Chapter 16. Coastal Oceans and Continental Shelves

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Key Findings

1. **Since the first SOCCR report significant progress has been made in quantifying coastal carbon dynamics because of expanding carbon-observing networks, improved model capabilities and better national and international coordination activities. Despite this, critical uncertainties remain.**

2. **Available estimates of air-sea carbon fluxes indicate that the North American margins act as a net sink for atmospheric CO_2. This net uptake is primarily driven by fluxes in high latitude regions. The magnitude of the net flux is not well constrained.**

3. **Existing data networks and high-resolution models are now available to begin the construction of coastal carbon budgets, but separation of anthropogenic carbon contributions from the natural background remains elusive.**

4. **Most efforts have focused on quantifying air-sea exchange of CO_2, but some studies have estimated other key fluxes, such as the exchange between shelves and the open ocean, by combining observations and models.**

5. **While all North American coastal regions require further observations and model development in order to facilitate full accounting of carbon budgets and human-caused carbon contributions, the Gulf of Alaska, the North American Arctic and the Central American Isthmus are particularly lagging.**

6. Further efforts are required to ensure sustained monitoring, synthesis of available observations, expanded model capabilities for describing the biogeochemistry and ecosystems, and forecasts of future scenarios for North American coastal oceans.

16.1 Introduction

Along ocean margins, the atmospheric, terrestrial, sedimentary, and deep-ocean carbon reservoirs come in contact with one another, which results in quantitatively significant carbon exchanges. Anthropogenic activities lead to secular trends in these exchanges. The drivers include rising atmospheric CO_2 levels, climate-driven changes in atmospheric forcing (e.g., winds and heat fluxes) and the hydrological cycle (e.g., freshwater input from rivers), and changes in riverine and atmospheric nutrient inputs from agricultural activities and fossil fuel burning. The collective impact of these factors on carbon processing and exchanges along ocean margins is complex and difficult to predict.

Here we focus on two particularly pressing issues within the much broader topic of carbon cycling along ocean margins: (1) the uptake of anthropogenic carbon and subsequent export to the deep ocean, and (2) drivers of and trends in coastal ocean acidification. The first is relevant for the overall quantification of the ocean's uptake of anthropogenic CO_2. The second is of direct relevance for coastal ecosystem health, fisheries and aquaculture.

Within this chapter we will use two different terms when referring to ocean margins: (1) coastal oceans, which we define as non-estuarine waters within 200 nautical miles (370 km) of the coast and (2) continental shelves, which refer to the submerged margins of the continental plates, operationally defined here as regions with water depths shallower than 200 m (indicated by xx color in **Figure 16.1**). Both definitions overlap, but there are important reasons for considering both. Along passive margins with broad shelves like the North American (NA) Atlantic coast (NAAC), the continental shelf is the relevant spatial unit for discussing carbon fluxes. Along active margins with narrow shelves like the NA Pacific Coast (NAPC), a larger region than just the shelf needs to be considered for a meaningful discussion of coastal carbon dynamics. The 200 nautical mile limit chosen here corresponds to the Exclusive Economic Zone and is thus pertinent to carbon accounting at the national level.

This chapter builds on and extends several previous synthesis and planning activities, including a report by the North American Continental Margins Working Group (Hales et al. 2008), the first SOCCR report (Chaves et al. 2007) and activities within the North American coastal interim synthesis (Benway et al. 2014, Najjar et al. 2012, Robbins et al. 2009, Alin et al. 2012). The first SOCCR report (Chavez et al. 2007) concluded that carbon fluxes for North American ocean margins were not well quantified because of insufficient observations and the complexity and highly localized spatial variability of coastal carbon dynamics. The report was inconclusive as to whether North American coastal waters act as an overall source or sink of atmospheric CO_2.

Our objective in this chapter is to provide a review and synthesis of recent findings with respect to coastal carbon uptake and ocean acidification for the margins of North America (NA). First, we briefly summarize the key variables and fluxes relevant to carbon budgets for coastal waters, outline mechanisms by which carbon can be removed from the atmospheric reservoir, and describe means for quantifying the resulting carbon removal (Section 16.2). Then we review the available research relevant to carbon budgets for NA coastal waters by region and assess whether enough information is available to derive

robust estimates of carbon export (Section 16.3). We then discuss climate-driven trends in coastal carbon fluxes and coastal ocean acidification (Section 16.4), briefly review current carbon-observing networks (Section 16.5) and provide conclusions (Section 16.5).

16.2 Current Understanding of Carbon Fluxes and Stocks

Carbon is present in various inorganic and organic forms in coastal waters. The dissolved inorganic species include aqueous CO_2 (a combination of dissolved CO_2 and carbonic acid), bicarbonate ion (HCO_3^-) and carbonate ion (CO_3^{2-}), which are collectively referred to as dissolved inorganic carbon (DIC), and methane (CH_4). The major particulate inorganic species is calcium carbonate ($CaCO_3$), also referred to as particulate inorganic carbon (PIC). Carbon is also present in various dissolved and particulate organic forms (DOC and POC). In shelf waters, the reduced carbon pool or total organic carbon pool (TOC) represents roughly 2 to 5% of the total carbon stock (Reference needed) and DOC constitutes over 90 to 95% of this TOC (Reference needed).

