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Surface Water and Groundwater Interactions in Salt Marshes and Their Impact on Plant Ecology and Coastal Biogeochemistry

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Abstract Salt marshes are highly productive intertidal wetlands providing important ecological services for maintaining coastal biodiversity, buffering against oceanic storms, and acting as efficient carbon sinks. However, about half of these wetlands have been lost globally due to human activities and climate change. Inundated periodically by tidal water, salt marshes are subjected to strong surface water and groundwater interactions, which affect marsh plant growth and biogeochemical exchange with coastal water. This paper reviews the state of knowledge and current approaches to quantifying marsh surface water and groundwater interactions with a focus on porewater flow and associated soil conditions in connection with plant zonation as well as carbon, nutrients, and greenhouse gas fluxes. Porewater flow and solute transport in salt marshes are primarily driven by tides with moderate regulation by rainfall, evapotranspiration and sea level rise. Tidal fluctuations play a key role in plant zonation through alteration of soil aeration and salt transport, and drive the export of significant fluxes of carbon and nutrients to coastal water. Despite recent progress, major knowledge gaps remain. Previous studies focused on flows in creek-perpendicular marsh sections and overlooked multi-scale 3D behaviors. Understanding of marsh ecological-hydrological links under combined influences of different forcing factors and boundary disturbances is lacking. Variations of surface water and groundwater temperatures affect porewater flow, soil conditions and biogeochemical exchanges, but the extent and underlying mechanisms remain unknown. We need to fill these knowledge gaps to advance understanding of salt marshes and thus enhance our ability to protect and restore them.

Plain Language Summary Salt marshes populated by salt-tolerant plants are commonly found at the land-ocean interface. As highly productive intertidal wetlands, salt marshes serve a variety of ecological functions. This review covers the current state of knowledge on surface water and groundwater interactions in salt marshes and their impacts on ecology and the environment. A global analysis is provided to understand the distribution, hydrological and ecological functions, status and potential challenges of salt marshes. Surface water and groundwater interactions under different driving forces and factors are discussed and linked to soil conditions and biogeochemical processes. Current approaches including field measurements, radionuclide analyses and numerical modeling provide a guide for applications targeted at specific questions in salt marsh science and management. Knowledge gaps identified by this review suggest needs for further investigations into the complex and dynamic surface water and groundwater interactions and associated biogeochemical processes that characterize salt marshes.

1. Introduction

Salt marshes are coastal ecosystems colonized by halophytes (i.e., salt-tolerant plants such as herbs, grasses, or low shrubs) and regularly flooded by tides (Adam, 1990; Allen & Pye, 1992; Chapman, 1960). Salt marshes

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are typically found in the upper intertidal zone (Adam, 2002; Chapman, 1960). They widely colonize low-slope and low-energy land-ocean interfaces with brackish or saline surface water (Balke et al., 2016). Salt marshes differ both physically and functionally from other coastal zones such as more inland freshwater wetlands, coastal tree-dominated mangroves, unvegetated tidal flats, and permanently inundated tidal zones where plants like sea-grasses are found.

Significant effort has been made to quantify the global distribution of salt marshes via remote sensing and field surveys (Campbell & Wang, 2020; McOwen et al., 2017; Y. Hu et al., 2021), but salt marsh areas are difficult to define due to active sedimentation, changes in land use, periodic tidal inundation and diverse vegetation. Reported areas of global salt marshes range from 22,000 to 400,000 km^2 (Chmura et al., 2003; C. M. Duarte et al., 2005). Based on recent data (McOwen et al., 2017), salt marshes are estimated to have a total global area of around 54,951 km^2 distributed across 43 countries and territories. They are generally found in estuaries, tidal inlets, embayments and open coasts where intertidal accommodation space is sufficient (Allen, 2000; Dijkema, 1987; French, 2019). Based on the polygon data set of McOwen et al. (2017), we calculated the area of salt marshes in per 10^4 km^2 and per 1° grid longitude or latitude cell (Figure 1). The analysis indicates that salt marshes are most extensive in North and Central America, Oceania and Northern Asia. About 55.1% and 13.4% of the world's salt marshes, respectively, lie in the northern temperate zone (23.5°–66.5°N) and the northern frigid zone (66.5°–73.6°N). Salt marshes are sparse in subtropical and tropical areas, which are dominated by mangroves (Chapman, 1976; Ellison, 2009). By country, the United States (18,849 km^2, 34.7%), Australia (13,259 km^2, 24.1%) and Russia (7,007 km^2, 12.8%) have the largest areas and host around 70% of the world's salt marshes (McOwen et al., 2017).

Located at the land-ocean interface as either shoreline ecosystems or near-shore islands, salt marshes are dynamic zones that are affected by both inland and oceanic forcing factors (Figure 2a). These interactions are driven by inland input (including river discharge and subsurface input of fresh groundwater), tides, waves, rainfall, evapo-transpiration, and long-term sea level change (Figure 2b), all of which are modified by climate change and human activities. Within salt marshes, there are strong and dynamic interactions between surface water and groundwater. These different factors influence sediment erosion and deposition, which are important drivers of morphological changes.

Multiple hydrological, ecological and geomorphological processes interact and affect salt marsh structures and functions (Figure 2b; Coco et al., 2013; Z. Zhou et al., 2016). Water flow transports sediment and solutes and crucially affects soil conditions. Sediment erosion and deposition are the main causes of morphologic change. Hydrodynamics, soil conditions and morphology cooperate to influence plant growth and distribution. Vegetation in turn influences soil conditions via transpiration and modification of soil properties including carbon accumulation. Water flow and sediment transport are also influenced by vegetation, which changes flow resistance and structure and traps suspended sediments (Fagherazzi et al., 2012; Mudd et al., 2010). Sediment properties are affected by plant roots and litter, feeding back into geomorphological processes as organic sediments (Neubauer, 2008).

The inherent linkages among processes affecting salt marshes regulate biogeochemical cycles and underpin the ecology of salt marshes, particularly outwelling and plant zonation. Outwelling refers to carbon and nutrients exported to the coastal sea that supports the productivity of near-shore marine ecosystems (Mitsch & Gosselink, 2000; Odum, 1968; Teal, 1962) or long term carbon sequestration in the ocean (Santos, Burdige, et al., 2021). Plant zonation is the organization of species into distinct spatial patterns ("zones") along a key gradient (A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015; Chapman, 1960; Marani et al., 2006; Pennings & Callaway, 1992; Silvestri et al., 2005). Understanding of surface water-groundwater (SW-GW) interactions is key to resolving these processes. More generally, complex interactions will drive how salt marshes respond to climate change, sea level rise and human activity, and influence strategies for mitigation of the associated impacts.

