16 Coastal Ocean and Continental Shelves

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KEY FINDINGS

1. Observing networks and high-resolution models are now available to construct coastal carbon budgets. Efforts have focused primarily on quantifying the net air-sea exchange of carbon dioxide (CO_2), but some studies have estimated other key fluxes, such as the exchange between shelves and the open ocean.

2. Available estimates of air-sea carbon fluxes, based on more than a decade of observations, indicate that the North American margins act as a net sink for atmospheric CO_2. This net uptake is driven primarily by fluxes in the high-latitude regions. The estimated magnitude of the net flux is 160 ± 80 teragrams of carbon per year (*medium confidence*) for the North American Exclusive Economic Zone, a number that is not well constrained.

3. The increasing concentration of CO_2 in coastal and open-ocean waters leads to ocean acidification. Corrosive conditions in the subsurface occur regularly in Arctic coastal waters, which are naturally prone to low pH, and North Pacific coastal waters, where upwelling of deep, carbon-rich waters has intensified and, in combination with the uptake of anthropogenic carbon, leads to low seawater pH and aragonite saturation states in spring, summer, and early fall (*very high confidence, very likely*).

4. Expanded monitoring, more complete syntheses of available observations, and extension of existing model capabilities are required to provide more reliable coastal carbon budgets, projections of future states of the coastal ocean, and quantification of anthropogenic carbon contributions.

Note: Confidence levels are provided as appropriate for quantitative, but not qualitative, Key Findings and statements.

16.1 Introduction

Along ocean margins, the atmospheric, terrestrial, sedimentary, and deep-ocean carbon reservoirs meet, resulting in quantitatively significant carbon exchanges. Anthropogenic activities lead to secular trends in these exchanges. The drivers underlying these trends include rising atmospheric carbon dioxide (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 quantify (Regnier et al., 2013).

This chapter focuses on two particularly pressing issues within the much broader topic of carbon cycling along ocean margins: 1) the uptake of atmospheric CO_2 and subsequent export to the deep ocean and 2) patterns and drivers of coastal ocean acidification. The first is relevant to overall quantification of the ocean's uptake of CO_2. The second is directly relevant to coastal ecosystem health, fisheries, and aquaculture.

Two different terms will be used here when referring to ocean margins: 1) the coastal ocean, defined in this report as nonestuarine 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 in gray in Figure 16.1, p. 651). Although the two definitions overlap, there are important reasons for considering both. Along passive margins with broad shelves like the North American Atlantic Coast, the continental shelf is the relevant spatial unit for discussing carbon fluxes. Along active margins with narrow shelves, such as the North American Pacific Coast, a larger region than just the shelf needs to be considered to meaningfully discuss coastal carbon dynamics. The 370-km limit chosen here to define the coastal ocean was recommended by Hales et al. (2008) and corresponds to the Exclusive

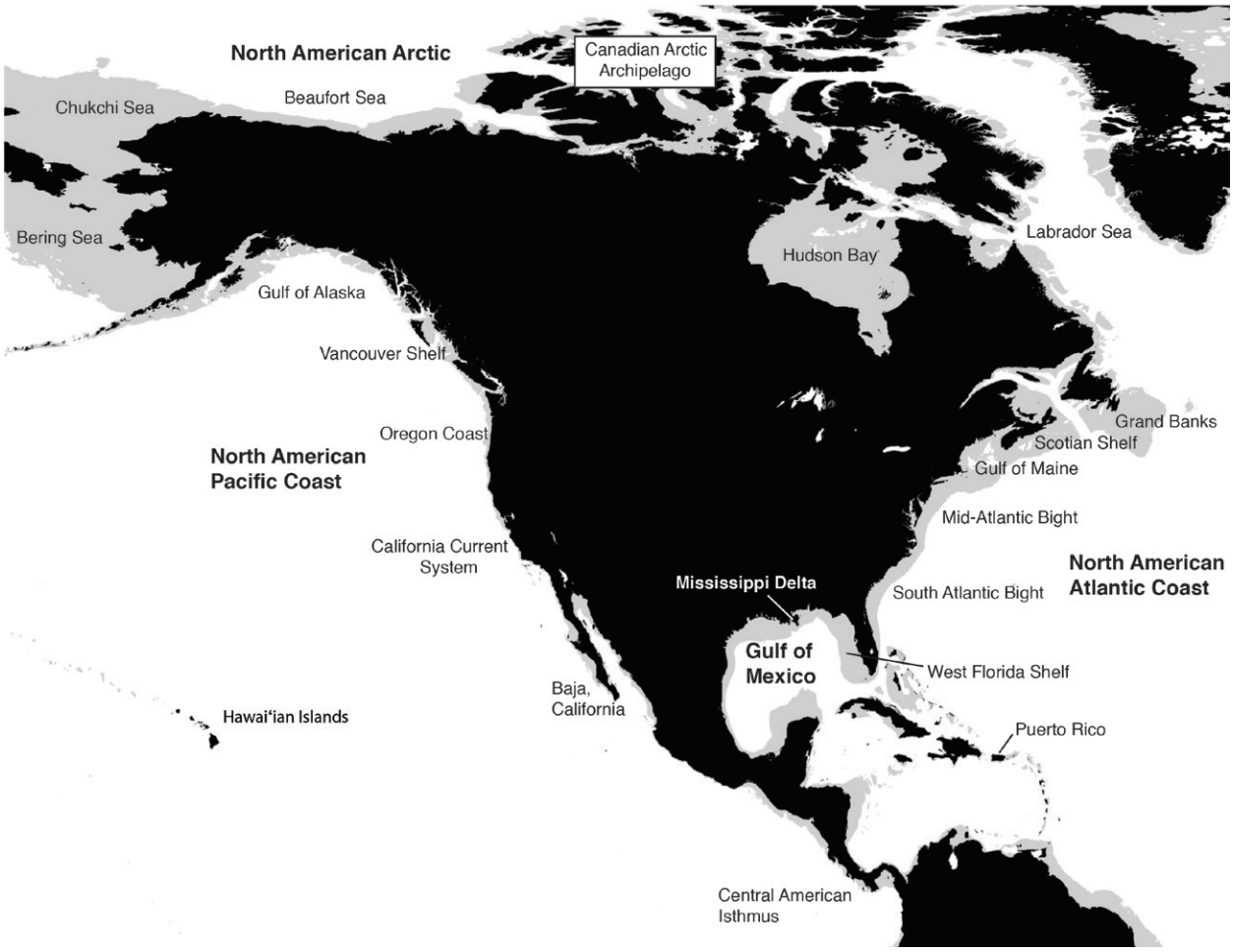

Figure 16.1. North American Shelf Seas. These seas (in gray) are defined as waters with bottom depths less than 200 m.

Economic Zone (EEZ, the region where a nation can claim exclusive rights for fishing, drilling, and other economic activities). Worth noting here is that ocean CO_2 uptake or loss is not credited to any nation under Intergovernmental Panel on Climate Change (IPCC) CO_2 accounting; instead, ocean uptake is viewed as an internationally shared public commons.

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 State of the Carbon Cycle Report* (SOCCR1; CCSP 2007; Chavez et al., 2007), and activities within the North American coastal interim synthesis (Benway et al., 2016; Alin et al., 2012; Najjar et al., 2012; Mathis and Bates 2010; Robbins et al., 2009). SOCCR1 (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.

The objective here 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. Summarized first are the key

variables and fluxes relevant to carbon budgets for coastal waters, followed by descriptions of 1) the mechanisms by which carbon can be removed from the atmospheric reservoir and 2) the means for quantifying the resulting carbon removal (see Section 16.2, this page). Next presented is available research relevant to carbon budgets for North American coastal waters by region, along with an assessment of whether enough information is available to derive robust estimates of carbon export to the open ocean (see Section 16.3, p. 655). Climate-driven trends in coastal carbon fluxes and coastal ocean acidification are then discussed (see Section 16.4, p. 669), followed by conclusions (see Section 16.5, p. 673).

16.2 Current Understanding of Carbon Fluxes and Stocks

Carbon is present in various inorganic and organic forms in coastal waters (see Figure 16.2, p. 653). Dissolved inorganic species include aqueous CO_2 (a combination of dissolved CO_2 and carbonic acid), bicarbonate and carbonate ions, and methane (CH_4); the first three carbon species are collectively referred to as dissolved inorganic carbon or DIC. 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 (Liu et al., 2010), and DOC constitutes more than 90% to 95% of this TOC (Vlahos et al., 2002).

Carbon is constantly transferred among 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 (see Figure 16.2, p. 653). 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 $CaCO_3$. Major internal transformations are the conversion of DIC into organic carbon (POC and DOC), through primary production, and respiration throughout the water column, returning most 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, CH_4 (i.e., in the relative absence of electron acceptors other than CO_2). Both DIC and CH_4 are released back into the overlying water. POC that is not respired (referred to as refractory POC) can be buried in sediments and stored for a very long time. Some organisms form internal or external body structures of $CaCO_3$, which either dissolve or become incorporated into the sediments and are buried. This discussion will refer to long-term storage of buried 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) seawater warming and cooling, which affects CO_2 solubility; 2) the under- or oversaturation of CO_2 resulting 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, sea ice extent, and surface films). 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 (Bauer et al., 2013; 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 organic and inorganic carbon inputs from land via rivers and estuaries (see Ch. 15: Tidal Wetlands and Estuaries, p. 596), from tidal wetlands, and exchanges between the coastal and open oceans 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