Carbon is constantly transferred between these different pools and exchanged across the interfaces that demarcate coastal waters: the land-ocean interface, the air-sea interface, and the interface between coastal and open ocean waters (**Figure 16.2**). The internal carbon transformations within coastal regions include photosynthetic primary production, respiration, transfers between lower and higher trophic levels of the food web, exchanges between sediment and overlying water, biogeochemical processes in the sediment, and the formation and dissolution of $CaCO_3$. A major internal transformation is the conversion of DIC into POC and DOC (primary production). In mid to high latitudes, this transformation occurs primarily in spring. Over the course of summer and fall, as primary production decreases, respiration throughout the water column returns some of the organic carbon into inorganic forms, primarily DIC. Some POC settles out of the water column and becomes incorporated into the sediments where most of this material is respired through a range of different redox processes that produce DIC and, under certain circumstances (i.e., in the relative absence of electron acceptors other than CO_2), methane. Both DIC and methane are released back into the overlying water. POC that is not respired (and referred to as refractory) can be buried in the sediments and stored for very long periods of time. Some organisms form internal or external body structures from $CaCO_3$, which ultimately become incorporated into the sediments and dissolve or are buried. We will refer to this long-term storage of POC and PIC in coastal sediments as permanent burial.

A major carbon exchange process along the ocean margin is the flux of CO_2 across the air-sea interface. The annual cycle of this flux is driven by: (1) the warming and cooling of seawater, which affects the solubility of CO_2; (2) the under- or oversaturation of CO_2 that results from primary production, respiration, and $CaCO_3$ precipitation and dissolution; (3) the transport of DIC to and from the ocean surface (e.g., upwelling and convection); and (4) factors that influence the resistance to gas exchange across the air-sea interface (e.g., winds and sea ice extent). The annual cycles of primary production, respiration and air-sea CO_2 flux tend to be of larger magnitude and more variable in coastal waters than in the open ocean (Liu et al. 2010, Muller-Karger et al. 2005, Thunell et al. 2007, Xue et al. 2016) and more pronounced in high latitudes. Other important exchange fluxes are carbon input from land via rivers and estuaries (Chapter 15), carbon inputs from tidal wetlands (Chapter 15), and exchanges between the coastal and open ocean across the continental shelf break or the operationally defined open-ocean boundary of the coastal ocean. Net removal of carbon from direct interaction with the atmospheric reservoir can occur by export to the deep ocean or by permanent burial in coastal sediments.

Although continental shelves make up only 7-10% of the global ocean surface area, they are estimated to contribute up to 30% of primary production and 30-50% of inorganic and 80% of organic carbon burial (Gattuso et al. 1998, Dunne et al. 2007). Given this, continental shelves have been argued to contribute disproportionately to the oceanic uptake of CO_2 (Muller-Karger et al. 2005, Liu et al. 2010, Cai 2011).

Carbon export, referring to the flux of organic and inorganic carbon from coastal waters to the deep ocean, can occur through the so-called "Continental Shelf Pump"—a term coined by Tsunogai et al. (1999) after they observed a large uptake of atmospheric CO_2 in the East China Sea. There are two distinct mechanisms underlying the Continental Shelf Pump (Fennel 2010). The first is physical in nature and thought to operate in mid- and high-latitude systems. In these shelf systems, surface water is cooled more strongly than in the adjacent open ocean because it is not subject to deep convection and, since the solubility of CO_2 increases with decreasing temperature, the cold shelf water experiences a larger influx of atmospheric CO_2. If this dense and carbon-rich water is transported off the shelf, it will sink due to its higher density and the associated carbon will be exported to the deep ocean. The second mechanism relies on biological processes that concentrate carbon below the seasonal pycnocline (i.e., photosynthetic production of organic carbon and subsequent sinking). If the carbon-rich water below the seasonal pycnocline is moved off the shelf horizontally, carbon could potentially be exported if this water is transported or mixed below the seasonal thermocline. The depth to which the shelf-derived carbon can be exported will be different for POC, which will sink, and DOC and DIC, which would primarily be advected laterally. Both of the above mechanisms for carbon export critically depend on physical transport of carbon-rich water off the shelf.

Carbon export from coastal waters to the deep ocean cannot be accurately quantified through direct observation. Thus, the only available estimates of such export are indirect, using mass balances of POC and dissolved oxygen (e.g., Hales et al. 2006) and model estimates (e.g., Fennel and Wilkin 2009, Xue et al. 2013, Turi et al. 2014, Fiechter et al. 2014, Bourgeois et al. 2016, Mannino et al., 2016). If the total carbon inventory in a coastal system is constant over a sufficiently long time scale (on the order of years), carbon export can be inferred from the sum of all other exchange fluxes across the system's interfaces over that same period. The influx of carbon from land and wetlands, its net exchange across the air-sea interface, lateral inputs due to advection, and any removal through permanent sediment burial must be balanced by export to the open ocean. The accuracy of the inferred export flux directly depends on the accuracy of the other flux estimates and the assumption of a constant carbon inventory. Quantification of internal transformation processes (e.g., primary and secondary production, respiration) does not directly enter this budgeting approach, but can elucidate the processes that drive fluxes across interfaces.

Presently available estimates of carbon fluxes across coastal interfaces come with significant uncertainties (Birdsey et al. 2009). These uncertainties are due to a combination of small-scale temporal and spatial variability, which is undersampled by presently available means of direct observation, and regional heterogeneity, which makes it difficult to scale up local observations. Geographical differences due to variations in shelf width, the presence or absence of large rivers, seasonal ice cover, and latitude (through its modulation of annual temperature and productivity cycles and hydrographic patterns via the Coriolis force; Sharples et al., in revision) all contribute to differences in regional carbon budgets and export. Below we will describe the regional characteristics of NA coastal waters, how these influence carbon dynamics on the regional scale and compile available estimates of carbon fluxes in an attempt to estimate export.

The underlying motivation for quantifying permanent burial of carbon and export of carbon from the coastal to the deep ocean is that both remove anthropogenic carbon from the atmospheric reservoir. The relevant carbon flux in this context is the burial or export of anthropogenic carbon, not total burial or export. Thus far we have considered total carbon fluxes, which are comprised of the anthropogenic flux component superimposed on a natural background flux. Only total fluxes—the sum of anthropogenic and background fluxes—can be observed directly. Distinction between anthropogenic fluxes and the natural background has not been attempted in regional observational or modeling studies. Observation-based estimates of the global anthropogenic uptake have been made by Sabine et al. (2004) and Sabine and Tanhua (2010); the study by Bourgeois et al. (2016) is the first to estimate coastal anthropogenic carbon uptake in a global model.