Salt marshes support a variety of important ecosystem functions:

1. They are highly productive, with net primary production of 200–4,000 g C m^{-2} y^{-1} (Curcó et al., 2002; Ibñez et al., 2002; Morris, Sundberg, & Hopkinson, 2013; Neves et al., 2007; Roman & Daiber, 1984). This is comparable to the productivity of the Amazon rainforest (1,000–1,600 g C m^{-2} y^{-1}; Araujo-Murakami et al., 2014; Girardin et al., 2016; Malhi et al., 2009).
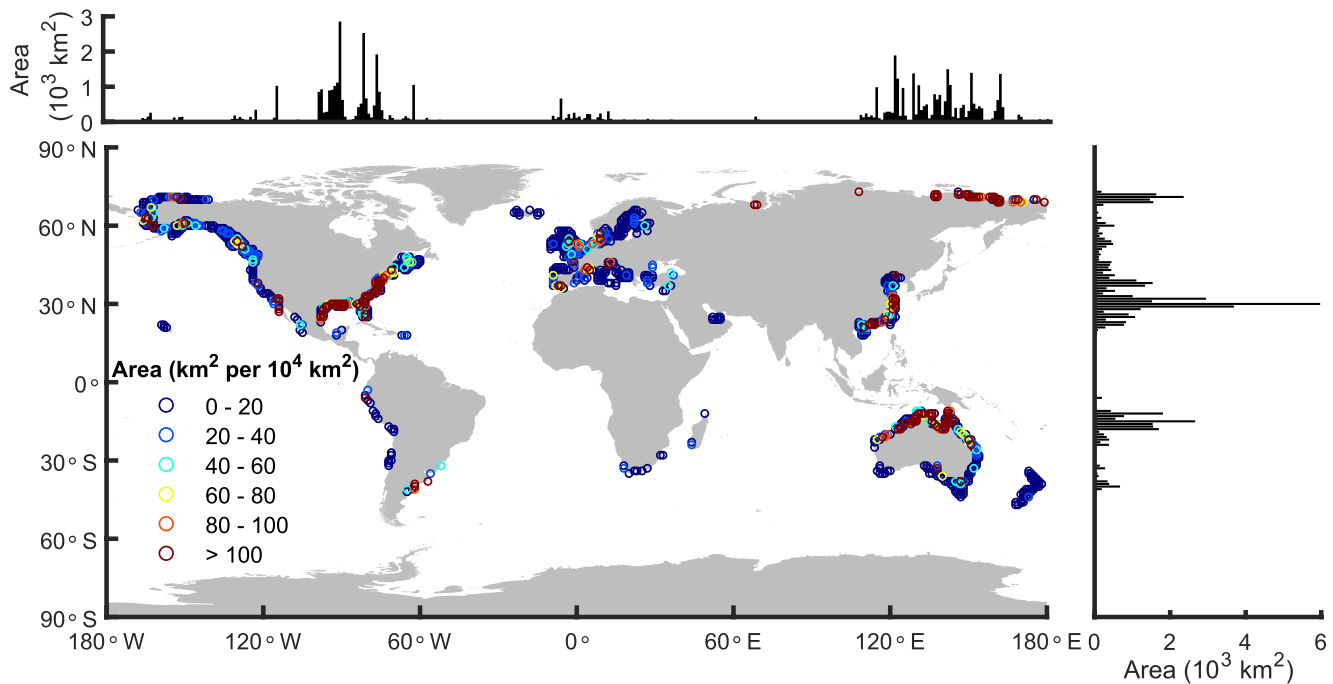
Figure 1. Global distribution of salt marshes based on data from McOwen et al. (2017). The latitudinal and longitudinal areal extent of salt marshes per 1° grid cell are shown in above and right subplots, respectively. The original polygon data was converted to binary (unvegetated or vegetated) data to 0.0048° grid cells. Then, the sum of vegetated grid areas in targeted zones were calculated.

2. They provide essential habitats for many organisms, including mammals, birds, fish and invertebrates, and are called a "gene pool" of diverse species (Barbier et al., 2011). In particular, they provide important feeding and stopover habitats for migratory birds along international flyways (Maslo & Lockwood, 2013).

3. They provide natural buffering against coastal hazards such as storms (Temmerman et al., 2013) because vegetation greatly increases friction coefficients, which reduce current flow and dissipates wave energy. In this way, salt marshes contribute to buffering storm surges and protect shorelines against erosion (Gedan et al., 2011; Shepard et al., 2011).

4. At the interface between land and ocean, salt marshes serve as natural filters for the removal of nutrients, metals and other pollutants (Breaux et al., 1995; Bystrom, 2000; Verhoeven et al., 2006). In this way, salt marshes are referred to as a "kidney" of the coast (Mitsch & Gosselink, 2000).

5. They are regarded as an important blue-carbon ecosystem, sequestering greenhouse gases and partially offsetting climate change (Lovelock & Duarte, 2019). About 430 Tg C are stored in the upper 0.5 m of soils in salt marshes globally (Chmura et al., 2003). The average global rate of carbon sequestration in salt marsh ecosystems is estimated to be 168 g C m^{-2} y^{-1}, leading to a total global rate of 48–59 Tg C y^{-1}, which is equivalent to 30% of the organic carbon buried in the entire seafloor (F. Wang et al., 2020).

Despite the enormous ecological value, more than 50% of the world's original salt marshes were lost in the last century (Barbier et al., 2008) at an annual rate of approximately 1%–2% (C. M. Duarte et al., 2008). Many existing salt marshes around the world suffer from serious degradation and loss of essential ecosystem functions:

1. More than 40% of the world's population currently lives within 100 km of the coastline (Martinez et al., 2007). By 2060, coastal populations are predicted to increase by 120% relative to the year 2000 (Neumann et al., 2015). To support such a dense population, salt marshes will likely continue to be transformed into urban, residential, and industrial land (Gedan et al., 2009). For example, at least 7,080 km^2 of salt marshes in China were "reclaimed" from the 1950s to the early 1990s (S. Yang & Chen, 1995). In Boston, 70% of the original salt marsh has been transformed to urban land (Bromberg & Bertness, 2005).
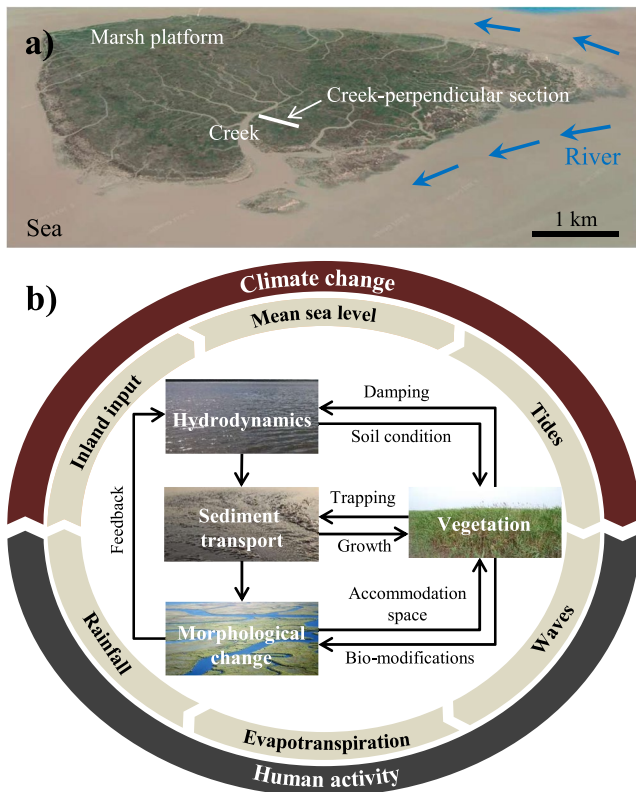
Figure 2. (a) A photo of a recently formed salt marsh (Jiuduansha Wetland, 31.20°–31.25°N and 121.87°–121.93°E) in the Yangtze River estuary, China (Google Earth, 9 April 2021). This marsh was created in 60 years from deposition of sediment carried by the Yangtze River. (b) A typical bio-morphodynamic loop considering biophysical processes in salt marshes (modified from Z. Zhou et al. (2016)). The biophysical processes affect chemical transformations and exchange between marsh and sea.

2. Salt marshes are threatened by changes in global mean sea level, which has risen by about 0.2 m in the last century (Church & White, 2011; Jevrejeva et al., 2008). Global sea level will rise between 0.43 (RCP, Representative Concentration Pathway 2.6) and 0.84 m (RCP 8.5) by 2100 relative to 1986–2005 (Oppenheimer et al., 2019). If their vertical accretion rates cannot keep pace with sea level rise, salt marshes will be submerged and lost permanently (Kirwan & Murray, 2007; Nicholls & Cazenave, 2010).

3. Both vertical accretion and lateral expansion of salt marshes need sufficient sediment supply (Kirwan et al., 2016). Riverine supply is a major sediment source, however rapid decreases in the sediment flux from rivers to oceans is a global issue. Syvitski et al. (2005) estimated that the flux of sediment reaching the world's coast is declining at a rate of 1.4 Pg (billion tonnes) per year due to human influence. For example, the annual sediment load of Yangtze River decreased from 0.51 to 0.13 Pg from 1954–1968 to 2004–2015 (Wu et al., 2020). That of Mississippi River reduced from 0.4 to 0.18 Pg mainly due to dam construction (Blum & Roberts, 2012). Declines in sediment supply may cause a salt marsh to shift from expanding to shrinking (Schuerch et al., 2018) and may drive entire regional marsh systems from high- to low-marsh states in the future (Moffett & Gorelick, 2016a).

4. Nutrients were previously deficient in coastal sediments, limiting growth of salt marsh plants. However, fertilizer use now causes excessive nutrient inputs (Galloway et al., 2008) and eutrophication of coastal areas (B. Wang et al., 2018; Canfield et al., 2010). Eutrophication can induce functional and structural changes in salt marshes, including changes to primary production, carbon sequestration and species composition (Graham & Mendelssohn, 2014; Morris, Shaffer, & Nyman, 2013). Eutrophication decreases belowground biomass (e.g., roots and rhizomes), which are important for stabilizing salt marshes (Deegan et al., 2012; Wasson et al., 2017). This may contribute to vegetation retreat and creek bank erosion, resulting in salt marsh loss. In this way, salt marshes may transform from carbon sinks into carbon sources.

5. Salt marshes are vulnerable to the invasion of alien species (Strong & Ayres, 2013). Recently, invasion of *Spartina alterniflora* has attracted global attention. This cordgrass species is native to the East Coast of North America and the Gulf of Mexico (Kirwan et al., 2009) but is now found in estuaries, lagoons, and coastal intertidal zones worldwide. For example, *S. alterniflora* was responsible for one of the largest invasions in the USA at Willapa Bay (Washington). Its cover grew to 270 km^2 of intertidal and supratidal zones during the twentieth century (Strong & Ayres, 2016). *S. alterniflora* was initially introduced to China in 1979 (An et al., 2007) and now covers 616 km^2, which is half of the total Chinese marsh area (1,275 km^2; Y. Hu et al., 2021). Because of its high tolerance to salinity and inundation (B. Li et al., 2009; Strong & Ayres, 2016), *S. alterniflora* spreads rapidly and significantly alters ecological structures and functions of native ecosystems.

Given this status, it is urgent to develop and implement effective approaches to protect existing salt marshes and restore lost and declining marshes. The objective of this paper is to provide a comprehensive review of SW-GW interactions and associated biogeochemical processes and plant ecology in salt marshes. SW-GW interactions have been an active research topic in recent decades. Available reviews focus on inland hyporheic zones such as streambeds, river banks and floodplain areas (Boano et al., 2014; Cardenas, 2015). SW-GW interactions with a focus on seawater intrusion and submarine groundwater discharge (SGD; two complementary processes in coastal aquifers) were reviewed by Werner et al. (2013), Robinson et al. (2018), Taniguchi et al. (2019), Santos et al. (2012), W. S. Moore and Joye (2021), Alorda-Kleinglass et al. (2021), Santos, Chen, et al. (2021), and Garcia-Orellana et al. (2021). Recently, Guimond and Tamborski (2021) reviewed hydrological processes in salt marshes. They focused on groundwater flow and SGD and not plant ecology and coastal biogeochemistry.
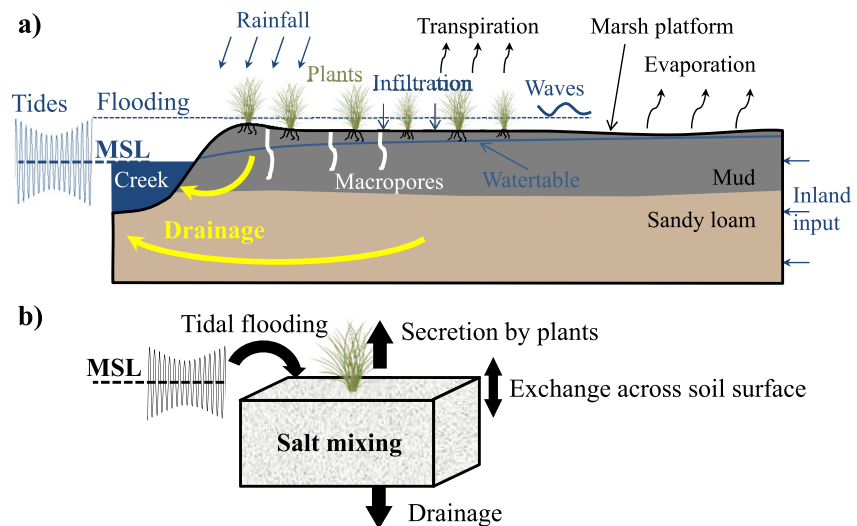
Figure 3. (a) Conceptual diagram of the creek-perpendicular marsh section showing net groundwater flow patterns with key hydrological factors and representative stratigraphy. (b) Conceptual model of salt balance in the marsh soil (modified from Morris (1995)). MSL means mean sea level.