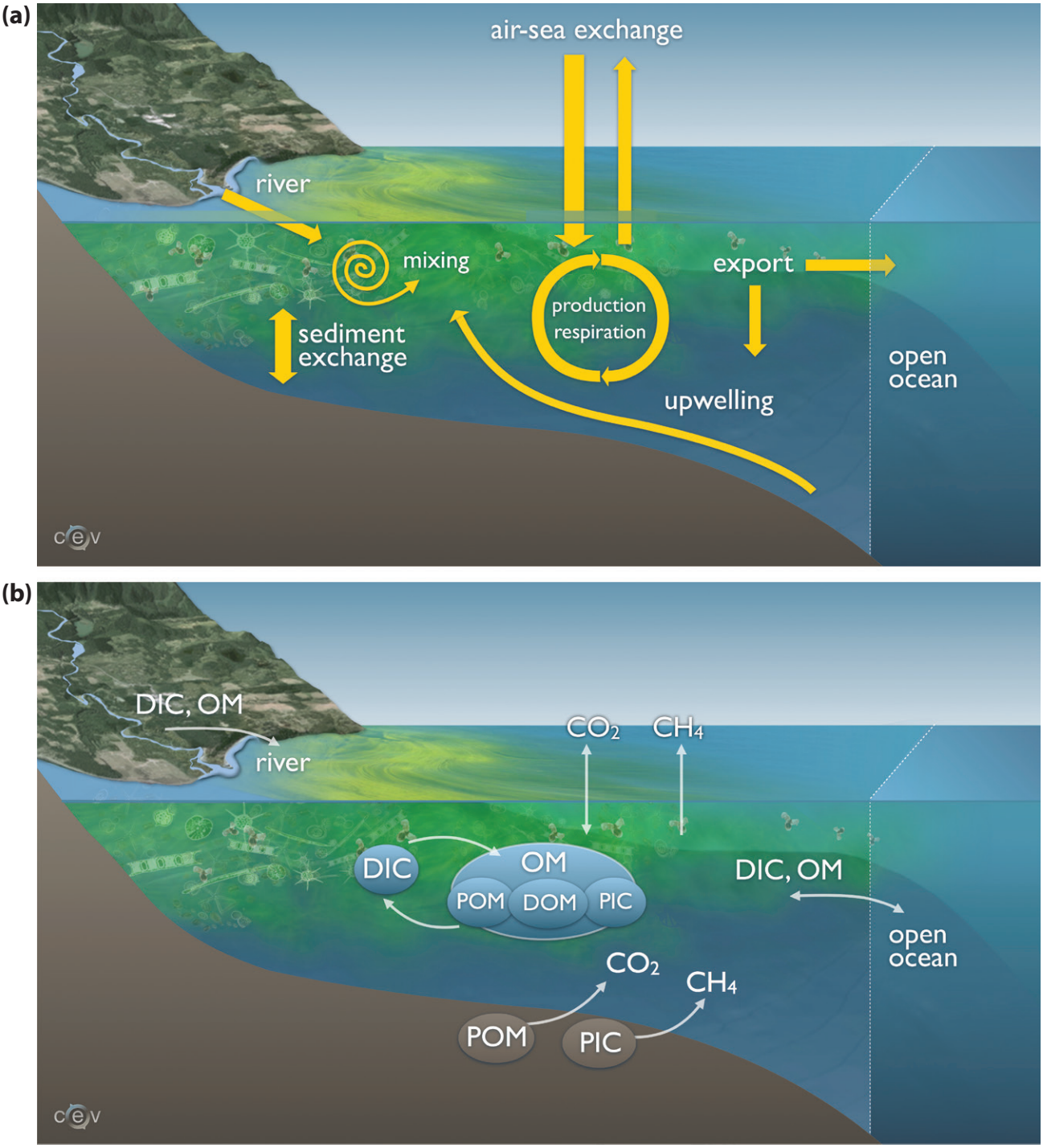

Figure 16.2. Major Coastal Carbon Pools and Fluxes. (a) Carbon in various forms (e.g., CO_2, carbon dioxide; CH_4, methane) is transferred among different pools and exchanged across interfaces between land, air, and ocean in coastal regions. **(b)** Carbon forms include dissolved inorganic carbon (DIC), organic matter (OM), particulate organic matter (POM), dissolved organic matter (DOM), and particulate inorganic matter (PIC). [Figure sources: Simone Alin, National Oceanic and Atmospheric Administration; Hunter Hadaway, University of Washington Center for Environmental Visualization; and Katja Fennel, Dalhousie University.]

reservoir can occur by export to the deep ocean or by permanent burial in coastal sediments.

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

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 winter, shelf water is cooled more strongly than surface water in the adjacent open ocean because the former is not subject to deep convection. The colder shelf water is denser and experiences a larger influx of atmospheric CO_2; both density and the solubility of CO_2 increase with decreasing temperature. 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 potentially could 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 primarily would be advected laterally. Both mechanisms for carbon export critically depend on physical transport of carbon-rich water off the shelf.

Carbon export flux from coastal waters to the deep ocean cannot be quantified easily or accurately through direct observation. Thus, the only available estimates of such export are indirect, using mass balances of POC and dissolved oxygen (Hales et al., 2006), mass balances of DOC (Barrón and Duarte 2015; Vlahos et al., 2002), mass balances of TOC and DIC (Najjar et al., 2018), and model estimates (Izett and Fennel 2018a, 2018b; Bourgeois et al., 2016; Fennel and Wilkin 2009; Fiechter et al., 2014; Mannino et al., 2016; Turi et al., 2014; Xue et al., 2013). If the total carbon inventory in a coastal system can be considered constant over a sufficiently long timescale (i.e., on the order of years), inferring carbon export is possible from using the sum of all other exchange fluxes across the system's interfaces over that same period. Export to the open ocean must balance the influx of carbon from land and wetlands, its net exchange across the air-sea interface, lateral exchange caused by advection, and any removal through permanent sediment burial. The accuracy of the inferred export flux directly depends on the accuracy of the other flux estimates and of the assumption of a constant carbon inventory. Quantifying internal transformation processes (e.g., respiration and primary and secondary production) does not directly enter this budgeting approach but can elucidate the processes that drive fluxes across interfaces.

Current estimates of carbon fluxes across coastal interfaces come with significant uncertainties (Regnier et al., 2013; Birdsey et al., 2009). These uncertainties are caused by a combination of 1) small-scale temporal and spatial variability, which is undersampled by currently available means of direct observation, and 2) regional heterogeneity, which makes scaling up observations from one region to larger areas difficult. Contributing to variability in regional carbon budgets and export are geographical differences arising from 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 of hydrography due to the rotation of the Earth (Sharples et al., 2017). Section 16.3, p. 655, describes the regional characteristics of North American

coastal waters and how these characteristics influence carbon dynamics. Available estimates of carbon fluxes are compiled in an attempt to estimate export.

The motivation for quantifying permanent burial of carbon and export of carbon from coastal waters to the deep ocean is that both processes remove CO_2 from the atmospheric reservoir. A more relevant but harder to obtain quantity in this context is the burial or export of anthropogenic carbon. The anthropogenic component of a given carbon flux is defined as the difference between its preindustrial and present-day fluxes. Thus, present-day carbon fluxes represent a superposition of the anthropogenic flux component and the 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 is difficult to assess for coastal ocean fluxes and has to rely on process-based arguments and models (Regnier et al., 2013). Observation-based estimates of the global open ocean's anthropogenic uptake have been made by Sabine et al. (2004), Sabine and Tanhua (2010), and Carter et al. (2017). Bourgeois et al. (2016) were the first to estimate coastal anthropogenic carbon uptake in their global model. Their estimates are presented in some detail in Section 16.3.5, p. 665.

16.3 Coastal Carbon Fluxes Around North America

16.3.1 North American Atlantic Coast

The North American Atlantic Coast borders on a wide, geologically passive margin shelf that extends from the southern tip of Florida to the continental shelf of the Labrador Sea (see Figure 16.1, p. 651). The shelf is several hundreds of kilometers wide in the north (Labrador shelf and Grand Banks) but narrows progressively toward the south in the Middle Atlantic Bight (MAB), which is between Cape Cod and Cape Hatteras, and the South Atlantic Bight (SAB), which is south of Cape Hatteras. The SAB shelf width measures only several tens of kilometers. Two major semi-enclosed bodies of water are the Gulf of Maine (GOM) and 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). Because 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 North American Atlantic Coast and significant seasonal and interannual variability. The net flux on the Scotian Shelf remains controversial. Shadwick et al. (2010), combining *in situ* and satellite observations, reported a large source of CO_2 to the atmosphere of 8.3 ± 6.6 grams of carbon (g C) per m^2 per year. In contrast, Signorini et al. (2013) estimated a relatively large sink of atmospheric CO_2, 14 ± 3.2 g C per m^2 per year, when using *in situ* data alone and a much smaller uptake, 5.0 ± 4.3 g C per m^2 per year, from a combination of *in situ* and satellite observations. The open GOM is a weak net source of 4.6 ± 3.1 g C per m^2 per year according to Vandemark et al. (2011) but with significant interannual variability, while Signorini et al. (2013) estimate the region to be neutral. The shallow, tidally mixed GOM regions (i.e., Georges Bank and Nantucket Shoals) are thought to be sinks, however (see Table 16.1, p. 657; Signorini et al., 2013). The MAB and SAB are net sinks. Observation-based

estimates for the MAB include sinks of 13 ± 8.3 g C per m^2 per year (DeGrandpre et al., 2002) and 13 ± 3.2 g C per m^2 per year (Signorini et al., 2013). Estimates for the SAB include sinks of 5.8 ± 2.5 g C per m^2 per year (Jiang et al., 2008) and 8.2 ± 2.9 g C per m^2 per year (Signorini et al., 2013). The change from neutral or occasional net source in the Scotian Shelf and GOM regions to net sink in the MAB arises because the properties of shelf water are modified during its southwestward flow by air-sea exchange, inflows of riverine and estuarine waters (Salisbury et al., 2008b, 2009), and exchange with the open North Atlantic across the shelf break (Cai et al., 2010a; 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. 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.

Shelf water in the SAB is distinct from that in the MAB and has almost no trace of Labrador Current water; instead, its characteristics are similar to those of the Gulf Stream, but its carbon signature is modified by significant organic and inorganic carbon and alkalinity inputs from coastal marshes (Cai et al., 2003; Jiang et al., 2013; Wang and Cai 2004; Wang et al., 2005). Herrmann et al. (2015) estimated that 59% of the 3.4 teragrams of carbon (Tg C) per year of organic carbon exported from U.S. East Coast estuaries is from the SAB. The subsequent respiration of this organic matter and direct outgassing of marsh-derived carbon make the nearshore 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 (DeGrandpre et al., 2002; Shadwick et al., 2010, 2011; Signorini et al., 2013; Vandemark et al., 2011; Wang et al., 2013). Hydrography and CO_2 dynamics on the Scotian Shelf are influenced by the significant freshwater 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., 2008a, 2009). Riverine and estuarine inputs become more important in the MAB with discharges from the 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 that further north (Wang et al., 2013), although sporadic phytoplankton blooms do occur because of 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 (Cai and Wang 1998; Wang and Cai 2004; Wang et al., 2005).