16.3 Coastal Carbon Fluxes Around North America

Table of acronyms

NAAC	North American Atlantic Coast
GOM	Gulf of Maine
MAB	Middle Atlantic Bight
SAB	South Atlantic Bight
NAPC	North American Pacific Coast
GAK	Gulf of Alaska
CCS	California Current System
CAI	Central American Isthmus (Pacific side)
GoMex	Gulf of Mexico
NAA	North American Arctic
CAA	Canadian Arctic Archipelago

16.3.1 North American Atlantic Coast

The NAAC is a passive margin shelf that extends from the southern tip of Florida to the continental shelf of the Labrador Sea. The shelf is several hundreds of km wide in the north (Labrador shelf and Grand Banks) but narrows progressively in the Middle Atlantic Bight (MAB, between Cape Cod and Cape Hatteras) and the South Atlantic Bight (SAB, south of Cape Hatteras) where its width is reduced to several tens of km. Two major semi-enclosed bodies of water are the Gulf of Maine (GOM) and the Gulf of St. Lawrence. Important rivers and estuaries north of Cape Hatteras include the St. Lawrence River and Estuary, the Hudson River, Long Island Sound, Delaware Bay and Chesapeake Bay. South of Cape Hatteras the coastline is characterized by small rivers and marshes. The SAB is influenced by the Gulf Stream, which flows northeastward along the shelf edge before detaching at Cape Hatteras and meandering eastward into the open North Atlantic Ocean. North of Cape Hatteras, shelf circulation is influenced by the confluence of the southwestward flowing fresh and cold shelf-break current (a limb of the Labrador Current) and the warm and salty Gulf Stream (Loder et al. 1998). Since shelf waters north of

Cape Hatteras are sourced from the Labrador Sea they are relatively cold, fresh and carbon-rich, while slope waters (those located between the shelf break and the northern wall of the Gulf Stream) are a mixture of Labrador Current and Gulf Stream water. Exchange between the shelf and open ocean across the shelf break is impeded by the presence of the Gulf Stream south of Cape Hatteras and by shelf break jets and fronts north of Cape Hatteras.

Air-sea fluxes of CO_2 exhibit a large-scale latitudinal gradient along the NAAC and significant seasonal variability. The Scotian Shelf is a net source of CO_2 to the atmosphere (Shadwick et al. 2010, 2011, Thomas et al. 2013), the GOM is a weak net source (Vandermark et al. 2011), and the MAB (DeGrandpre et al. 2002, Signorini et al. 2013) and SAB (Jiang et al. 2008) are weak net sinks. The change from net source in the Scotian Shelf region to net sink in the MAB arises because the properties of shelf water are modified during its southwestward flow by warming, inflows of riverine and estuarine waters (Salisbury et al. 2008, 2009), and exchange with the open North Atlantic across the shelf break (Cai et al. 2010, Wang et al. 2013). Outgassing of CO_2 on the Scotian Shelf is driven primarily by warming of cold, carbon-rich shelf water, which still carries a pronounced signature of its Labrador Sea origin (Rutherford et al., in prep.). The GOM, which is deeper than the Scotian Shelf and the MAB, and connected to the open North Atlantic through a relatively deep channel, is characterized by a mixture of cold, carbon-rich shelf waters and warmer, saltier slope waters. Shelf water in the MAB is sourced from the GOM and thus is a mixture of shelf and slope water. Freshwater inputs from several rivers and estuaries, and exchange across the shelf break further modify shelf water in the MAB (Wang et al. 2013).

Shelf water in the SAB is distinct from that in the MAB and has no trace of Labrador Current water; instead it is similar in its characteristics to the Gulf Stream but its carbon signature is modified by significant organic and inorganic carbon and alkalinity inputs from coastal marshes (Jiang et al. 2013, Wang et al. 2005, Wang and Cai 2004, Cai et al. 2003). Herrmann et al. (2015) estimated that 59% of the 3.4 Tg yr^{-1} of organic carbon exported from US east coast estuaries occurs in the SAB. The subsequent respiration of this organic matter and direct outgassing of marsh-derived carbon makes the near-shore regions a significant CO_2 source almost year-round. Despite the carbon inputs from marshes, uptake of CO_2 on the mid and outer shelf during the winter months is large enough to balance CO_2 outgassing in the other seasons and on the inner shelf making the SAB overall a weak net sink (Jiang et al. 2008).

North of Cape Hatteras, CO_2 dynamics are characterized by strong seasonality with solubility driven uptake by cooling in winter and biologically driven uptake in spring followed by outgassing in summer and fall due to warming and respiration of organic matter (Vandemark et al. 2011, Shadwick et al. 2010, 2011, Thomas et al. 2013, DeGrandpre et al. 2002, Signorini et al. 2013, Wang et al. 2013). Hydrography and CO_2 dynamics on the Scotian Shelf are influenced by the significant fresh water input from the St. Lawrence River. Riverine inputs of carbon and nutrients are relatively small in the GOM, but can cause local phytoplankton blooms, CO_2 drawdown, and low pH conditions (Salisbury et al., 2008, 2009). Riverine and estuarine inputs become more important in the MAB with discharges from Chesapeake Bay and the Delaware, Hudson and Connecticut Rivers (Wang et al. 2013). South of Cape Hatteras seasonal phytoplankton blooms do not occur regularly and biologically driven CO_2 uptake is less pronounced than further north (Wang et al., 2013), although sporadic phytoplankton blooms occur due to intrusions of high-nutrient subsurface Gulf Stream water (Wang et al. 2005, 2013). The influence of riverine inputs is small and localized in the SAB (Wang and Cai 2004, Wang et al. 2005, Cai and Wang 1998).