In this review, we build on earlier work by going deeper into the driving forces of SW-GW interactions and how these essential hydrological processes drive plant ecology and biogeochemical exchange between salt marshes and the sea. We focus on bi-directional interactions between surface water and groundwater mainly occurring in shallow sediments of salt marshes. It should be noted that accommodation space of salt marshes is dynamic and affected by geomorphological processes (e.g., marsh accretion/erosion, creek evolution, and sediment subsidence). Geomorphological processes likely happen at long-term and interannual scales and have been reviewed by Allen (2000), Coco et al. (2013), Fagherazzi et al. (2012), and FitzGerald and Hughes (2019).

We first review the state of knowledge of SW-GW interactions in salt marshes with a focus on porewater flow and soil conditions (soil saturation, groundwater salinity, and oxygen availability) as driven by different hydrological forcings and modified by eco-geomorphological factors (Section 2). We outline current approaches used to examine SW-GW interactions at various spatio-temporal scales, including field measurements, radionuclide analyses and numerical models (Section 3). With a link to SW-GW interactions, we synthesize an understanding of hydrology-dependent plant growth and species competition/facilitation in salt marshes, which underpin plant zonation and thus marsh ecology (Section 4). We then discuss how SW-GW interactions affect biogeochemical transformations and to what extent salt marshes export carbon, greenhouse gases and nutrients to the adjacent sea (Section 5). Knowledge gaps are identified to guide future research in Section 6 before drawing conclusions in Section 7.

2. Surface Water and Groundwater Interactions

SW-GW interactions significantly modify soil saturation (Armstrong et al., 1985; H. Li et al., 2005; Howes & Goehringer, 1994; L. R. Gardner, 2005a), groundwater salinity (Carol et al., 2012; Moffett, Robinson, & Gorelick, 2010; Pennings et al., 2005), and biogeochemical cycling (A. M. Wilson & Gardner, 2006; Hemond et al., 1984; Montalto et al., 2007; Nuttle & Hemond, 1988; Valiela et al., 1978) in salt marshes (Figure 3a). Salt marshes are dissected by tidal creeks, which are an essential morphological component and behave as exchange pathways for marsh water, sediments and chemicals with the adjacent sea (Figure 2a). Therefore, studies of SW-GW interactions mainly focused on these creeks. In some cases, this creek-centered focus is incomplete. For example, in some cases marsh surface exchange via non-channelized sheet flow are key salt or biogeochemical pathways (H. Zhang, Moffett, et al., 2014; Moffett, Robinson, & Gorelick, 2010).

The conceptual model in Figure 3a depicts net groundwater flow through a marsh platform that slopes gently toward the creek, with a natural levee at the creekbank. Note that marshes in back-barrier systems can display a concave-up profile moving from the creek edges toward the interior marsh (Beeftink, 1966; Beeftink &

Rozema, 1988). Although the sediments below marshes are usually homogeneous, the conceptual model shows fine-grained sediments overlying a more permeable sandy-loam layer because this structure is common in many marshes (C. E. Hughes et al., 1998; J. W. Harvey et al., 1987; L. R. Gardner & Porter, 2001; Wiegert & Freeman, 1990; Xin et al., 2009). Due to the macropores created by plant root and organism burrowing, the permeability of the upper marsh sediments can be much higher than muds (C. E. Hughes et al., 1998; Guimond, Seyfferth, et al., 2020). Groundwater flow through this system is driven by tidal fluctuations and infiltration of rainwater into the upland and the marsh surface. Net groundwater flow is toward the creek on average, with flow concentrated in more permeable sediment layers. As discussed later, saline groundwater that discharges to tidal creeks is enriched in nutrients and other solutes, and therefore provides a source of nutrients to surface water.

At the much smaller scale of the root zone (Figure 3b), dynamic soil conditions strongly influence plant growth and zonation (Armstrong et al., 1985; Pennings et al., 2005; Pezeshki, 2001). The dynamic soil conditions are affected by interacting processes. For example, infiltration of tidal creek water provides salt to the marsh soils, and the resulting brackish-to-saline groundwater is then concentrated or diluted by evapotranspiration or rainfall-induced freshwater infiltration. Salt in the marsh soil is also removed by lateral drainage, exchange across the marsh surface and removal by plants (some marsh plants such as *S. alterniflora* take up saline porewater and excrete salt into the environment; Morris, 1995). Higher soil saturations result in poor soil aeration. As the oxygen diffusivity and concentration in groundwater are much lower than those in the soil air, poor aeration can lead to low oxygen availability in the soil, which inhibits plant growth (Armstrong & Drew, 2002; Mendelssohn et al., 1981).

Although it is clear from this conceptual model (Figure 3a) that multiple processes interact to control the hydrology of these marsh systems, prior studies typically examined only one aspect of the flow system at a time. Therefore, we first describe each aspect in detail and then describe the interactions of multiple factors and their net impact on groundwater flow and soil conditions.