Regional biogeochemical models reproduce the large-scale patterns of air-sea CO_2 flux with oceanic uptake increasing from the SAB to the GOM (Cahill et al., 2016; Fennel et al., 2008; Previdi et al., 2009). 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 with that in a low-NAO year. In the MAB, the decrease resulted primarily from 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 future, climate-driven warming and trends in atmospheric forcing (primarily wind) on air-sea CO_2 flux (without considering the

Table 16.1. Regional Estimates of Net Air-Sea Carbon Dioxide Exchange from Observations and Regional Models[a,b]

Region	Area (km²)	Air-Sea Exchange		Observation-Based Estimate or Model	Reference
		g C per m² per year[a,b]	Tg C per year[a,b]		
North American Atlantic Coast (NAAC)					
Scotian Shelf	2.2×10^5	8.3 ± 6.6	1.8	Combination of *in situ* and satellite observations (10-year average, 1999–2008)	Shadwick et al. (2010)
	1.28×10^5	−14 ± 3.2	−1.9	Observation-based estimate (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
		−5.0 ± 4.3	−0.64	Combination of *in situ* and satellite observations (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
	1.2×10^5	−28 ± 0.72	−3.3	Model (2-year average, 2004–2005)	Fennel and Wilkin (2009)
Gulf of Maine (without Georges Bank and Nantucket Shoals)	1.28×10^5	0.48 ± 2.6	0.061	Observation-based estimate (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
		0.12 ± 0.96	0.015	Combination of *in situ* and satellite observations (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
		4.6 ± 3.1	0.58	Observation-based estimate (5-year mean, 2004–2008)	Vandemark et al. (2011)
Georges Bank and Nantucket Shoals	0.58×10^5	−8.5 ± 2.6	−0.49	Observation-based estimate (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
		−16 ± 2.9	−0.95	Combination of *in situ* and satellite observations (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
Gulf of Maine (with Georges Bank and Nantucket Shoals)	1.7×10^5	−20 ± 4.9	−3.4	Model (2-year average, 2004–2005)	Fennel and Wilkin (2009)
	0.87×10^5	−27 ± 8.4	−1.9	Model (4-year average, 2004–2007)	Cahill et al. (2016)

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Table 16.1. Regional Estimates of Net Air-Sea Carbon Dioxide Exchange from Observations and Regional Models[a,b]

Region	Area (km²)	Air-Sea Exchange		Observation-Based Estimate or Model	Reference
		g C per m² per year[a,b]	Tg C per year[a,b]		
Mid-Atlantic Bight (MAB)	1.25×10^5	−13 ± 8.3	−1.6	Observation-based estimate	DeGrandpre et al. (2002)
		−14	−1.8	Model (2004)	Fennel et al. (2008)
	0.93×10^5	−13 ± 3.2	−1.2	Observation-based estimate (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
		−21 ± 2.3	−2.0	Combination of *in situ* and satellite observations (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
	0.86×10^5	−11 ± 2.6	−0.92	Model (2-year average, 2004–2005)	Fennel and Wilkin (2009)
	1.15×10^5	−14 ± 2.4	−1.7	Model (4-year average, 2004-2007)	Cahill et al. (2016)
South Atlantic Bight (SAB)	1.02×10^5	−5.8 ± 2.5	−0.59	Observation-based estimate	Jiang et al. (2008)
		−8.2 ± 2.9	−0.83	Observation-based estimate (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
		−8.0 ± 1.9	−0.82	Combination of *in situ* and satellite observations (reference year, 2004)	Signorini et al. (2013); using Ho et al. (2011) gas transfer param.
	0.92×10^5	−6 ± 2.4	−0.55	Model (4-year average, 2004–2007)	Cahill et al. (2016)
Gulf of Mexico (GMx)					
Whole Gulf of Mexico	15.6×10^5	−2.3 ± 0.96	−3.6	Observation-based estimate	Robbins et al. (2014)
		−8.5 ± 6.5	−13	Model (7-year average, 2005–2010)	Xue et al. (2016)
Open Gulf of Mexico	10.1×10^5	−5.8 ± 0.84	−5.8	Observation-based estimate	Robbins et al. (2014)
		−12 ± 5.5	−13	Model (7-year average, 2005–2010)	Xue et al. (2016)
West Florida Shelf	1.5×10^5	4.4 ± 1.3	0.67	Observation-based estimate	Robbins et al. (2014)
		4.6 ± 0.58	0.68	Model (7-year average, 2005–2010)	Xue et al. (2016)

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Table 16.1. Regional Estimates of Net Air-Sea Carbon Dioxide Exchange from Observations and Regional Models[a,b]

Region	Area (km²)	Air-Sea Exchange		Observation-Based Estimate or Model	Reference
		g C per m² per year[a,b]	Tg C per year[a,b]		
Northern Gulf of Mexico	1.5×10^5	−5.3 ± 4.4	−0.79	Observation-based estimate	Robbins et al. (2014)
		−3.8 ± 8.9	−0.58	Model (7-year average, 2005–2010)	Xue et al. (2016)
	unknown	−11 ± 44		Observation-based estimate	Huang et al. (2015)
	unknown	−13 ± 3.6		Combination of *in situ* and satellite observations	Lohrenz et al. (2018)
Western Gulf of Mexico	0.8×10^5	2.2 ± 0.6	0.17	Observation-based estimate	Robbins et al. (2014)
		4.1 ± 3.8	0.33	Model (7-year average, 2005–2010)	Xue et al. (2016)
Mexico Shelf	1.8×10^5	−1.1 ± 0.6	−0.19	Observation-based estimate	Robbins et al. (2014)
		−2.3 ± 4.2	−0.41	Model (7-year average, 2005–2010)	Xue et al. (2016)
North America Pacific Coast (NAPC)					
Gulf of Alaska	3×10^6	−11	−36	Observations, climatology of 1991–2011, 0 to 400 km offshore	Evans and Mathis (2013)
British Columbia coastal ocean		−35		Observations, 1995–2001	Evans et al. (2012)
British Columbia Vancouver Island shelf		−6		Model, annual average	Ianson and Allen (2002)
Oregon Shelf		−3.6 ± 82		Observations inshore of 200-m isobath	Evans et al. (2011)
Oregon Shelf		−88		Observations	Hales et al. (2005)
50° to 22°N	1.76×10^6	−7.9	−14	Satellite-based prediction of pCO_2 and satellite-based wind speed, within 370 km of coast	Hales et al. (2012)
35° to 40°N			0.6	Model, 0 to 100 km from coast, 1999–2005	Fiechter et al. (2014)
40° to 45°N			−0.4	Model, 0 to 100 km from the coast, 1999–2005	Fiechter et al. (2014)

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(Continued)

Table 16.1. Regional Estimates of Net Air-Sea Carbon Dioxide Exchange from Observations and Regional Models[a,b]

Region	Area (km^2)	Air-Sea Exchange		Observation-Based Estimate or Model	Reference
		g C per m^2 per year[a,b]	Tg C per year[a,b]		
30° to 46°N	1.49×10^6	0.6 ± 2.4	0.9±3.6	Model, 0 to 800 km from the coast, 12-year simulation with climatological forcing	Turi et al. (2014)
North American Arctic (NAA)					
Chukchi Sea	2.9×10^5	−15	−4.4	Observations	Evans et al. (2015b)
	5.95×10^5	−175 ± 44	−38 ± 7	Observations	Bates et al. (2006)
	5.95×10^5	−35	−12.1	Observations	Gao et al. (2012)
		−17 ± 17		Satellite-based prediction of pCO_2 and satellite-based wind speed	Yasunaka et al. (2016)
Beaufort Sea (Amundsen Gulf)		−14		Observations	Shadwick et al. (2011)
Beaufort Sea (Cape Bathurst Polynya)		−44 ± 28		Observations	Else et al. (2013)
Beaufort Sea	9.2×10^5	−4.4	−4.0	Observations	Evans et al. (2015b)
Beaufort Sea		−10 ± 15		Observations	Mucci et al. (2010)
Western Arctic Coastal Ocean	1.2×10^6	−8.8 ± 4.8	−11 ± 5.7	Observations	Evans et al. (2015b)
Hudson Bay	7.32×10^5	−3.2 ± 1.8	−0.58 ± 0.3	Observations	Else et al. (2008)
Bering Sea	6.94×10^5	−9.6	−6.7	Observations	Cross et al. (2014a)
		−5.3	−3.7	Observations	Takahashi et al. (2009)

Notes

a) Positive fluxes indicate a source to the atmosphere.

b) C, carbon; CO_2, carbon dioxide; Tg, teragrams; g, grams; 1 Tg = 10^{12} g.

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 with that in 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) showed that the effective alkalinity flux resulting from denitrification in sediments of the North American Atlantic Coast reduces the simulated ocean uptake of CO_2 by 6% compared to a simulation without sediment denitrification.

The passive-margin sediments along the Atlantic coast have not been considered an area of significant CH_4 release until recently (Brothers et al., 2013; Phrampus and Hornbach 2012; Skarke et al., 2014). Phrampus and Hornbach (2012) predicted that

massive seepage of CH_4 from upper-slope sediments is occurring in response to warming of intermediate-depth Gulf Stream waters. Brothers et al. (2013) and Skarke et al. (2014) documented widespread CH_4 plumes in the water column and attributed them to gas hydrate degradation. Estimated CH_4 efflux from the sediment in this region ranges from 1.5×10^{-5} to 1.8×10^{-4} Tg CH_4 per year, where the uncertainty range reflects different assumptions underlying the conversion from CH_4 plume observations to seepage rates. The fraction of the released CH_4 that escapes to the atmosphere remains uncertain (Phrampus and Hornbach 2012).

16.3.2 North American Pacific Coast

The North American Pacific Coast extends from Panama to the Gulf of Alaska and is an active margin with varying shelf widths (see Figure 16.1, p. 651). The continental shelf is narrow along the coasts of California, Oregon, and Washington, with a width on the order of 10 km but widening significantly in the Gulf of Alaska, where shelves extend up to 200 km offshore. In the Gulf of Alaska, freshwater and tidal influences strongly affect cross-shelf exchange, and the shelf is dominated by downwelling circulation. The region from Vancouver Island to Baja California is a classic eastern boundary current upwelling region influenced by the California Current System (Chavez et al., 2017). Winds drive a coastal upwelling circulation characterized by equatorward flow in the California Current and by coastal jets and their associated eddies and fronts that extend offshore, particularly off the coasts of Baja California, California, Washington, and Oregon. The northern California Current System experiences strong freshwater influences and seasonality in wind forcing that diminish in the southern part of the system. 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 runs from Panama to the southern tip of Baja California and experiences intense and persistent wind events, large eddies, and high waves that combine to produce upwelling and strong nearshore mixing (Chapa-Balcorta et al., 2015; Franco et al., 2014). 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 (Chapa-Balcorta et al., 2015; Chelton et al., 2000a, 2000b; Gaxiola-Castro and Muller-Karger 1998; Lluch-Cota et al., 1997). The California Current brings water from the North Pacific southward into the southern California and Central American Isthmus regions, while the California Undercurrent transports equatorial waters northward in the subsurface.