Regional biogeochemical models reproduce the large-scale patterns of air-sea CO_2 flux with oceanic uptake increasing from the SAB to the GOM (Fennel et al. 2008, Previdi et al. 2009, Cahill et al. 2016). These model studies elucidate the magnitude and sources of interannual variability as well as long-term trends in air-sea CO_2 fluxes. Previdi et al. (2009) investigated opposite phases of the North Atlantic Oscillation (NAO) and found that the simulated air-sea flux in the MAB and GOM was 25% lower in a high-NAO year compared to a low-NAO year. In the MAB the decrease was primarily due to changes in wind forcing, while in the GOM changes in surface temperature and new production were more important. Cahill et al. (2016) investigated the impact of climate-driven warming and trends in atmospheric forcing (primarily wind) on air-sea CO_2 flux (without considering the atmospheric increase in CO_2). Their results suggest that warming and changes in atmospheric forcing have modest impacts on air-sea CO_2 flux in the MAB and GOM compared to the SAB where surface warming turns the region from a net sink into a net source of CO_2 to the atmosphere. Model studies also illustrate the effects of interactions between biogeochemical transformations in the sediment and the overlying water column on carbon fluxes. For example, Fennel et al. (2008) show that the effective alkalinity flux resulting from denitrification in NAAC sediments reduces the simulated ocean uptake of CO_2 by 6% compared to a simulation without sediment denitrification.

16.3.2 North American Pacific Coast

The North American Pacific Coast (NAPC) extends from Panama to the Gulf of Alaska (GAK) and is an active margin with varying shelf widths. The continental shelf is narrow along the coasts of California, Oregon and Washington, ranging in width from 5 km in the south to 10 km in the north, but widens significantly in the GAK where shelves extend up to 200 km offshore. In the GAK, freshwater and tidal influences exert strong influence on the cross-shelf exchange, and the shelf is dominated by downwelling circulation. The region from Vancouver Island to Baja California is a classic eastern boundary upwelling region and influenced by the California Current System (CCS). Winds drive a coastal upwelling circulation characterized by equatorward flow in near-surface jets and associated eddies and fronts that extend offshore, particularly off the coast of California. The northern CCS experiences strong freshwater influences and seasonality in wind forcing that diminish in the southern CCS. In addition to the Columbia River and the Fraser River a variety of small mountainous rivers with highly variable discharge supply freshwater. The Central American Isthmus (CAI) runs from Panama to the southern tip of Baja California and experiences intense, persistent wind events, large eddies, and high waves, which combine to produce upwelling and strong nearshore mixing (Franco et al. 2014, Chapa-Balcorta et al. 2015). In addition to alongshore winds, strong seasonal wind jets that pass through the Central American cordillera create upwelling hotspots and drive production during boreal winter months in the Gulfs of Tehuantepec, Papagayo and Panama (Chelton et al. 2000a,b, Chapa-Balcorta et al. 2015).

Observation-based studies of air-sea CO_2 flux suggest that the coastal ocean from Baja to GAK ranges from a weak to moderate sink of atmospheric CO_2 over this broad longitudinal range. Central California coastal waters have long been understood to have near neutral air-sea CO_2 exchange due to large and counter-balancing periods of efflux during upwelling conditions and influx during periods of relaxation and high primary productivity, and that this pattern is strongly modulated by El Niño-La Niña conditions (Friederich et al. 2002). Hales et al. (2005) used seasonal data to estimate 20 mmol CO_2 m^{-2} d^{-1} uptake by Oregon coastal waters, which is 15 times larger than the global mean (1.3 mmol CO_2 m^{-2} d^{-1}). Using data with greater temporal coverage, Evans et al. (2011) revealed how large flux events can significantly alter the estimation of net exchanges for the Oregon shelf. Due to capturing a large and short-lived efflux

event, their annual estimate was only -0.7 ± 18.7 mmol m^{-2} d^{-1} for this same setting. The disparity between these results illustrates the importance of basing regional flux estimates on data that are well resolved in time and space. Capitalizing on the increased and more uniform spatiotemporal coverage of satellite data, Hales et al. (2012) estimated an annual mean air-sea flux of 1.8 mmol CO_2 m^{-2} d^{-1} between 22° and 50°N within 370 km of shore. The most northern estimates for the NAPC by Evans et al. (2012) and Evans and Mathis (2013), determined influxes near 6 mmol CO_2 m^{-2} d^{-1} for British Columbian coastal waters shoreward of the 500 m isobath and near 4 mmol CO_2 m^{-2} d^{-1} for GAK coastal waters shoreward of the 1500 m isobaths.

Models for the upwelling region (Turi et al. 2014, Fiechter et al. 2014) reproduce the pattern of CO_2 outgassing near shore, CO_2 uptake further offshore, as well as the along-shore pattern of the air-sea CO_2 flux. These models also illustrate the intense eddy-driven variability near shore. Turi et al. (2014) predict essentially neutral CO_2 fluxes of 0.05±0.20 mol CO_2 m^{-2} yr^{-1} for the region from 30° to 46°N extending 800 km of shore. Fiechter et al. (2014) predict a sink of atmospheric CO_2 of 6 Tg C yr^{-1} for the region from 35° to 45°N within 600 km of shore. Both models simulate strong outgassing within the first 100 km of shore, driven by intense upwelling of nutrient- and carbon-rich water, compensated by biologically driven CO_2 uptake from the atmosphere as upwelled nutrients are consumed by photosynthesis within several hundreds of km of shore. The disagreement in mean flux estimates may partly result from different choices of averaging region and period, and differences in model forcing, such as the climatological forcing in Turi et al. (2014) versus realistic variability in Fiechter et al. (2014). It should be noted, however, that observations for the Oregon shelf by Evans et al. (2015) showed intense summer upwelling that led to strong outgassing with pronounced variability in air-sea fluxes, but only found weak stimulation of primary production. They hypothesized that nutrient-rich waters might be subducted offshore at convergent surface temperature fronts before nutrients are fully consumed by primary producers.