2.1. Tides

Tides are an essential part of salt marsh functioning. The baseline tidal elevation is further influenced by wind, as illustrated by the example of storm surges that can significantly raise sea level (A. M. Wilson et al., 2011). Tides propagate through tidal creeks and flood marsh platforms, and so are a critical driver for movement of water and solutes in marsh systems. Tides then ebb off the marsh surface both by direct drainage to tidal creeks and by sheet flow over the marsh surface. Tidal fluctuations have multiple harmonic constituents, for example, solar semi-diurnal (S2), principle lunar semidiurnal (M2), solar diurnal (S1), and lunar diurnal (O1) (Godin, 1972). For simplicity, prior studies, and particularly modeling studies, considered limited tidal constituents, for example, monochromatic semidiurnal tides and spring-neap tides with only two harmonic constituents (see Section 3.3). In early studies, this led to simplified binary considerations of inundating and non-inundating tides, depending on whether the high tide level is higher than the marsh platform elevation (A. M. Wilson & Gardner, 2006; Montalto et al., 2006).

The relationship between the elevations of the marsh surface and local surface water fluctuations determines the frequency and duration of tidal inundation (Montalto & Steenhuis, 2004). Salt marshes usually slope gently toward the sea or may display a concave-up profile from creeks to interior areas in back-barrier systems. High marsh areas are subjected to less frequent flooding events and shorter inundation periods (J. W. Harvey et al., 1987). In general, tide-induced watertable fluctuations are most evident near creeks and are attenuated due to damping (by friction and water spreading) as the tide propagates toward the marsh interior away from creeks (Jacob, 1950). As the slope of the water surface over marsh platforms is often on the order of 10^{-4} (Rinaldo et al., 1999), the associated hydraulic gradient (without waves) has a negligible effect on subsurface flow.

During rising and high inundating tides, surface water overtops creekbanks, flows over the marsh platform and infiltrates through the marsh sediment surface (A. M. Wilson & Gardner, 2006; Howes & Goehringer, 1994; J. W. Harvey et al., 1987; Montalto et al., 2006). Tidal water also enters through the creek bank during the rising tide (Figure 4), although the bulk of the recharge occurs through the marsh platform (A. M. Wilson & Gardner, 2006; Howes & Goehringer, 1994; L. R. Gardner, 2005a). Vertical water infiltration is rapid during inundation, so the marsh can become fully saturated at the high tide, with the watertable coupled to the surface water level.

When the marsh platform is gradually exposed during falling tide, groundwater seeps out from the marsh sediments across the creek bank and bottom. This drainage lowers the watertable and leads to the desaturation/reaer-
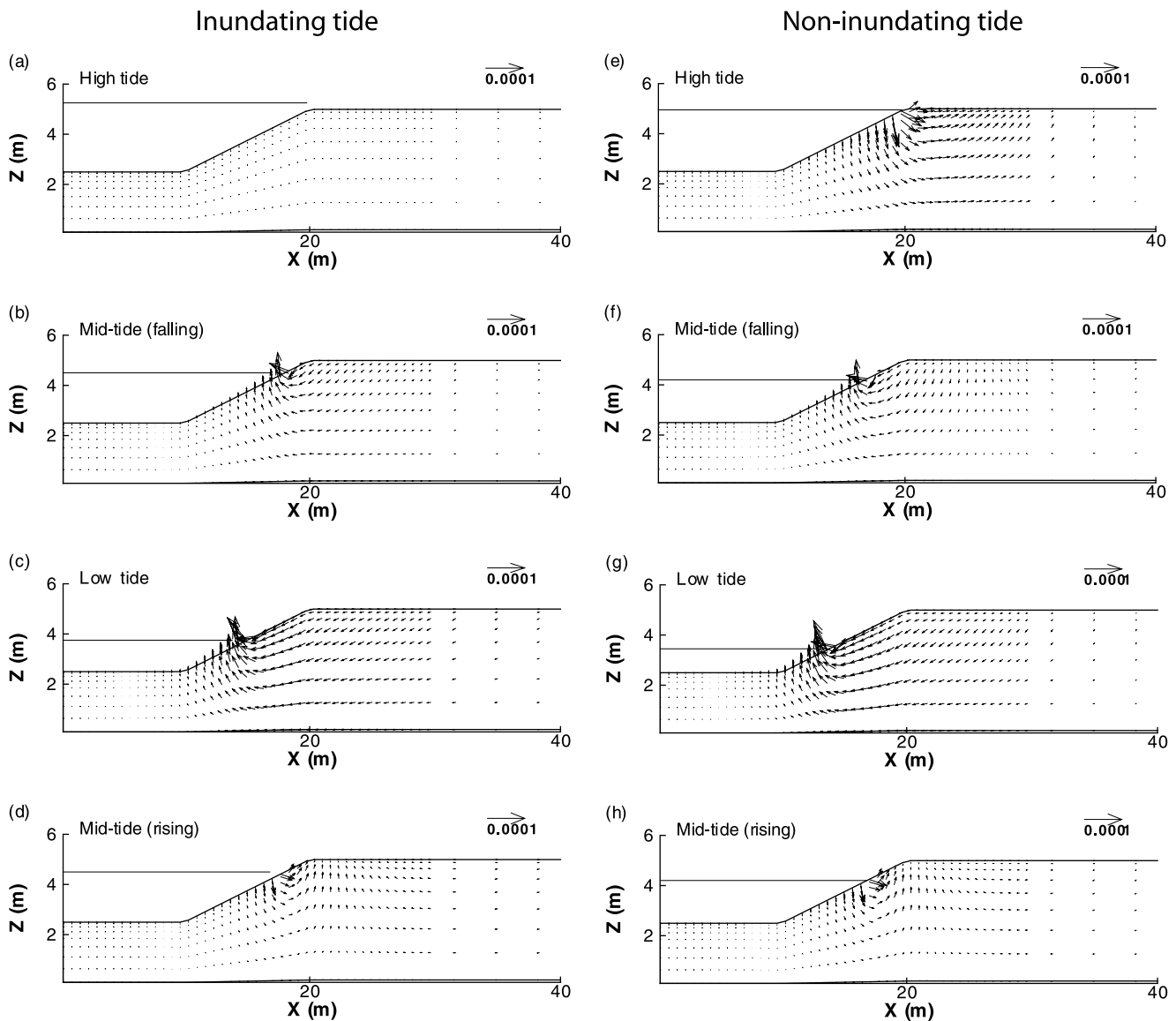
Figure 4. Flow velocities (in m/s) in an example vertical marsh section perpendicular to a creek and subjected to tidal fluctuations. The modeled marsh is assumed to be homogeneous. The horizontal lines show the tidal levels. (a and e) High tide; (b and f) mid-tide, ebb; (c and g) low tide; and (d and h) mid-tide, rising. The left panel is for inundating tide and the right panel is for non-inundating tide (adopted from A. M. Wilson and Gardner (2006)). X is the distance to the creek center and Z is the elevation.