The net exchange of CO_2 with the atmosphere across the North American Pacific Coast is characterized by strong spatial and temporal variation and reflects complex interactions between biological uptake of nutrients and degassing of nutrient- and carbon-rich upwelled waters. A growing number of coastal air-sea flux studies have used extrapolation techniques to estimate fluxes across the coastal ocean on regional to continental scales. Observation-based studies of air-sea CO_2 flux suggest that estimates for the coastal ocean from Baja California to the Gulf of Alaska range 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 because of their large and counter-balancing periods of efflux during upwelling conditions and influx during periods of relaxation and high primary productivity; this pattern is strongly modulated by El Niño–La Niña conditions (Friederich et al., 2002). Hales et al. (2005) used seasonal data to estimate an uptake of 88 g C per m^2 per year by Oregon coastal waters, which is about 15 times larger than the global mean of 6 g C per m^2 per year. Using data with greater temporal coverage, Evans et al. (2011) showed how large flux events can significantly alter the estimation of net exchanges for the Oregon shelf. After capturing a large and short-lived efflux event, their annual estimate was outgassing of 3.1 ± 82 g C per m^2 per year for this same region. The disparity illustrates the importance of basing regional flux estimates on observations 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 uptake of 7.9 g C per m^2 per year between 22° and 50°N within 370 km offshore. The most northern estimates for the North American Pacific Coast by Evans et al. (2012) and Evans and Mathis (2013) determined influxes of 26 g C per m^2 per year for British Columbian coastal waters shoreward of the 500-m isobath and 18 g C per m^2 per year for Gulf of Alaska coastal waters shoreward of the 1500-m isobath.

Models for the upwelling region (Fiechter et al., 2014; Turi et al., 2014) reproduce the pattern of CO_2 outgassing nearshore and CO_2 uptake further offshore. They also illustrate the intense eddy-driven variability nearshore. Turi et al. (2014) simulate a weak source of 0.6 ± 2.4 g C per m^2 per year for the region from 30° to 46°N, extending 800 km of shore, an amount which is inconsistent with the observations of Hales et al. (2012) that describe the same region as a sink of 7.9 g C per m^2 per year. Fiechter et al. (2014) simulate a source of atmospheric CO_2 of 0.6 Tg C per year for the region from 35° to 45°N within 600 km of shore, an estimate which is in contrast to the observation-based estimate of a 14 Tg C sink published by Hales et al. (2012). 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 during subsequent offshore advection within several hundreds of kilometers of the coast. The disagreement in mean simulated fluxes may result partly 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). Notable, however, is that observations for the Oregon shelf by Evans et al. (2015a) showed intense summer upwelling that led to strong outgassing with pronounced variability in air-sea fluxes but found only weak stimulation of primary production. The research team hypothesized that nutrient-rich waters might be subducted offshore at convergent surface temperature fronts before nutrients are fully consumed by primary producers.

Less is known about the air-sea flux of CH_4 along the North American Pacific Coast margin. Recent studies inventoried sedimentary sources of CH_4 hydrates, derived from terrestrial and coastal primary production, and suggested that extensive deposits along the Cascadia margin are beginning to destabilize because of warming (Hautala et al., 2014; Johnson et al., 2015).

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 also have been shown to increase offshore transport in the northern part of the system (Barth et al., 2002; Hales et al., 2006). Uncertainties in published estimates are high, ranging from very small (Ianson and Allen 2002; Pennington et al., 2010) to very high fractions of primary production (Hales et al., 2005; Turi et al., 2014), again as a result of the region's large spatial and temporal variability.

16.3.3 Gulf of Mexico

The Gulf of Mexico (GMx) is a semi-enclosed marginal sea at the southern coast of the conterminous United States. The passive margin shelves of its northern portion are relatively wide (up to 250 km west of Florida), but, in contrast to shelf waters of the North American Atlantic Coast, those of the GMx are not separated from 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 (Muller-Karger et al., 2015; Rivas et al., 2005). While shelf circulation is influenced primarily 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 GMx, where the Mississippi-Atchafalaya River System delivers large loads of freshwater, nutrients, and sediments.

Observational estimates indicate that the GMx, as a whole, is a weak net sink of atmospheric CO_2 with an annual average of 2.3 ± 0.96 g C per m^2 per year (Robbins et al., 2014). Robbins et al. (2014) also provide flux estimates, as follows, for smaller shelf regions, namely, the West Florida Shelf, the northern Gulf shelf, the western Gulf shelf, and the Mexico shelf. The West Florida Shelf and western Gulf shelf act as sources to the atmosphere, with estimated annual average fluxes of 4.4 ± 1.3 and 2.2 ± 0.6 g C per m^2 per year, respectively. The northern Gulf acts as a sink, with an estimated flux of 5.3 ± 4.4 g C per m^2 per year, and the Mexican shelf is almost neutral, with an estimated flux of 1.1 ± 0.6 g C per m^2 per year. Huang et al. (2015) estimated a larger uptake on the northern Gulf shelf of 11 ± 44 g C per m^2 per year (i.e., about twice the estimate of Robbins et al., 2014) and reported a much larger uncertainty. In an analysis that combines satellite and *in situ* observations, Lohrenz et al. (2018) estimated a similar uptake for the northern GMx of 13 ± 3.6 g C per m^2 per year. The overall carbon exchanges in the Gulf vary significantly from year to year because of 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), reproducing the same spatial pattern though their simulated Gulf-wide uptake of 8.5 ± 6.5 g C per m^2 per year is larger. This discrepancy results largely from a greater simulated sink in the open Gulf. Also, the uncertainty estimates of the model-simulated fluxes by Xue et al. (2016) are much larger than those of Robbins et al. (2014); the latter might be too optimistic in reporting uncertainties of the flux estimates.

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

Quantitative understanding of CH_4 dynamics in GMx coastal and oceanic environments is limited. Solomon et al. (2009) speculated that deep CH_4 hydrate seeps in the Gulf potentially are a significant CH_4 source to the atmosphere. They estimated ocean-atmosphere fluxes from seep plumes of 1,150 ± 790 to 38,000 ± 21,000 g CH_4 per m^2 per day compared with 2.2 ± 2.0 to 41 ± 8.2 g CH_4 per m^2 per day for background sites. Subsequent acoustic analyses of bubble plume characteristics question the finding that CH_4 bubbles make their way to the surface (Weber et al., 2014), and the fate of CH_4 emissions from seeps and their overall contribution to atmospheric CH_4 remain uncertain.

16.3.4 North American Arctic

The North American Arctic coastal ocean comprises broad (~300 km) shallow shelves in the Bering and Chukchi seas, the narrower (<100-km) Beaufort Sea shelf, the Hudson Bay, and the extensive Canadian Arctic Archipelago (CAA). Shelf water enters these regions from the North Pacific and follows a large-scale pathway from its entrance into the North American Arctic through the Bering Strait via the Chukchi and Beaufort seas into the CAA and, ultimately, the North Atlantic (Carmack et al., 2006, 2015). Hudson Bay receives significant inputs of freshwater (Dery et al., 2005). Except for the southernmost Bering Sea, most of the coastal region is covered with sea ice from about October to June. Areas of persistent multiyear sea ice at the northernmost extent of the CAA are rapidly declining (Stroeve et al., 2012). Reoccurring polynyas (i.e., holes in the ice) are found in all three of its major regions (Smith and Barber 2007). The North American Arctic is sparsely populated with communities heavily reliant on subsistence fishing and hunting; the rapid regional changes associated with global warming are affecting these communities. Globally, the pace of increasing air temperatures is the highest in the North American Arctic and adjacent Arctic regions, resulting in significant reductions in both summer and winter sea ice cover that profoundly affect the marine ecosystems across the northern extent of the continent (Moore and Stabeno 2015; Steiner et al., 2015).

Coastal waters in the North American Arctic have been described consistently as a net sink for

atmospheric CO_2 (Bates et al., 2006, 2011; Chen et al., 2013; Cross et al., 2014a; Dai et al., 2013; Else et al., 2008; Evans et al., 2015b; Laruelle et al., 2014; Mucci et al., 2010; Shadwick et al., 2011). This general trait is caused by low surface water pCO_2, the partial pressure of CO_2, relative to the atmosphere during ice-free months. These levels are set by the combination of low water temperatures and seasonally high rates of both ice-associated and open-water primary production (Cai et al., 2010b, 2014; Steiner et al., 2014), as well as by 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 is higher. Suppressed gas exchange through sea ice has been a source of debate within the Arctic CO_2 flux community, likely a result of inconsistencies between methodologies and the challenge of data collection in such a harsh environment, particularly during winter. The typical approach of calculating air-sea CO_2 flux (from measured air-sea pCO_2 differences and gas transfer rates parameterized using wind speed relationships) can differ markedly from flux estimations determined by eddy correlations. The latter suggest high rates of CO_2 exchange relative to open-water fluxes (Else et al., 2011). Three arguments indicate that the high, initial eddy correlation–based fluxes may be overestimates: 1) the potential for unaccounted CO_2 and water vapor cross-correlation possibly affecting the measurement (Landwehr et al., 2014); 2) independent analysis of the ^{222}Radon isotope showing near-zero gas exchange in areas covered by sea ice (Rutgers van der Loeff et al., 2014); and 3) recent demonstration of dampened gas-transfer velocities via concurrent, properly corrected eddy covariance–based fluxes and air-sea pCO_2 difference measurements in the Antarctic marginal ice zone supporting linear scaling methods that calculate fluxes using percent sea ice cover (Butterworth and Miller 2016).

However, despite the dampening effect of sea ice, 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 because rising temperatures may allow some degree of exchange to take place. To date, measurements of wintertime exchange have been limited to very few studies (Else et al., 2011, 2013; Miller et al., 2015). In recent years, the role of sea ice growth and decay has been shown to significantly affect the air-sea CO_2 flux (Rysgaard et al., 2007, 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 to sea ice DIC and the underlying seawater, because DIC is a component of the brine whereas total alkalinity precipitates in the brine channels as a form of $CaCO_3$ known as ikaite (Dieckmann et al., 2008; Rysgaard et al., 2013). During sea ice decay, ikaite dissolves, leading to excess total alkalinity relative to DIC and undersaturation of CO_2 in meltwater.

Estimates of air-sea CO_2 flux in the Chukchi and Beaufort seas, Hudson Bay, and the western CAA all indicate atmospheric CO_2 uptake (Bates et al., 2006; Else et al., 2008, 2013; Gao et al., 2012; Mucci et al., 2010; Semiletov et al., 2007; Shadwick et al., 2011; see Table 16.1, p. 657) with significantly higher uptake over the broad and productive Chukchi shelf. A recent synthesis of a decade of coastal ocean data collected within 400 km of land determined an annual mean uptake of 8.8 g C per m^2 per year (Evans et al., 2015b). Variability in wind patterns and sea ice cover affects the water column structure and connectivity between the surface ocean and overlaying atmosphere, thus influencing the magnitude of air-sea CO_2 exchange.