The net exchange of CO_2 with the atmosphere across the NAPC is characterized by strong spatial and temporal variation and reflects complex interactions between degassing of nutrient- and carbon-rich upwelled waters near the coast and biological uptake of upwelled nutrients further offshore.

Less is known about the air-sea flux of methane along the NAPC margin. Recent studies inventoried sedimentary sources of methane hydrates, derived from terrestrial and coastal primary production, and suggested that extensive deposits along the Cascadia margin are beginning to destabilize due to climate warming (Hautala et al. 2014, Johnson et al. 2016).

Cross-shelf exchange of carbon occurs in the California Current System mostly in response to wind-driven circulation and eddies, but river plumes and tides have been shown to also increase offshore transport in the northern CCS (Barth et al. 2002, Hales et al. 2006). Uncertainties in published estimates are high, and range from very small (Pennington et al. 2010, Ianson and Allen 2002) to very high fractions of primary production (Hales et al. 2005, Turi et al. 2014), again due to the large spatial and temporal variability in the region. Comparison between published studies and more recent results is difficult because of the different spatial domains used for describing the key carbon fluxes.

16.3.3 Gulf of Mexico

The Gulf of Mexico (GoMex) is a semi-enclosed marginal sea at the southern coast of the conterminous United States. It covers approximately 1.6 x 10^6 km^2, with waters deeper than the continental shelf

covering about 8.6 x 10^5 km^2. The passive margin shelves of the northern GoMex are relatively wide (up to 250 km west of Florida) but, in contrast to the NAAC, shelf waters are not separated from the open-ocean waters by shelf-break fronts or currents. Ocean water enters the Gulf mainly through the Yucatan Channel, where it forms the northeastward meandering Loop Current (LC), which sheds anticyclonic eddies and exits the Gulf through the Florida Straits (Rivas et al. 2005, Muller-Karger et al., 2015). While shelf circulation is primarily influenced by local wind and buoyancy forcing, outer shelf regions are at times influenced by LC eddies that impinge on and interact with the shelf (Lohrenz and Verity 2004). Riverine input is substantial in the northern GoMex, where the Mississippi-Atchafalaya river system delivers large loads of freshwater, nutrients and sediments.

Observational estimates indicate that the GoMex as a whole is a weak net sink of atmospheric CO_2 with an annual average of 0.19±0.08 mol CO_2 m^{-2} yr^{-1} (Robbins et al. 2014). Robbins et al. (2014) also provide flux estimates for smaller shelf regions, namely the West Florida Shelf (WFS), the northern Gulf shelf, the western Gulf shelf and the Mexico shelf as follows. The WFS and western shelf act as sources to the atmosphere with estimated annual average fluxes of -0.37±0.11 and -0.18±0.05 mol CO_2 m^{-2} yr^{-1}, respectively. The northern Gulf acts as a sink with an estimated flux of 0.44±0.37 mol CO_2 m^{-2} yr^{-1} and the Mexican shelf is almost neutral with an estimated flux of 0.09±0.05 mol CO_2 m^{-2} yr^{-1}. Huang et al. (2015) estimated a larger uptake on the northern Gulf shelf with 0.96±3.7 mol CO_2 m^{-2} yr^{-1} (about twice the estimate of Robbins et al.) and also reported a much larger uncertainty. The overall carbon exchanges in the Gulf can be modified significantly based on interannual variability in wind, temperature, and precipitation (Muller-Karger et al., 2015).

Model-simulated air-sea CO_2 fluxes by Xue et al. (2016) agree relatively well with the estimates of Robbins et al. (2014) in that the same spatial pattern is reproduced, though the simulated Gulf-wide uptake (0.71±0.54 mol CO_2 m^{-2} yr^{-1}) is larger. This discrepancy is largely due to a larger simulated sink in the open Gulf. Also, the standard deviations of the model-simulated fluxes by Xue et al. (2016) are much larger than those of Robbins et al. (2014), indicating that the latter might be too optimistic in reporting uncertainties of the flux estimates.

Overall, the various observation- and model-derived estimates for GoMex regions agree in terms of their broad patterns, but the existing discrepancies and at times large uncertainties indicate that current estimates need further refinement.

Quantitative understanding of methane dynamics in the GoMex coastal and oceanic environments is limited. Solomon et al. (2009) argued that there is a potentially significant contribution to the atmosphere from the deep methane hydrate seeps in the Gulf of Mexico. They estimated ocean-atmosphere fluxes from seep plumes of 197±135 to 6,520±3,530 μmol m^{-2} d^{-1} compared to -0.38±0.34 to 7.0±1.4 μmol m^{-2} d^{-1} for background sites. Subsequent acoustic analyses of bubble plume characteristics questioned the finding that methane bubbles made their way to the surface (Weber et al., 2014), and the fate of methane emissions from seeps and their overall contribution to atmospheric methane remains uncertain.

16.3.4 North American Arctic

The North American Arctic (NAA) coastal ocean consists of broad (~300 km) shallow shelves in the Bering and Chukchi Seas, the narrower (<100 km) Beaufort Sea shelf, and the extensive Canadian Arctic Archipelago (CAA). Shelf waters in these regions originate in the North Pacific and follow a large-scale pathway from their entrance into the NAA through the Bering Strait via the Chukchi and Beaufort Seas

into the CAA and ultimately the North Atlantic (Carmack et al. 2015; Carmack et al. 2006). Most of the coastal region, with the exception of the southernmost Bering Sea, is covered with sea ice from approximately October to June. Areas of persistent multi-year sea ice cover currently exist at the northernmost extent of the CAA, and reoccurring polynyas are found in all three of the major regions (Smith and Barber 2007). The NAA is sparsely populated with communities that are heavily reliant on subsistence fishing and hunting; the rapid changes across the region associated with global warming are impacting these communities. Globally, the pace of increasing air temperatures is the highest in the NAA and adjacent Arctic regions, and this is resulting in significant reductions in both summer and winter sea ice cover with profound impacts on the marine ecosystems across the northern extent of the continent (Jefferies et al. 2015; Moore and Stabeno 2015, Steiner et al. 2015).