ation of the soil (Cao et al., 2012; H. Li et al., 2005; Marani et al., 2006; Ursino et al., 2004), thus providing a mechanism for solute exchange near the creek bank, as well as preventing a buildup of salt or other solutes (A. M. Wilson & Gardner, 2006; L. R. Gardner, 2005a). In contrast, drainage diminishes with increasing distance from the creek (A. L. H. Hughes et al., 2012; Howes & Goehringer, 1994; L. R. Gardner, 2005a; Nuttle, 1988a) where the average watertable height and soil saturation increase (Marani et al., 2006; Silvestri et al., 2005; Xin, Kong, et al., 2013; Xin et al., 2011). Increases in tidal amplitude enhance groundwater flow. Indeed, water and nutrient exchange are larger in macrotidal than microtidal marshes (A. M. Wilson, Evans, Moore, Schutte, & Joye, 2015; A. M. Wilson & Morris, 2012).

Contrasts between the salinity of surface water and marsh porewater can further affect groundwater flow and solute transport (Shen et al., 2015, 2016; Xiao, Li, et al., 2019). Marsh platforms can have an upward density gradient (surface water salinity is higher than porewater salinity) or downward density gradient. As indicated by numerical simulations from Shen et al. (2015), upward salinity gradients form salt fingers (unstable groundwater flow in form

of fingers), which can significantly modify local groundwater flow and solute transport. However, due to the dominant role of tidally driven near-creek circulation in salt marshes, particularly in those with high-permeability, the impact of salt fingers is limited to the marsh interior. Even with a stable, downward density gradient, substantial salt exchange can occur diffusively across the marsh surface. Field investigation (Moffett, Robinson, & Gorelick, 2010) and numerical simulations (Xiao, Li, et al., 2019) indicated that SW-GW exchange is enhanced when surface water has variable salinity.

During periods of non-inundating tides, large areas of the marsh platform remain exposed, and the effects of tidal forcing on unflooded areas are limited compared to when inundation occurs (A. M. Wilson & Gardner, 2006; L. R. Gardner et al., 2002). In the absence of inundation, the watertable is nearly flat across the majority of the marsh interior except for fluctuations near creeks, if one assumes spatially homogeneous sediment and evapotranspiration properties (Montalto et al., 2006). If sediment hydraulic properties or extraction of water from the root zone by evapotranspiration vary spatially, then even in a 2D transect the marsh watertable and pressure head distributions would not proceed uniformly across the marsh, but would necessarily respond to sharper gradients at sediment or vegetation zone boundaries (Moffett et al., 2012). Subjected to non-inundating tides, tidally driven groundwater circulation mainly occurs in a narrow near-creek zone, with influx and efflux focused, respectively, around the high and low tide limits if bank sediments are uniform (A. M. Wilson & Gardner, 2006; L. R. Gardner, 2005a; Figure 4), or else focused on a higher-permeability stratigraphic unit, if present (L. R. Gardner, 2007). Therefore, the tide-induced water exchange in marshes is weaker during non-inundating tides than during inundating tides (A. M. Wilson & Gardner, 2006; A. M. Wilson & Morris, 2012). At higher elevations away from creeks, there is a thicker unsaturated soil layer with improved aeration (A. M. Wilson & Morris, 2012).

Spring-neap tidal variations are generated by bichromatic signals, that is, a semi-diurnal solar tide (12 hr period) and semi-diurnal lunar tide (12.42 hr period), forming a spring-neap cycle with a period of 14.78 days (Godin, 1972). Over a spring-neap tidal cycle, the variable elevation of the high tide affects the extent of inundation. In practice, inundation is frequent during spring tides. During neap tides, the marsh platform remains exposed for longer time. During spring tides, temporal variations of the local watertable and soil saturation are essentially similar to those for monochromic inundating tides (Xin et al., 2010). As the tidal regime changes from spring to neap, the marsh platform is subjected to a long-period emergence that allows the marsh interior to drain. This causes a continuous reduction in soil saturation (Nuttle, 1988a; Xin, Li, & Barry, 2013). Groundwater discharge can increase by a factor of two during spring tides relative to neap tides (A. M. Wilson, Evans, Moore, Schutte, & Joye, 2015; A. M. Wilson et al., 2011). If spatially homogeneous sediments and uniform evapotranspiration are assumed, one can model a relatively well-aerated near-creek zone (induced by spring tides) and a poorly aerated interior zone (induced by neap tides) over a spring-neap tidal cycle. The extents of the two zones mainly depend on the amplitudes of bichromatic tidal signals and permeability of the marsh sediment (Xin et al., 2010). If sediments and evapotranspiration rates vary spatially, even along 2D transects, then some interior zones may be more well-aerated than others (H. Zhang, Moffett, et al., 2014; Moffett et al., 2012).