With regard to Arctic CH_4 fluxes, much more is known about the emission potential, distribution, and functioning of terrestrial sources (McGuire et al., 2009); knowledge of marine CH_4 sources is developing slowly due to sparse observations and the logistical challenges of Arctic marine research. The largest marine CH_4 source in the Arctic is dissociation of gas hydrates stored in continental margin sediments (Parmentier et al., 2013, 2015). As sea ice continues to retreat and ocean waters warm, CH_4 hydrate stability is expected to decrease

Table 16.2. Regional Estimates of Net Air-Sea Carbon Dioxide Exchange from Two Data Syntheses and a Process-Based Model for the MARCATS Regions[a,b]

MARCATS Segment No.[b]	MARCATS System[b]	Class	Shelf Area (10^3 km^2)	Chen et al. (2013)	Laruelle et al. (2014)	Bourgeois et al. (2016)
				Flux[a,b] (Tg C per year)	Flux[a,b] (Tg C per year)	Flux[a,b] (Tg C per year)
1	Northeastern Pacific	Subpolar	460	−19	−6.8	−10 ± 0.82
2	California Current	Eastern Boundary Current	210	−5.7	−0.13	−0.48 ± 0.15
3	Tropical Eastern Pacific	Tropical	200	−0.1	0.19	−0.22 ± 0.095
9	Gulf of Mexico	Marginal Sea	540	−1.3	−2.1	−4.5 ± 0.63
10	Florida Upwelling	Western Boundary Current	860	−11	−2.7	−15 ± 1.3
11	Labrador Sea	Subpolar	400	−10	−19	−8.8 ± 1.2
12	Hudson Bay	Marginal Sea	1100	11	NA	−3.8 ± 3.4
13	Canadian Arctic Archipelago	Polar	1200	−57	−14	−6.2 ± 0.75
	Total		4900	−94	−44	−49

Notes

a) Positive fluxes indicate a source to the atmosphere.

b) MARCATS, MARgins and CATchments Segmentation; C, carbon; CO_2, carbon dioxide; Tg, teragrams; g, gram; Tg = 10^{12} g

with potentially large and long-term implications. An additional potential marine CH_4 source, unique to polar settings, is release from subsea permafrost layers, with fluxes from thawed sediments reported to be orders of magnitude higher than fluxes from adjacent frozen sediments (Shakhova et al., 2015).

16.3.5 Summary Estimates for CO_2 Uptake by North American Coastal Waters

Despite the variability in regional estimates discussed above and summarized in Table 16.1, p. 657, North American coastal waters clearly act as a net sink of atmospheric carbon. Because of discrepancies among studies, these various regional estimates would be difficult to combine into one number with any confidence. Instead, this chapter considers estimates of net air-sea CO_2 exchange in North American coastal waters from two global data syntheses (Chen et al., 2013; Laruelle et al., 2014) and a process-based global model (Bourgeois et al., 2016; see Table 16.2, this page). The data syntheses use a global segmentation of the coastal zone and associated watersheds known as MARCATS (MARgins and CATchments Segmentation; Laruelle et al., 2013), which, at a resolution of 0.5°, delineates a total of 45 coastal segments, eight of which surround North America. The data synthesis of Chen et al. (2013) is a summary of individual studies, whereas Laruelle et al. (2014) analyze the Surface Ocean CO_2 Atlas 2.0 database (Bakker et al., 2014) to derive regional estimates. The data syntheses of Chen et al. (2013) and Laruelle et al. (2014)

estimate the North American coastal uptake to be 94.4 and 44.5 Tg C per year, respectively, and the process-based model of Bourgeois et al. (2016) estimates an uptake of 48.8 Tg C per year (see Table 16.2, p. 665). Although there are significant regional discrepancies between the latter two estimates for the eastern tropical Pacific Ocean (i.e., the Central American Isthmus), the GMx, the Florida Upwelling region (actually covering the eastern United States), the Labrador Sea, and the CAA, the overall flux estimates for North America are in close agreement. This, and the fact that Laruelle et al. (2014) used a consistent methodology to estimate air-sea CO_2 flux, builds some confidence in these numbers.

The net CO_2 flux and its anthropogenic component from the process-based global model of Bourgeois et al. (2016) are also reported for a regional decomposition of the EEZs of the United States, Canada, and Mexico (see Table 16.3, this page) in Table 16.4, p. 667. The model simulates a net uptake of CO_2 in North American EEZ coastal waters (excluding the EEZ of the Hawaiian and other islands) of 160 Tg C per year with an anthropogenic flux contribution of 59 Tg C per year. This chapter adopts 160 Tg C per year as the net uptake by coastal waters of North America, excluding tidal wetlands and estuaries. Unfortunately, there are no formal error estimates for this uptake. Instead, estimates adopted here project an error by first noting that the Bourgeois et al. (2016) model is in good agreement with the more recent of the two observation-based estimates for the MARCATS regions of North America. Furthermore, the error estimate for the uptake by continental shelves globally is about 25%, with the North American MARCATS regions having mainly "fair" data quality (Laruelle et al., 2014). Hence, assuming an error of ±50% for the uptake by North American EEZ waters seems reasonable.

Table 16.3. Subregions of the Combined Exclusive Economic Zone of Canada, the United States, and Mexico[a]

Region Number	Area (10^3 km^2)	Acronym	Name
1	500	MAB	Mid-Atlantic Bight
2	160	GOM	Gulf of Maine
3	220	SS	Scotian Shelf
4	860	GStL	Gulf of St. Lawrence and Grand Banks
5	1,100	LS	Labrador Shelf
6	1,200	HB	Hudson Bay
7	1,000	CAA	Canadian Arctic Archipelago
8	950	BCS	Beaufort and Chukchi Seas
9	2,200	BS	Bering Sea
10	1,500	GAK	Gulf of Alaska
11	460	CCSN	Northern California Current System
12	640	CCSC	Central California Current System
13	1,200	CCSS	Southern California Current System
14	1,400	Isthmus	Isthmus
15	1,600	GMx	Gulf of Mexico and Yucatan Peninsula
16	500	SAB	South Atlantic Bight
17	7,500	Islands	Hawai'i and other Pacific and Caribbean islands

Notes

a) Area is calculated for the mask that was used to define subregions for averaging.

16.3.6 Summary Carbon Budget for North American Coastal Waters

Combining the atmospheric CO_2 uptake estimate with estimates of carbon transport from land and carbon burial in ocean sediments enables a first attempt at constructing a carbon budget for the North American EEZ (see Table 16.5, p. 668). Carbon delivery to the coastal ocean from land via rivers and from tidal wetlands after estuarine processing (i.e., CO_2 outgassing and carbon burial in estuaries) is estimated to be 106 ± 30 Tg C per year (see Ch. 15:

Table 16.4. Estimates of Carbon Burial and Primary Production,[a] Net Primary Production (NPP),[b] and Simulated NPP and Air-Sea Exchange of Carbon Dioxide[c] for the Exclusive Economic Zone Decomposition in Table 16.2[d,e,g]

Region[f]	Carbon Burial[a]		Satellite NPP[a]		Satellite NPP[b]	NPP from Global Model[c]		Air-Sea Exchange of CO_2[c]	
Number and Acronym	g C per m^2 per year	Tg C per year	g C per m^2 per year	Tg C per year	Tg C per year	g C per m^2 per year	Tg C per year	g C per m^2 per year	Tg C per year
1, MAB	23	101	360	170	170	260	120	31 (14)	15 (6.8)
2, GOM	46	5.5	490	58	81	180	26	33 (7.1)	4.9 (1.1)
3, SS	9.8	2.0	300	63	64	170	43	33 (11)	8.2 (2.8)
4, GStL	16	11	260	190	230	150	130	24 (6.5)	21 (5.6)
5, LS	2.3	2.3	120	120	70	82	88	33 (9.5)	36 (10)
6, HB	19	17.1	144	130	13	130	150	–0.48 (1.4)	–0.50 (1.7)
7, CAA	2.6	1.6	42	26	Not available	19	20	4.1 (0.96)	4.3 (0.96)
8, BCS	12	10	120	110	Not available	49	47	8.0 (1.2)	7.6 (1.1)
9, BS	17	34	240	490	470	130	270	13 (4.0)	28 (8.6)
10, GAK	7.2	10.0	260	360	420	130	210	19 (4.6)	29 (7.1)
11, CCSN	6.1	2.54	270	110	150	160	73	9.4 (4.2)	4.3 (1.9)
12, CCSC	1.2	0.65	260	150	210	170	110	1.1 (4.4)	0.72 (2.9)
13, CCSS	0.99	1.1	210	230	280	150	190	–4.3 (3.1)	–5.5 (4.0)
14, Isthmus	0.42	0.53	230	300	210	150	200	–2.3 (3.6)	–3.2 (4.9)
15, GMx	6.2	8.7	250	350	390	220	360	4.8 (3.7)	7.9 (6.2)
16, SAB	5.4	2.4	210	92	110	260	130	9.7 (6.6)	5.0 (3.4)
17, Islands	0.0055	0.041	120	890	580	80	620	–1.4 (4.1)	–11 (31)
Total	NA	120	NA	3,400	NA	NA	2,800	NA	150 (100)
Total w/o 17	NA	120	NA	2,500	NA	NA	2,200	NA	160 (59)

Notes

a) Dunne et al. (2007).

b) Balcom and Continental Shelf Associates (2011).

c) Bourgeois et al. (2016).

d) Included in carbon dioxide (CO_2) exchange estimates are total and anthropogenic fluxes calculated by averaging the model years 1993–2012. Here all fluxes are relative to the coastal ocean reservoir (i.e., positive fluxes are a source to the coastal ocean, while negative fluxes are a sink).

e) NPP, net primary production; g, grams; C, carbon; Tg, teragrams.

f) See Table 16.3, p. 666, for region descriptions.

g) Key: g C, grams of carbon; Tg C, teragrams of carbon.

Tidal Wetlands and Estuaries, p. 596). Estimates of carbon burial, based on the method of Dunne et al. (2007) for the regional decomposition of the North American EEZ, are reported in Table 16.4, p. 667, with a total flux of 120 Tg C per year. Here these fluxes are considered to be an upper bound because they are substantially larger than other estimates. The Dunne et al. (2007) global estimates of organic carbon burial in waters shallower than 200 m are 19 ± 9 g C per m^2 per year, much larger than the estimates of 6 and 1 g C per m^2 per year by Chen (2004) and Muller-Karger et al. (2005), respectively, although areas are slightly different in the three studies. The organic carbon burial estimates of Dunne et al. (2007) for the GOM, MAB, and SAB (see Table 16.4, p. 667) are larger by factors of 8, 17, and 3, respectively, than the best estimates of the empirical model of Najjar et al. (2018). However, due to different definitions of the boundary between coastal waters and the open ocean, the combined area of the GOM, MAB, and SAB in Najjar et al. (2018) is about a third of that in Dunne et al. (2007). Finally, Dunne et al. (2007) estimated the organic carbon burial in Hudson Bay to be 19 g C per m^2 per year, compared to a mean estimate of 1.5 ± 0.7 g C per m^2 per year of burial from sediment cores (Kuzyk et al., 2009). Given these results, SOCCR2 considers the estimates of Dunne et al. (2007) to be an upper bound and assumes that a reasonable lower bound is about an order of magnitude smaller, thus placing the North American organic carbon burial estimate at 65 ± 55 Tg C per year.