Coastal waters in the NAA have consistently been described as a net sink for atmospheric CO_2 (Bates et al. 2011, Bates et al. 2006, Chen et al. 2013, Cross et al. 2014, Dai et al. 2013, Evans et al. 2015, Laruelle et al. 2014, Mucci et al. 2010, Shadwick et al. 2011). This general trait is due to low surface water pCO_2 levels relative to the atmosphere set by the combination of cold water and seasonally high rates of both ice-associated and open water primary production (Cai et al. 2010b, 2014, Steiner et al. 2014), and limited gas exchange through sea ice relative to open water (Butterworth and Miller 2016; Rutgers van der Loeff et al. 2014) during winter months when under-ice pCO_2 levels are higher. However, sea ice is porous and its permeability is a known function of temperature (Golden et al. 2007); therefore, as Arctic winter temperatures continue to rise, the role of wintertime air-ice CO_2 exchange may become increasingly important. To date, measurements of wintertime exchange have been limited to very few studies (Else et al. 2011; Else et al. 2013; Miller et al. 2011). In recent years, the role of sea ice growth and decay has been shown to significantly affect the air-sea flux of CO_2 (Rysgaard et al. 2007; Rysgaard et al. 2009). During sea ice formation, brine rejection forms dense high-saline water that is exported from the surface layer. This process alters the ratio of total alkalinity (TA) to DIC of the sea ice and the underlying seawater, as DIC is a component of the brine whereas TA precipitates in the brine channels as a form of calcium carbonate known as ikaite (Dieckmann et al. 2008; Rysgaard et al. 2013). During sea ice decay, ikaite dissolves leading to excess TA relative to DIC and undersaturation of CO_2 in meltwater.

Estimates of air-sea CO_2 flux in the Chukchi and Beaufort Seas and western CAA all report atmospheric CO_2 uptake ranging from near neutral to 40 mmol m^{-2} d^{-1} (Bates, 2006; Gao et al., 2012; Semiletov et al., 2007; Else et al 2013; Mucci et al, 2010; Shadwick et al., 2011) with significantly higher uptake occurring over the broad and productive Chukchi shelf. A recent synthesis of a decade of coastal ocean data for this portion of the NAA collected within 400 km of land determined annual mean uptake of 2 mmol m^{-2} d^{-1} (Evans et al., 2015). Variability in the physical environment in terms of the wind patterns and sea ice cover affect the water column structure and connectivity between the surface ocean and overlaying atmosphere via gas exchange, which in turn affects the magnitude and sign of air-sea CO_2 exchange.

16.4 Climate Trends and Feedbacks

16.4.1 Trends in Coastal Carbon Fluxes

Important questions with respect to coastal carbon fluxes include: How will the coastal ocean change as a CO_2 sink? What is the anthropogenic component of the CO_2 sink? And how will the total and anthropogenic flux proportions be impacted by changing climate and other forcings? As stated already briefly in Section 16.2, when considering the ocean's role in sequestering anthropogenic carbon, the

anthropogenic flux component is relevant, not the total uptake flux. However, since neither the quantification of the anthropogenic carbon flux component nor the prediction of its future trend is straightforward, we discuss likely trends in total carbon fluxes with the implicit assumption that both will behave similarly.

A direct effect of increasing atmospheric CO_2 may be an increase in net uptake by the coastal ocean. In addition to rising atmospheric CO_2 levels, changes in climate forcings (i.e., surface heat fluxes, winds and freshwater input) affect carbon fluxes in coastal NA waters in several important ways. Ocean warming reduces the solubility of gases and thus directly affects gas concentrations near the surface; this will likely reduce the net air-sea flux of CO_2 by reducing the undersaturation of CO_2 (see Cahill et al. 2016 for the NAAC). Surface warming also strengthens vertical stratification and thus impedes vertical mixing, which will affect upward diffusion of nutrients and DIC. Enhanced stratification could thus lead to decreases in biologically driven carbon uptake and decreases in outgassing of CO_2. In coastal upwelling systems, surface warming will also increase the horizontal gradient between cold, freshly upwelled source waters and warm offshore surface water, leading to a greater tendency for the subduction of upwelled water at offshore surface temperature fronts during periods of persistent and strong upwelling-favorable winds. The cumulative effect of these processes for the NAPC may be greater and more persistent CO_2 outgassing near shore and lower productivity offshore as upwelled nitrate is exported before it may be used by the phytoplankton community (Evans et al., 2015). It is clear that rates of warming are faster in higher latitudes, but it is not easy to predict what the net effect of these warming-induced changes will be in the NAA. Furthermore, warming in the Arctic leads to reductions in ice cover and increases the duration of ice-free periods, both of which directly affect air-sea gas exchange (Bates and Mathis 2009). Another profound effect of Arctic warming is the melting of permafrost, which leads to the release of large quantities of CH_4 to the atmosphere, from both the land surface and the coastal ocean (Crabeck et al. 2014, Parmentier et al. 2013).

