJA), and September, October, and November (SOM).

For example, the majority of the near-bottom water along the Seward Line is seasonally under-saturated with regard to arago-nite between June and January. Such long and reoccurring aragonite undersaturation events may be harmful to CO_2 sensitive organisms. Furthermore, between January and March, conditions are unfavorable for pteropods across the entire water column with aragonite saturation state < 1.5. This study also made it apparent that more observations across the shelf and different

5 times of the year are needed in order to improve the model and evaluate its skills in areas other than near Seward.

Future work will focus on the progression of ocean acidification and climate change and impacts on the inorganic carbon chemistry. We also anticipate this simulation to be a useful tool for the study of the duration and intensity of extreme events and climate-ocean teleconnections. With increased confidence in the model, another logical next step would be to guide the future expansion, diversification, and optimization of OA observing systems in the GOA-LME.

Code and data availability. The code and forcing files are available on zenodo.org under DOI's 10.5281/zenodo.3647663, 10.5281/zenodo.3661518, and 10.5281/zenodo.3647609. The model output is publicly available (https://doi.org/10.24431/rw1k43t) and can be visualized with a user-friendly web interface. This is a product of our collaboration with Axiom Data Science.

Author contributions. K.H., C.H., S.D., and C.S. developed and tuned the model system for the GOA. C.H. and B.I. prepared the model
5 output. C.H., C.S., B.I., and S.C.D. analyzed the model output. R.D. ported the COBALT code into ROMS. All authors commented on the manuscript.

Competing interests. We declare that no competing interests are present.

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