If these estimates of net air-sea flux, carbon burial, and carbon input from land are accurate, then the residual must be balanced by an increase in carbon inventory in coastal waters and a net transfer of carbon from coastal to open-ocean waters. In their global compilation, Regnier et al. (2013) report an increase in the coastal carbon inventory of 50 Tg C per year, which is a quarter of their estimated anthropogenic carbon uptake by air-sea exchange in the coastal waters of 200 Tg C per year. The latter estimate is uncertain. In their global modeling study, which did not account for anthropogenic changes in carbon delivery from land, Bourgeois et al. (2016) estimated an accumulation of carbon in the coastal ocean of 30 Tg C per year. This amount is a third of their estimated uptake of anthropogenic carbon from air-sea gas exchange in the coastal ocean of 100 Tg C per year and approximately half of their estimated cross-shelf export of anthropogenic carbon of 70 Tg C per year. The rate of carbon accumulation in the North American EEZ from the model of Bourgeois et al. (2016) is 50 Tg C per year (see Table 16.5, this page). Here again, this chapter assumes an uncertainty of ±50%. The residual of 151 ± 105 Tg C per year is the inferred export of carbon to the open ocean (see Table 16.5, this page). The fact that the error in this residual is large in absolute and relative terms emphasizes the need for more accurate carbon budgets for coastal waters of North America. The challenge, however, is that many of these terms are small compared to internal carbon cycling in coastal waters, which is dominated by primary production and respiration. Two separate estimates of primary production (see Table 16.4, p. 667) are in broad agreement and reveal that terms in the Table 16.5 budget are just a few percent of primary

Table 16.5. Approximate Summary Carbon Budget for the Exclusive Economic Zone of North America[a–d]

Process	Flux (Tg C per year)[b,d]
Input from land	106 ± 30
Uptake from atmosphere	160 ± 80
Burial	–65 ± 55
DIC[c] accumulation in coastal waters	–50 ± 25
Inferred open-ocean export (residual)	–151 ± 105

Notes

a) Exclusive Economic Zone (EEZ) excludes EEZs of the Hawaiian and other islands.

b) Positive fluxes are a source to the coastal ocean, while negative fluxes are a sink.

c) The accumulation of dissolved inorganic carbon (DIC) is reported with a negative sign to illustrate that all fluxes balance.

d) Tg C, teragrams of carbon.

production. This also emphasizes that small changes in carbon cycling in coastal waters can result in large changes in atmospheric uptake and transport to the open ocean.

16.4 Climate Trends and Feedbacks

16.4.1 Trends in Coastal Carbon Fluxes

Important questions with respect to coastal carbon fluxes include:

- What is the anthropogenic component of the CO_2 sink?
- How will the coastal ocean change as a CO_2 sink?
- How will changing climate and other forcings affect the total and anthropogenic flux proportions?

As stated in Section 16.2, p. 652, when considering the ocean's role in sequestering anthropogenic carbon, the relevant component is anthropogenic flux, not the total uptake flux. Neither quantifying the anthropogenic carbon flux component nor predicting its future trend is straightforward. Here the likely trends in total carbon fluxes are described; by definition, changes in total carbon fluxes imply changes in anthropogenic fluxes as well.

A direct effect of increasing atmospheric CO_2 will 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) may affect carbon fluxes in North American coastal waters. Ocean warming reduces the solubility of gases and thus directly affects gas concentrations near the surface; this likely will decrease the net air-sea flux of CO_2 by reducing the undersaturation of CO_2 (see Cahill et al., 2016, for the North American Atlantic Coast). Surface temperature increases also strengthen vertical stratification and thus impede vertical mixing, effects which will affect upward diffusion of nutrients and DIC. Enhanced stratification, therefore, could lead to decreases in both biologically driven carbon uptake and CO_2 outgassing. However, model projections for the northern GMx show that the direct effect of increasing atmospheric CO_2 overwhelms the other more secondary effects (Laurent et al., 2018). Furthermore, temperature trends in coastal waters around North America show complex patterns with some regions having cooled from 1982 to 1997 followed by warming from 1997 to 2013 (e.g., the MAB), some regions having warmed from 1982 to 1997 followed by cooling from 1997 to 2013 (e.g., the SAB and Gulf of Alaska), and other regions showing no consistent warming from 1982 to 2013 (e.g., the NAA; Liao et al., 2015). Temperature anomalies from a time series in the central California Current System show warm surface waters for the decade prior to 1997 followed by a prolonged cooler period until the strong surface warming associated with a marine heatwave and the 2015 to 2016 El Niño interrupted the cool anomalies (Chavez et al., 2017). However, deeper waters in the California Undercurrent have shown a multidecadal trend (1980 to 2012) toward warmer, saltier, lower-oxygen, and higher-CO_2 waters at a depth associated with increased northward transport of Pacific equatorial waters (Meinvielle and Johnson 2013).

Some studies suggest that trends in the air-sea pCO_2 gradient (ΔpCO_2) are indicative of a strengthening or weakening of the net CO_2 uptake by shelf systems, where an increasing ΔpCO_2, implying that ocean pCO_2 rises more slowly than atmospheric pCO_2, corresponds to increased net uptake and cross-shelf export (Laruelle et al., 2018). In their observation-based analysis of decadal trends in shelf pCO_2, Laruelle et al. (2018) found that coastal waters lag compared to the rise in atmospheric CO_2 in most regions. For North American coastal waters, they found that the MAB has an increase in ΔpCO_2 of 1.9 ± 3.1 microatmospheres (µatm) per year, a finding which means that in this region surface ocean pCO_2 does not increase or else increases at a rate that is substantially slower than in the atmosphere. For the shelves of the Labrador Sea, the Vancouver Shelf, and the SAB, they found rates of 0.68 ± 0.61 µatm per year, 0.83 ± 1.7 µatm per year, and 0.51 ± 0.74 µatm per year, respectively, implying that surface ocean pCO_2 does not increase or increases at a slower rate than atmospheric CO_2. The only North

American coastal region that exhibits a negative trend is the Bering Sea, with –1.1 ± 0.74 µatm per year, meaning that surface ocean pCO_2 increases at a faster rate than in the atmosphere. Laruelle et al. (2018) concluded that the lag in coastal ocean pCO_2 increase compared to that in the atmosphere in most regions indicates an enhancement in the coastal uptake and export of atmospheric CO_2, although they did not investigate alternative explanations.

Trends in coastal ocean uptake of pCO_2 are highly variable regionally and result from a complex interplay of factors. In coastal upwelling systems, surface warming will 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 North American Pacific Coast may be greater and more persistent CO_2 outgassing nearshore and lower productivity offshore as upwelled nitrate is exported before it can be used by the phytoplankton community (Evans et al., 2015a). Rates of warming clearly are faster in higher latitudes, but predicting the net effect of these warming-induced changes in the North American Arctic is not easy. Furthermore, warming in the Arctic leads to reductions in ice cover and longer 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 example, for the North American Atlantic Coast, changes in wind stress were shown to significantly modify air-sea fluxes (Cahill et al., 2016; Previdi et al., 2009). Large-scale changes in wind patterns also affect ocean circulation with a range of implications (Bakun 1990). Upwelling-favorable winds along the North American Pacific Coast have intensified in recent years, especially in the northern parts of the upwelling regimes (García-Reyes et al., 2015; Rykaczewski and Checkley 2008; Rykaczewski et al., 2015; Sydeman et al., 2014), a change which has led to 1) shoaling of subsurface nutrient-rich waters (Aksnesa and Ohman 2009; Bograd et al., 2015), 2) increased productivity (Chavez et al., 2011, 2017; Jacox et al., 2015; Kahru et al., 2015), 3) higher DIC delivery to the surface (Turi et al., 2016), and 4) declining oxygen levels (Crawford and Peña 2016; Peterson et al., 2013; Bograd et al., 2015). In the North American Arctic, late-season air-sea CO_2 fluxes may become increasingly more directed toward the atmosphere as Arctic low-pressure systems with storm-force winds occur more often over open water, thus ventilating CO_2 respired from the high organic carbon loading of the shallow shelf (Evans et al., 2015b; Hauri et al., 2013; Steiner et al., 2013) and affecting net annual exchanges. The intense warming observed across the North American Arctic also influences mid-latitude weather patterns (Kim et al., 2014), with probable cascading effects on CO_2 exchanges through adjustments in the wind field.

16.4.2 Acidification Trends in North America's Coastal Ocean

Increasing atmospheric CO_2 emissions lead to rising atmospheric CO_2 levels (see Figure 16.3, p. 671) and a net ocean uptake of CO_2. Since about 1750, the ocean has absorbed 27% of anthropogenic CO_2 emissions to the atmosphere from fossil fuel burning, cement production, and land-use changes (Canadell et al., 2007; Le Quéré et al., 2015; Sabine and Tanhua 2010). As a result of this uptake, the surface ocean pCO_2 has increased (see Figure 16.3, p. 671) and oceanic pH, carbonate ion concentration, and carbonate saturation state have decreased (Caldeira and Wickett 2003; Feely et al., 2004, 2009; Orr et al., 2005). Commonly called ocean acidification, this suite of chemical changes is defined more precisely as "any reduction in the pH of the ocean over an extended period, typically decades or longer, that is caused primarily by uptake of CO_2 from the atmosphere but also can be caused by other chemical additions or subtractions from the ocean" (IPCC 2011, p. 37). In addition to uptake of

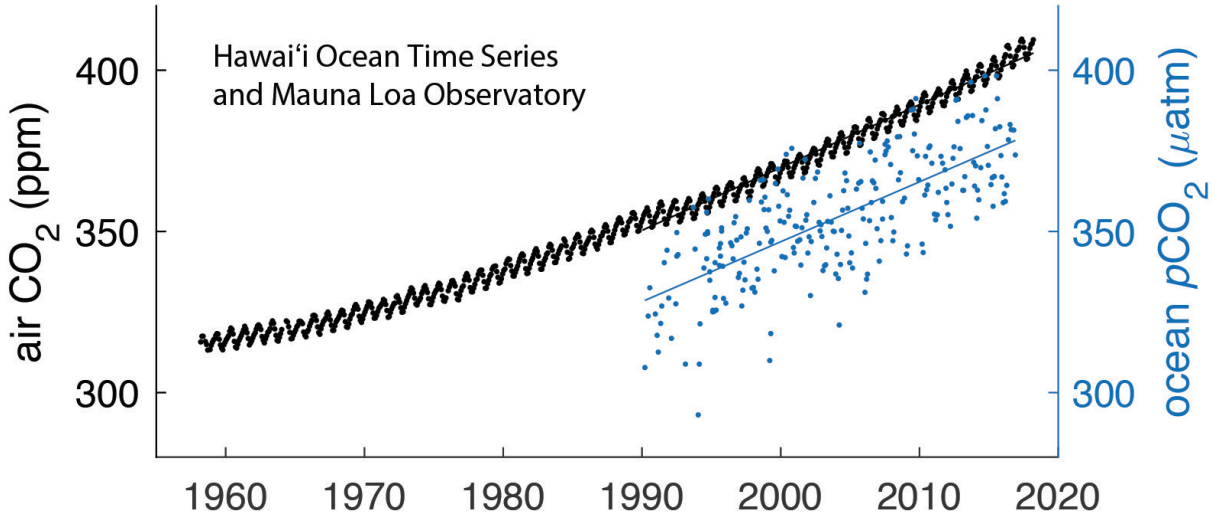

Figure 16.3. Trends in Measured Atmospheric Carbon Dioxide (CO_2) and Surface Ocean Partial Pressure of CO_2 (pCO_2). Black dots represent atmospheric CO_2 measured in parts per million (ppm) at the Mauna Loa Observatory in Hawai'i beginning in 1958. Surface ocean pCO_2 data (blue dots) are measured in microatmospheres (µatm) from the Hawai'i Ocean Time-series (HOT) station near Hawai'i (see Figure 16.4, p. 672, for site location). Black and blue lines indicate linear trends after 1990. Atmospheric CO_2 increased by 1.86 ppm per year; surface ocean pCO_2 increased by 1.95 µatm per year. [Data sources: Mauna Loa, www.esrl.noaa.gov/gmd/ccgg/trends/data.html; HOT, hahana.soest.hawaii.edu/hot/hot-dogs/interface.html.]