Most of the salt marsh literature has relied on 2D vertical marsh transects perpendicular to a tidal creek, with spatially homogeneous sediments and spatially uniform boundary conditions on each edge (A. M. Wilson & Gardner, 2006; H. Li et al., 2005; L. R. Gardner, 2005a; Marani et al., 2006; Ursino et al., 2004; Xin et al., 2011). These assumptions neglect the system's spatial variability and omit flow in the along-creek or cross-surface (sheet flow) directions. Furthermore, most studies assume that the surface water level is flat and without feedback from groundwater. These assumptions are useful to compare contrasting scenarios of marsh conditions at a scale much coarser than spatial variations in porous media, surface roughness and boundary conditions (e.g., evapotranspiration), and at a scale much finer than surface water slope and surface water salinity variations. In contrast, some 3D modeling studies have considered coupled SW-GW interactions (e.g., Guimond, Yu, et al., 2020; Moffett et al., 2012; Xin, Li, & Barry, 2013; Xin et al., 2011, 2012; Y. Zhang et al., 2018, details in Section 3.3). Xin et al. (2011) modeled a simple homogeneous 3D marsh with a seaward sloping marsh platform, a central meandering tidal creek and a homogeneous surface boundary condition. A monochromatic and inundating tide induced three key circulations processes with time scales differing by orders of magnitude (Figure 5c): near-creek circulation, meander-modulated circulation and near-channel circulation (Figures 5a and 5b). In contrast, with a relatively straight creek, 2D behavior dominates and tidally driven groundwater circulation is mainly perpendicular to the creek, as revealed by 2D modeling studies (A. M. Wilson & Gardner, 2006; A. M. Wilson & Morris, 2012; Marani et al., 2006; Ursino et al., 2004). Superimposed on this conceptual template, whether 2D or
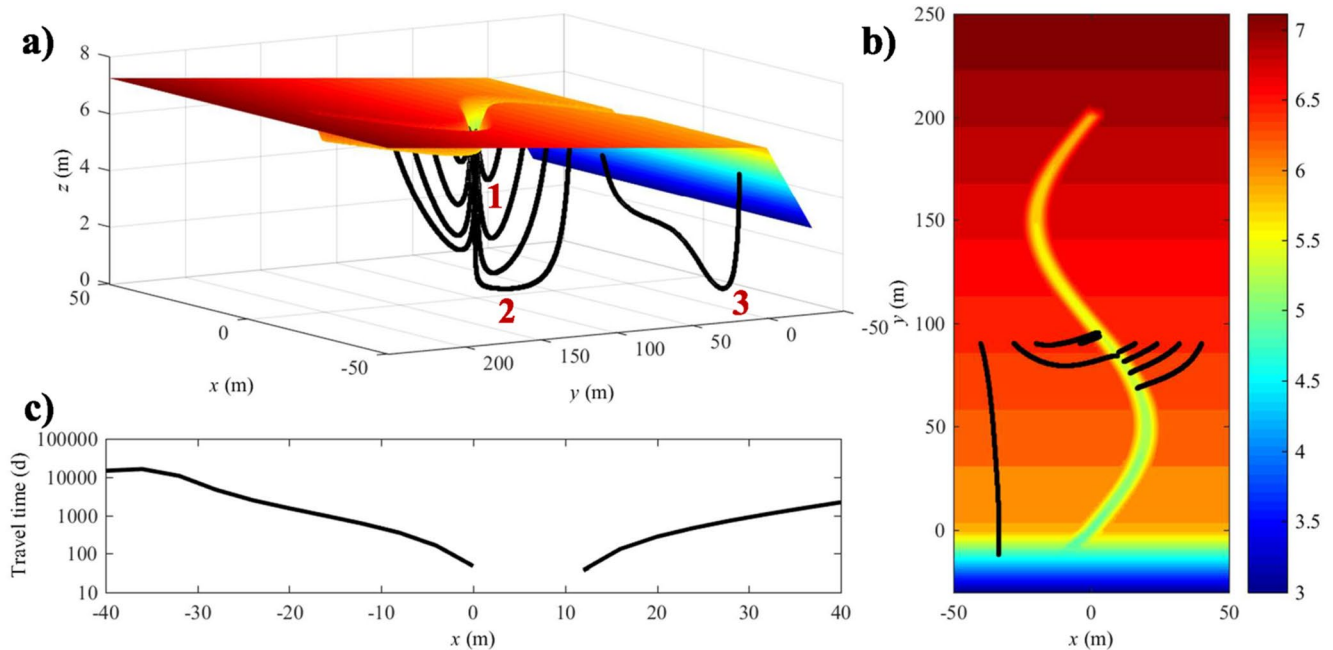
Figure 5. Particle tracking in a simplified homogeneous 3D marsh with a meandering tidal creek and a main channel. The horizontal Cartesian coordinates are x (along the main channel direction) and y (perpendicular to the main channel), while z is the vertical coordinate. (a) 3D particle traces and (b) plan view. The colored contour map represents the marsh topography (sediment surface elevation). The particles are released at y (the distance to the main channel) = 90 m (0.5 m soil depth). (c) Travel times that particles take to move through the marsh sediment: The x-axis shows where the particle is released initially. The tidally averaged flow leads to topographically affected 3D porewater circulation: near the creek bank (1), around the creek meander (2), and over long marsh sections inclined toward the main channel (3) (adapted from Xin et al. (2011)).

3D, is additional variation in groundwater circulation caused by both variable sediment permeability (both matrix and macro-porosity) and porewater extraction by evapotranspiration.

Salt marsh sediments often exhibit heterogeneity. Moffett et al. (2012) simulated 3D homogeneous and heterogeneous marsh systems based on field measurements and still found that groundwater discharge occurs mainly via the tidal channel network, driven by spring and neap tides. The model and observations showed that a flooding tide masks hydraulic differences in marsh soils (Moffett et al., 2012). Spatial heterogeneity of the sediment permeability (according to sediment zones) and additional spatial variability in porewater extraction by evapotranspiration (according to vegetation zones) increased spatial hydraulic differences and led to a mosaic of compound ecohydrological zones that exhibited distinct soil saturation and groundwater velocity dynamics contrast to homogeneous marsh models. Recently, Guimond, Yu, et al. (2020) established a 3D marsh model consisting of an estuary and the surrounding coastal marsh and forest, including soil hydraulic parameters. The simulation results demonstrated five hydrologic zones related to marsh topography and spring-neap tides: subtidal zone, tidal near-creek zone, spring-neap zone, tidal interior zone, and upland zone.

Storms affect water flow in salt marshes through wind-driven storm surges. A. M. Wilson et al. (2011) found that elevated tides associated with a tropical storm reduced groundwater discharge from a salt marsh by roughly a factor of two compared to the periods before and after the surge. During the same period, high tides drove large volumes of surface water from the creek into salt marsh sediments through the creek bank and geochemical evidence showed changes more than 25 m from the creek bank (A. M. Wilson et al., 2011).

2.2. Waves

Waves can be another important factor in the intertidal zone, but they are rarely examined in SW-GW studies of salt marshes because high-frequency swell waves are sequentially attenuated in estuaries, low-lying tidal flats often existing in front of salt marshes, and finally marsh plants. Waves propagating into the shallow water zone will break (Longuet-Higgins, 1983; Nielsen, 2009). Thus, waves are only expected to induce SW-GW exchange across marsh sediments during inundating tides and when large areas are inundated. Over a flat seabed, swell

waves drive surface water into the benthic zone under a high hydraulic head, and groundwater seeps out beneath under a low hydraulic head (Russoniello et al., 2018; Sawyer et al., 2013). This wave induced benthic exchange likely increases with soil permeability and water surface slope caused by waves (King et al., 2009). Marine benthic studies indicate that water transport is dominated by bioturbation or diffusion in sediments with permeabilities below 10^{-12} m^2 (Huettel et al., 2014). Thus, wave-driven flow in the subsurface is likely to be limited in all but very sandy marsh systems, although it may be important in macropore fluid turnover. This wave-driven benthic exchange is expected to be altered by marsh plants and modified by sediment geometry and heterogeneity. Marsh plants could attenuate wave energy and modify hydraulic heads across the marsh sediment surface (Rupprecht et al., 2017; Schoutens et al., 2019; Ysebaert et al., 2011). Waves would affect marsh geomorphology by regulating sediment deposition/erosion and creek evolution which in turn affect SW-GW interactions (Leonardi et al., 2016; Marani et al., 2011; Mariotti & Fagherazzi, 2013). However, these processes and mechanisms remain unexamined in salt marshes to date.