Changes in wind stress also directly affect air-sea gas fluxes because stronger winds intensify gas exchange. For the NAAC, changes in wind stress can lead to significant modifications of air-sea fluxes (Previdi et al. 2009, Cahill et al. 2016). Large-scale changes in wind patterns also affect circulation patterns with a range of implications. Upwelling-favorable winds along the NAPC have intensified over recent years, especially in the northern parts of the upwelling regimes (Rykaczewski et al. 2015, Garcia-Reyes et al. 2015, Sydeman et al. 2014, Rykaczewski and Checkley 2008), which has led to shoaling of subsurface nutrient-rich waters (Aksnes and Ohman, 2009), increased productivity (Kahru et al. 2015, Jacox et al. 201?, Chavez et al., 2011), higher DIC delivery to the surface (Turi et al. 2016), and declining oxygen levels (Peterson et al. 2013, Crawford et al. 2015?). In the NAA, late season air-sea CO_2 fluxes may become increasingly more directed toward the atmosphere as Arctic low pressure systems with storm force winds co-occur more often with periods of open water, ventilating CO_2 respired from the high organic carbon loading of the shallow shelf bottom waters (Evans et al. 2015, Hauri et al. 2013, Steiner et al. 2013) and affecting net annual exchanges. The intense warming observed across the NAA also influences mid-latitude weather patterns (Kim et al. 2014), with likely cascading effects on CO_2 exchanges through adjustments in the wind field.

16.4.2 Acidification Trends in North American Coastal Oceans

Increasing atmospheric CO_2 emissions and the consequent ocean uptake of CO_2 reduce the pH of seawater, its carbonate ion concentration and its carbonate saturation states, which can significantly affect

growth, metabolism and life cycles of marine organisms (e.g., Gattuso and Hanson 2011). Since about 1750, the ocean has absorbed approximately a quarter to a third of anthropogenic CO_2 emissions to the atmosphere from fossil fuel burning, cement production and land use changes (Sabine and Tanhua 2010). Due to this uptake of atmospheric CO_2, the surface ocean CO_2 partial pressure has increased while oceanic pH, carbonate ion concentration, and the carbonate saturation state have all decreased. This suite of chemical changes is commonly called "ocean acidification," which is defined more precisely as: "any reduction in the pH of the ocean over an extended period, typically decades or longer, which is caused primarily by uptake of CO_2 from the atmosphere, but can also be caused by other chemical additions or subtractions from the ocean" (IPCC, 2011, p. 37). DIC concentrations in natural waters vary due to biological production and respiration, which themselves can be altered by inputs of nutrients or organic matter from land. Ocean acidification most directly affects marine calcifiers, organisms that precipitate calcium carbonate to form internal or external body structures. Of the different forms of calcium carbonate precipitated by calcifying organisms, aragonite is the most soluble form.

While trends of acidification in open ocean waters tend to occur at a similar rate as the increase in atmospheric CO_2, acidification in coastal waters can be more sporadic and much faster. Polar regions are naturally prone to acidification because of their low temperatures (Orr et al. 2005). In Arctic coastal waters, pH and carbonate saturation state are naturally low relative to lower latitude coastal settings due to higher CO_2 solubility, the influence of multiple sources of freshwater (including riverine, glacial melt and sea ice melt) with varying CO_2 chemistries, and the high respiratory DIC content in bottom waters (the Beaufort and Chukchi Sea continental shelves experience inflows of naturally corrosive Pacific seawater with pH as low as 7.6; Mathis et al. 2011). The main contributing factor to the relative high rates of acidification in polar waters is retreating sea ice, which adds melt water from multi-year ice and by increasing the surface area of open water, thereby enhancing the uptake of atmospheric CO_2 (Steiner et al. 2013a, Cai et al. 2010b). These factors, in combination with increasing atmospheric CO_2 levels, have set a faster pace of ocean acidification in the Arctic than projected trends in other coastal regions (Fabry et al. 2009, Feely et al. 2009, Mathis et al. 2015). Model projections for the Bering and Chukchi Seas predict annual average aragonite undersaturation (i.e., favoring dissolution) by the year 2070 and 2030, respectively (Mathis et al. 2015). The Beaufort Sea upper halocline and deep waters now regularly show aragonite undersaturation (Mathis et al. 2015, Miller et al 2014). These chemical seawater signatures are propagated via M'Clure Strait and Amundsen Gulf into the CAA and beyond (AMAP, 2013, Azetsu-Scott et al. 2010; Turk et al. 2016; Yamamoto-Kawai et al. 2013). Model projections based on the Intergovernmental Panel on Climate Change high-CO_2 emissions scenario, Representative Concentration Pathway 8.5 (RCP8.5), suggest the surface Beaufort Sea will become undersaturated with respect to aragonite around 2025 (Steiner et al. 2014). As these conditions intensify, it is expected that the negative impacts for calcifying marine organisms will become a critical issue reshaping ecosystems and fisheries across the NAA domain (Mathis et al. 2014, Moore and Stabeno 2015).

In the NAPC region, climate-driven changes in upwelling circulation result in coastal acidification events. As mentioned in section 16.4.1, upwelling-favorable winds along the NAPC have intensified over recent years, especially in the northern parts of the upwelling regimes (Rykaczewski et al., 2015; Garcia-Reyes et al. 2015; Sydeman et al., 2014; Rykaczewski & Checkley, 2008). Intensified upwelling supplies deep water to the shelf that is rich in DIC and nutrients but poor in oxygen. Ocean acidification and hypoxia are thus strongly linked ecosystem stressors as low-oxygen, high-CO_2 conditions both derive from the microbial respiration of organic matter (Chan et al. 2016). In the northern CCS, pCO_2, pH and aragonite

saturation reach levels known to be harmful to ecologically and economically important species during the summer upwelling season (Chapter 17, Feely et al. 2008; Harris et al. 2013; Bednarsek et al. 2014). During four upwelling season cruises, northern CCS conditions over the continental shelf saw pCO_2 levels above 1000 µatm roughly 25% of the time and aragonite below saturation up to 50% of the time (Feely et al. 2016). Observations on an Oregon coast mooring show regular excursions of pCO_2 up to 1200 µatm at the surface during upwelling events (Evans et al., 2011, Harris et al. 2013). Along the CAI at the southern end of the NAPC region, mountain wind jets can drive upwelling of waters with pCO_2 of ~1000 µatm and aragonite saturation states near equilibrium (Chapa-Balcorta et al. 2015). In the GAK, where surface pCO_2 is oversaturated with respect to the atmosphere during winter months, aragonite saturation is near equilibrium in winter (Evans and Mathis, 2013).