CO_2 from the atmosphere, variations in DIC concentrations and thus pH can be caused by biological production and respiration. Ocean acidification can significantly affect growth, metabolism, and life cycles of marine organisms (Fabry et al., 2008; Gattuso and Hansson 2011; Somero et al., 2016) and most directly affects marine calcifiers, organisms that precipitate $CaCO_3$ to form internal or external body structures. When the carbonate saturation state decreases below the equilibrium point for carbonate precipitation or dissolution, conditions are said to be corrosive, or damaging, to marine calcifiers. These conditions make it more difficult for calcifying organisms to form shells or skeletons, perform metabolic functions, and survive.

Acidification trends in open-ocean surface waters tend to occur at a rate that is commensurate with the rate of the increase in atmospheric CO_2 (see, for example, trends of atmospheric CO_2 in comparison to surface ocean pCO_2 at the Hawai'i Ocean Time-series in Figure 16.3, this page). Acidification in coastal waters is more variable because of a combination of changes in circulation and upwelling, larger-amplitude seasonal signals in production and respiration than in the open ocean, and atmospheric CO_2 uptake (see Figure 16.4, p. 672; Feely et al., 2008, 2016, 2018; Chavez et al., 2017). In many coastal regions, pCO_2 rises more slowly than in the open ocean (see Section 16.4.1, p. 669; Laruelle et al., 2018). Along the North American Pacific Coast, climate-driven changes in upwelling circulation result in coastal acidification events. As mentioned in Section 16.4.1, upwelling-favorable winds along this coast have intensified over recent years, especially in the northern parts of the upwelling regimes (García-Reyes et al., 2015; McClatchie et al., 2016; Rykaczewski and Checkley 2008; Rykaczewski et al., 2015; Sydeman et al., 2014). Intensified upwelling supplies deep water to the shelf that is rich in DIC and nutrients but poor in oxygen. Ocean acidification and hypoxia thus are strongly linked ecosystem stressors because

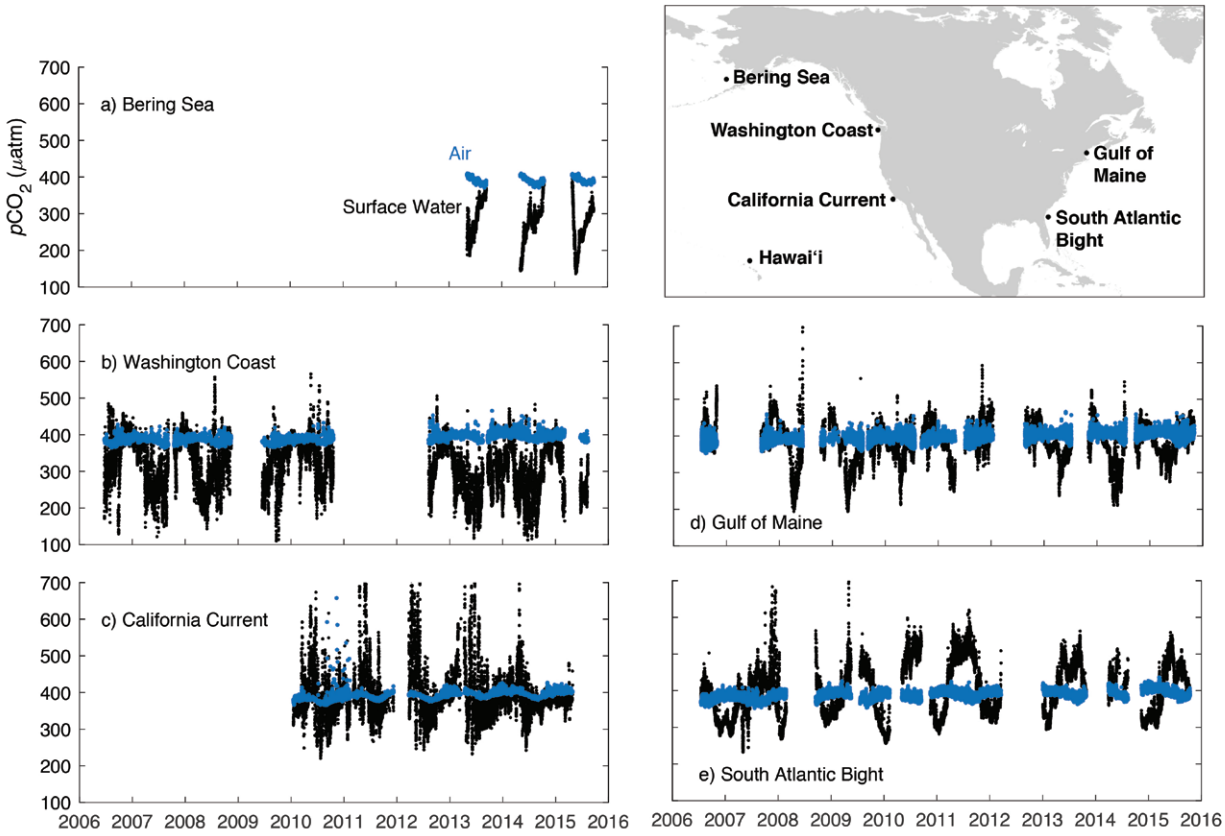

Figure 16.4. Partial Pressure of Carbon Dioxide (pCO_2) Data for the Surface Ocean (black) and Overlying Atmosphere (blue) at Five Coastal Sites. Data are in microatmospheres (µatm); map shows mooring locations. [Data sources: Bering Sea (mooring M2), Cross et al., 2014b. Washington coast (Cape Elizabeth mooring), Mathis et al., 2013. California Current (mooring CCE2), Sutton et al., 2012. Coastal Western Gulf of Maine mooring, Sutton et al., 2013. South Atlantic Bight (Gray's Reef mooring), Sutton et al., 2011.]

low-oxygen, high-CO_2 conditions derive from the microbial respiration of organic matter (Chan et al., 2016; Feely et al., 2008, 2016, 2018). In the northern California Current System, pCO_2, pH, and aragonite saturation reach levels known to be harmful to ecologically and economically important species during the summer upwelling season (see Ch. 17: Biogeochemical Effects of Rising Atmospheric Carbon Dioxide, p. 690; Barton et al., 2012, 2015; Bednaršek et al., 2014, 2016, 2017; Feely et al., 2008, 2016; Harris et al., 2013). In the Gulf of Alaska, aragonite saturation drops to near saturation values during the winter months when deep mixing occurs and surface ocean pCO_2 exceeds atmospheric pCO_2 (Evans and Mathis 2013). Along the Pacific Coast, 50% of shelf waters are projected to experience year-long undersaturation by 2050 (Gruber et al., 2012; Hauri et al., 2013; Turi et al., 2016).

Polar regions are naturally prone to acidification because of their low temperatures (Orr et al., 2005; Steinacher et al., 2009). In many Arctic coastal regions, pH and carbonate saturation state are naturally low relative to lower-latitude coastal settings. These low levels result from higher CO_2 solubility, the influence of multiple sources of freshwater (e.g., 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 relatively high rates of acidification in polar waters is retreating sea ice, which adds meltwater from multiyear ice and increases the surface area of open water, thereby enhancing the uptake of atmospheric CO_2 (Cai et al., 2010b; Steiner et al., 2013). 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 (Feely et al., 2009; Mathis et al., 2015a). Models predict annual average aragonite undersaturation (i.e., favoring dissolution) for the Bering Sea and the Chukchi Sea by 2070 and 2030, respectively (Mathis et al., 2015a). The Beaufort Sea upper halocline and deep waters now regularly show aragonite undersaturation (Mathis et al., 2015a; Miller et al., 2014). These chemical seawater signatures are propagated via M'Clure Strait and Amundsen Gulf into the CAA and beyond (Azetsu-Scott et al., 2010; Turk et al., 2016; Yamamoto-Kawai et al., 2013). Model projections based on the IPCC high-CO_2 emissions scenario, Representative Concentration Pathway 8.5 (RCP8.5), suggest the Beaufort Sea surface water will become undersaturated with respect to aragonite around 2025 (Steinacher et al., 2009; Steiner et al., 2014). As these conditions intensify, negative impacts on calcifying marine organisms are expected to become a critical issue, reshaping ecosystems and fisheries across the North American Arctic domain (Mathis et al., 2015b; Moore and Stabeno 2015).

In the northern GMx, 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). Here excessive nutrient inputs from the Mississippi River result in hypoxia and eutrophication-induced acidification of near-bottom waters (Cai et al., 2011; Laurent et al., 2017). Similar to the California Current System, low-oxygen and high-CO_2 conditions coincide and derive from microbial respiration of organic matter (Cai et al., 2011; Laurent et al., 2017; Feely et al., 2018). Currently, aragonite saturation states are around 2 in hypoxic bottom waters and thus well above the saturation threshold. Projections suggest that aragonite saturation states of these near-bottom waters will drop below the saturation threshold near the end of this century (Cai et al., 2011; Laurent et al., 2018).

Recent studies indicate that the northern regions of the North American Atlantic Coast (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 and lower aragonite saturation states than more southern coastal regions. These properties are driven primarily by a decrease in mean total alkalinity of shelf water from the SAB northward to the GOM. Seasonal undersaturation of aragonite in subsurface water is occurring in the GOM with photosynthesis and respiration playing a major role in controlling the seasonal variability of aragonite saturation states; dissolution of aragonite might already occur in fall and winter (Wang et al., 2017). With a significant shellfish industry, the GOM displays the lowest pH and aragonite saturation levels along the East Coast in summer (Wang et al., 2013).

16.5 Conclusions

The research community has made tremendous progress in improving understanding and constraining rates of carbon cycling in coastal waters since SOCCR1 (CCSP 2007), 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 the interfaces between land and coastal ocean, coastal and open-ocean waters, and water and sediment. Net exchange fluxes and trends are relatively small signals masked by a large and fluctuating background. At present, most of these fluxes are not quantified well enough to derive well-constrained carbon budgets for North American coastal waters or to project how those 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. 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 thus far elusive for coastal waters.

Estimates of air-sea CO_2 fluxes currently provide the best evidence for the contribution of coastal waters to overall carbon uptake by the ocean. In the broad shelf system of the North American Atlantic Coast, shelf water is separated from the adjacent open ocean by persistent shelf break currents and density fronts. Available estimates suggest that the overall North American Atlantic Coast is a weak sink, with some subregions acting as sources (e.g., nearshore regions of the SAB), while others are either neutral (Scotian Shelf and GOM) or act as weak sinks (MAB and outer SAB). Large sections of the narrow shelf of the North American Pacific Coast are dominated by upwelling circulation, which leads to strong CO_2 outgassing near the coast. However, compensating for this outgassing is 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 GMx are considered a weak net sink, with the West Florida Shelf and the western Gulf shelf acting as sources; the Mexico shelf being neutral; and only the northern shelf a clear sink that is driven largely by anthropogenic nutrient inputs from the Mississippi River. The wide, seasonally ice-covered shelves in the North American Arctic consistently are acting as a sink for atmospheric CO_2. The low surface-water pCO_2 in this region primarily results from low water temperatures and the decreased uptake of atmospheric CO_2 during a significant fraction of the year because of seasonal ice cover. Overall, North American coastal waters act as a sink, but regional variations and uncertainties are large.

Several drivers influence secular trends in coastal carbon fluxes and will continue to do so in the future. These drivers include 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 (particularly, the nitrogen cycle). Coastal surface pCO_2 clearly does not closely track atmospheric pCO_2. Although there are a number of plausible mechanisms for potential future changes in coastal carbon uptake, the total effect cannot be predicted with any confidence. Regional model studies are beginning to address these challenges.

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, one of the more soluble forms of biogenic $CaCO_3$ in the ocean. Aragonite saturation states are routinely below saturation (i.e., favoring dissolution) in North American Arctic coastal waters. In the North American Pacific Coast region, atmospheric CO_2 uptake in combination with intensified upwelling that brings low-pH, low-oxygen water onto the shelves leads to aragonite levels below the saturation threshold in large portions of the subsurface waters. In the northern GMx, 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 assessments of the carbon cycle in these regions are paramount. Critical needs are maintaining and expanding existing coastal observing programs, continuing national and international coordination and integration of observations, increasing development of modeling capabilities, and addressing stakeholder needs.

SUPPORTING EVIDENCE

KEY FINDING 1

Observing networks and high-resolution models are now available to construct coastal carbon budgets. Efforts have focused primarily on quantifying the net air-sea exchange of carbon dioxide (CO_2), but some studies have estimated other key fluxes, such as the exchange between shelves and the open ocean.

Description of evidence base

Observing networks are in place along the Atlantic, Pacific, and Arctic coasts of North America and the U.S. Gulf Coast (Alin et al., 2015; Bates et al., 2006, 2011; Cai et al., 2010a; Chen et al., 2013; Cross et al., 2014a; Dai et al., 2013; DeGrandpre et al., 2002; Evans et al., 2011, 2012, 2015b; Hales et al., 2005, 2012; Jiang et al., 2008; Mucci et al., 2010; Najjar et al., 2018; Robbins et al., 2009, 2014; Salisbury et al., 2008b, 2009; Shadwick et al., 2010, 2011; Vandemark et al., 2011; Wang et al., 2013, 2017).

Regional models are in place for the same regions (Cahill et al., 2016; Fennel et al., 2008; Fiechter et al., 2014; Pilcher et al., 2018; Previdi et al., 2009; Turi et al., 2014; Xue et al., 2016).

The emphasis on quantifying air-sea exchange is illustrated by the fact that the references listed in Table 16.1, p. 657, all provide an estimate of this flux, but few provide estimates of other fluxes. Few studies exist that do provide estimates of carbon exchange between shelves and open ocean; they include Fennel and Wilkin (2009), Barth et al. (2002), Hales et al. (2006), Xue et al. (2016), and Najjar et al. (2018).

Major uncertainties

This key message essentially contains statements of fact. Hence, this statement is not considered uncertain.

KEY FINDING 2

Available estimates of air-sea carbon fluxes, based on more than a decade of observations, indicate that the North American margins act as a net sink for atmospheric CO_2. This net uptake is driven primarily by fluxes in the high-latitude regions. The estimated magnitude of the net flux is 160 ± 80 teragrams of carbon per year (*medium confidence*) for the North American Exclusive Economic Zone, a number that is not well constrained.

Description of evidence base

This statement is supported by the numbers summarized in Tables 16.1, p. 657, and 16.2, p. 665. Consistent reports of outgassing exist only for the Gulf of Maine (GOM), where the net flux is almost neutral, and the West Florida Shelf. Contradictory reports exist for the Scotian Shelf. Everywhere else the net flux is reported as net uptake (i.e., sink), although with large uncertainties. Three independent studies also provide estimates of net air-sea CO_2 exchange in North American coastal waters. Two are global data syntheses (Chen et al., 2013; Laruelle et al., 2014), and one is from a process-based global model (Bourgeois et al., 2016; see Table 16.2, p. 665). The model of Bourgeois et al. (2016) estimates a net air-sea CO_2 flux of 160 teragrams of carbon

(Tg C) per year for the North American Exclusive Economic Zone (EEZ). The estimate is that the uncertainty is 50%.

These individual estimates cannot be combined because of discrepancies in numbers and gaps in coverage.

Major uncertainties

The consistency among studies pointing at North American coastal waters as a sink provides confidence, although each individual estimate is uncertain.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

The statement that North American coastal waters act as a sink overall can be made with high confidence and reflects the fact that studies are consistent in supporting this conclusion, even though each number itself comes with a large uncertainty. The overall uptake estimate is uncertain; hence, there is high confidence in stating that this flux estimate is poorly constrained.

Summary sentence or paragraph that integrates the above information

The consistency of many independent estimates reporting coastal uptake of atmospheric CO_2 builds confidence that these waters indeed act as a sink.

KEY FINDING 3

The increasing concentration of CO_2 in coastal and open-ocean waters leads to ocean acidification. Corrosive conditions in the subsurface occur regularly in Arctic coastal waters, which are naturally prone to low pH, and North Pacific coastal waters, where upwelling of deep, carbon-rich waters has intensified and, in combination with the uptake of anthropogenic carbon, leads to low seawater pH and aragonite saturation states in spring, summer, and early fall (*very high confidence, very likely*).

Description of evidence base

In Arctic coastal waters, pH and carbonate saturation state are naturally low (Cai et al., 2010b; Mathis et al., 2011; Steiner et al., 2013). The pace of ocean acidification is faster in the Arctic than in other coastal and open-ocean regions (Fabry et al., 2009; Feely et al., 2009; Mathis et al., 2015a). The Beaufort Sea upper halocline and deep waters now regularly show aragonite undersaturation (Mathis et al., 2015a; Miller et al., 2014). These chemical seawater signatures are propagated via M'Clure Strait and Amundsen Gulf into the Canadian Archipelago and beyond (Azetsu-Scott et al., 2010; Turk et al., 2016; Yamamoto-Kawai et al., 2013). Variability in the carbon content of freshwater end members also has been shown to contribute to undersaturation events in coastal waters of the Gulf of Alaska region (Siedlecki et al., 2017; Evans et al., 2014)

In the North America Pacific Coast (NAPC) region, anthropogenic CO_2 uptake combined with climate-driven changes in upwelling circulation result in coastal acidification events. Upwelling-favorable winds along the NAPC have intensified over recent years, especially in the northern parts of the upwelling regimes (García-Reyes et al., 2015; McClatchie et al., 2016; Rykaczewski and Checkley 2008; Rykaczewski et al., 2015; Sydeman et al., 2014). In the northern California Current System, pCO_2 (partial pressure of CO_2), pH, and aragonite saturation

reach levels known to be harmful to ecologically and economically important species during the summer upwelling season (see Ch. 17: Biogeochemical Effects of Rising Atmospheric Carbon Dioxide, p. 690; Barton et al., 2012, 2015; Bednaršek et al., 2014, 2016, 2017; Feely et al., 2008, 2016, 2018; Harris et al., 2013; Siedlecki et al., 2016).

Major uncertainties

Statement is well supported by the literature. No major uncertainties.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

Statement is well supported by the literature. No major uncertainties.

Estimated likelihood of impact or consequence, including short description of basis of estimate

Corrosive waters have been observed in the Arctic and North Pacific coastal regions (Feely et al., 2008, 2016; Mathis et al., 2015a; Miller et al., 2014). A more comprehensive list of references is given in the description above and in the chapter body.

Summary sentence or paragraph that integrates the above information

Statement that corrosive waters regularly occur is well supported by the literature because these conditions have been directly observed. There are no major uncertainties.

KEY FINDING 4

Expanded monitoring, more complete syntheses of available observations, and extension of existing model capabilities are required to provide more reliable coastal carbon budgets, projections of future states of the coastal ocean, and quantification of anthropogenic carbon contributions.

Description of evidence base

The underlying motivation for constructing complete carbon budgets for coastal waters is that permanent burial of carbon in coastal sediments and export of carbon from coastal waters to the deep ocean 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. Only total fluxes can be observed directly. Distinction between anthropogenic fluxes and the natural background has not been attempted in regional observational or modeling studies, because more comprehensive accounting than is available for carbon fluxes and improved modeling capabilities would be needed. The study by Bourgeois et al. (2016) is the first to estimate coastal anthropogenic carbon uptake in a global model. The estimated net air-sea exchange of CO_2 from this global model is reported for a regional decomposition of the EEZs of the United States, Canada, and Mexico in Table 16.3, p. 666. The model simulates a net uptake of CO_2 in North American coastal waters that is of similar magnitude to estimates of organic carbon burial and riverine carbon input, but the latter two numbers are uncertain because they are each taken from one individual study and not corroborated by multiple references. However, the similar magnitudes of these numbers illustrate that current coastal carbon budgets are uncertain and that constraining just the air-sea gas exchange will not be sufficient to quantify the export of anthropogenic carbon by coastal processes.

Major uncertainties

This report's synthesis of the current literature shows that the magnitudes of several significant components of coastal carbon budgets are currently uncertain.

Summary sentence or paragraph that integrates the above information

The synthesis in this chapter shows that coastal carbon budgets and anthropogenic contributions to the underlying fluxes are currently uncertain. Thus, more observations and modeling efforts could reduce these uncertainties.

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