In the northern GoMex, where surface aragonite saturation states typically range from 3.6 to 4.5 and are thus well above the dissolution threshold (Wang et al. 2013, Wanninkhof et al., 2015), excessive nutrient inputs from the Mississippi River result in eutrophication-induced acidification of bottom waters (Cai et al. 2011, Laurent et al. in revision). Similar to the CCS, low-oxygen and high-CO_2 conditions coincide and derive from the microbial respiration of organic matter (Cai et al. 2011, Laurent et al, in revision). Currently aragonite saturation states are around 2 in hypoxic bottom waters and thus well above equilibrium. Projections suggest that aragonite saturation states will drop below saturation toward the end of this century (Cai et al., 2011, Laurent et al. in prep.).

Recent studies indicate that the northern regions of the NAAC (the MAB and GoM) are more prone to acidification than the SAB (Wang et al. 2013, Wanninkhof et al. 2015). Coastal waters in this region have, on average, lower pH, lower aragonite saturation states and higher CO_2 fugacity than more southern coastal regions, which is primarily driven by a decreasing trend in mean total alkalinity of shelf water from the SAB northward to the Gulf of Maine. The GOM, which has a significant shellfish industry, displays the lowest pH and aragonite saturation levels along the east coast in summer (Wang et al. 2013).

16.5 Conclusions

There has been tremendous progress in improving understanding and constraining rates of carbon cycling in coastal waters since the last SOCCR report, primarily because of a greatly expanded suite of observations, process studies and models. However, quantification of many coastal carbon fluxes remains a significant challenge. Carbon is constantly exchanged across the air-sea interface as well as interfaces between land and coastal ocean, between coastal and open ocean waters, and between water and sediment. Net exchange fluxes and trends are relatively small signals masked by a large fluctuating background. At present, most of these fluxes are not constrained well enough to derive complete carbon budgets for NA coastal waters, or to project how fluxes will change in the future due to various drivers.

This chapter focused primarily on the role of ocean margins in sequestering atmospheric CO_2 and coastal ocean acidification. In the coastal ocean a net removal of carbon from direct interaction with the atmospheric reservoir can occur by export of dissolved or particulate carbon to the deep ocean or by permanent burial in sediments. Neither of these is easily observed or well quantified at present. The best-observed flux is gas exchange across the air-sea interface although extracting the small net flux and its trend from a variable background remains a challenge. Ultimately, the removal of anthropogenic carbon is the relevant quantity for assessing the contribution of ocean margins to the uptake of anthropogenic

carbon; however, the separation of anthropogenic fluxes from the natural background is elusive for coastal waters thus far.

At present, estimates of air-sea CO_2 fluxes provide the best evidence for the contribution of coastal waters to overall carbon uptake by the ocean. In the broad shelf system of the NAAC, shelf water is separated from the adjacent open ocean by persistent shelf break currents and density fronts. Available estimates suggest that the NAAC overall is a weak sink with some subregions acting as sources (Scotian Shelf and near-shore regions of the SAB), some being neutral (GOM) and some acting as weak sinks (MAB and outer SAB). Large sections of the narrow shelf of the NAPC are dominated by upwelling circulation, which leads to strong outgassing of CO_2 near the coast; however, outgassing is compensated for by biologically driven uptake from upwelled nutrients further offshore. Recent estimates are consistent in suggesting that the region is a weak to moderate sink of atmospheric CO_2. The relatively wide shelves in the Gulf of Mexico are considered a weak net sink, with the WFS and the western shelf acting as sources, the Mexico shelf being neutral and only the northern shelf a clear sink, which is in large part driven by anthropogenic nutrient inputs from the Mississippi River. The wide, seasonally ice-covered shelves in the NAA are consistently acting as sink for atmospheric CO_2. The low surface water pCO_2 in this region is primarily due to water temperatures and the fact that seasonal ice cover hinders uptake of atmospheric CO_2 during a signification fraction of the year. Overall NA coastal waters act as sink, but regional variations and uncertainties are large.

There are several drivers that are influencing secular trends in coastal carbon fluxes and will continue to do so in the future including rising atmospheric CO_2 levels, changes in atmosphere-ocean interactions (e.g., wind forcing and heat fluxes), changes in the hydrological cycle and anthropogenic perturbations of global nutrient cycling (in particular the nitrogen cycle). It is clear that coastal surface pCO_2 does not closely track atmospheric pCO_2. We have identified a number of plausible mechanisms for how coastal carbon uptake may change in the future, but cannot predict the total effect with any confidence. Regional model studies are beginning to address these.

A major concern is coastal acidification, which can affect the growth, metabolism and life cycles of many marine organisms, specifically calcifiers, and can trigger cascading ecosystem-scale effects. Most vulnerable are those organisms that precipitate aragonite, the most soluble form of biogenic $CaCO_3$ in the ocean. Aragonite saturation states are routinely below saturation (i.e., favoring dissolution) in NAA coastal waters. In the NAPC region, intensified upwelling brings low-pH, low-oxygen water onto the shelves and aragonite saturation levels drop below saturation during the summer upwelling season. In the northern GoMex aragonite saturation states are well above the dissolution threshold. Although eutrophication-induced acidification occurs in bottom waters influenced by Mississippi River inputs of nutrients and freshwater, saturation levels remain well above the dissolution threshold.

Given the importance of coastal margins, both in contributing to carbon budgets and in the societal benefits they provide, further efforts to improve our assessments of the carbon cycle in these regions is paramount. Maintaining and expanding existing coastal observing programs, continued national and international coordination and integration of observations, modeling capabilities and stakeholder needs are critical.

Supporting Evidence

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Figures