2.3. Evapotranspiration

As an integral part of the hydrologic cycle, evapotranspiration is an important driver for groundwater flow and associated solute transport in salt marshes (Dacey & Howes, 1984; H. Li et al., 2005; Marani et al., 2006; Moffett, Robinson, & Gorelick, 2010; Morris, 1995; Silvestri et al., 2005). Observed evapotranspiration rates in natural marshes typically range from 4 to 15.6 mm d^{-1} (Marani et al., 2006; Stannard et al., 2013), which is generally larger than the precipitation rate in subpolar and temperate zones where salt marshes are distributed (Adler et al., 2018). Both soil evaporation and plant transpiration depend on soil moisture in the shallow soil layer (related to capillarity and watertable depth; Or et al., 2013; Rose et al., 2005; Shimojima et al., 1990), with transpiration being further controlled by plant physiology (Mendelssohn et al., 1981; Pezeshki, 2001; Ungar, 1991). With a shallow watertable, a hydraulic connection is established between the evaporating soil surface and groundwater. The soil water loss caused by evaporation near the surface can be quickly replenished by capillary transport such that the evaporation rate is close to the potential rate (i.e., that for open water and controlled by atmospheric demand; Nachshon, Shahraeeni, et al., 2011). In contrast, the evaporation rate for a deep watertable is controlled by soil moisture rather than the atmospheric demand (Nachshon, Weisbrod, et al., 2011; Rose et al., 2005). In this way, the local evaporation rate changes periodically with tidal fluctuations, which regulate watertable and soil moisture condition (Hemond & Fifield, 1982; Shen et al., 2018). Evapotranspiration rates can also vary substantially according to overall available (radiative) energy, atmospheric humidity, and boundary layer aerodynamics (wind speed, vegetation height, and effective aerodynamic roughness), and more locally at the plant-to-patch scale (Moffett & Gorelick, 2012; Moffett, Wolf, et al., 2010). In vegetated systems, these factors are as significant as watertable depth in governing evapotranspiration rates. However, they are typically omitted from salt marsh hydrological studies and comparisons of biological versus physical controls on marsh evapotranspiration has yet to be explored.

Soil evaporation removes freshwater but leaves salt behind. In shallow marsh soils, the porewater salinity increases and the soil saturation decreases (Nuttle & Hemond, 1988), with the result that porewater salinity exceeds seawater salinity (A. M. Wilson et al., 2011; C. E. Hughes et al., 1998; Silvestri et al., 2005). When continuous evaporation increases the porewater salinity beyond solubility, salt precipitation can occur in the shallow soil layer. This would in turn affect the water transport in the soil and hence the evaporation process (C. Zhang, Li, & Lockington, 2014).

In addition to soil evaporation, plant transpiration affects local watertables and soil aeration. This leads to a positive feedback: pioneer plants initially grow successfully near the creek where soil aeration is better. The increase in the local evapotranspiration in turn improves soil aeration for further development of other plant communities (Dacey & Howes, 1984; H. Li et al., 2005; Marani et al., 2006; Ursino et al., 2004; Xin, Kong, et al., 2013). Based on numerical simulations that set plant respiration at the marsh surface (Ursino et al., 2004) or within the plant rhizosphere (H. Li et al., 2005; Marani et al., 2006), it is posited that enhanced evapotranspiration induces a permanently aerated zone below the soil surface in low-permeability salt marshes. Boaga et al. (2014) conducted an in situ test that suggested that low soil permeability and high evapotranspiration favor development of such a zone. Simulations by H. Zhang, Moffett, et al. (2014) showed that in higher-permeability sediments vertical infiltration could fill the aerated area during the flood tide, but an unsaturated "pocket" could persist in sediments of lower vertical permeability. Therefore, there should be key governing ratios of vertical permeability, tidal level

and duration and aeration depth that indicate the potential for a permanently aerated zone, although these have yet to be elucidated. If a permanently aerated zone exists during tidal inundation, it could theoretically remain aerobic, facilitating plant root respiration and growth. The main argument against this hypothesis is that macropores are common in shallow salt marsh sediments, enhancing sediment permeability (Guimond, Seyfferth, et al., 2020) and water infiltration during tidal inundation (Xiao, Wilson, et al., 2019; Xin et al., 2009). However, empirical geophysical (saturation; Boaga et al., 2014), geochemical (methyl mercury gas; H. Zhang, Moffett, et al., 2014), and hydrological (manometers; Chapman, 1938) evidence point to the persistence of some unsaturated pockets in the marsh subsurface during flood tides.

The upward flow induced by evapotranspiration is a dominant feature in the marsh interior (Dacey & Howes, 1984; Hemond & Fifield, 1982; Moffett et al., 2012; Xin, Kong, et al., 2013). This upward flow is augmented by hydraulic gradients between the upland area and marsh platform, which supports more rapid evapotranspiration (A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015). However, not all marshes have adjacent uplands with strong hydraulic gradients, nor do all marshes have high evapotranspiration demand. Due to greater water supply, and often greater plant leaf areas, evapotranspiration rates are typically higher for the intertidal zone than the upper zone not reached by tides (C. E. Hughes et al., 1998). Higher evapotranspiration rates over longer durations without tidal flooding or substantial rainfall result in a comparatively lower watertable and desaturation in deep sediments unless otherwise resupplied by upward groundwater flow (A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015) or higher porewater salinity (C. Zhang, Li, & Lockington, 2014; Montalto et al., 2006).

2.4. Freshwater Recharge

Freshwater recharge in the form of regional groundwater input or rainfall-induced infiltration affect groundwater and solute transport in salt marshes (A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015; Nuttle & Harvey, 1995). The overall freshwater input is determined by meteorological conditions, principally precipitation (Boorman, 2019). Precipitation causes freshwater infiltration into marsh soils if the marsh is not inundated by the tide. The rate of infiltration depends on rainfall intensity and the infiltration capacity of the soil. Once the rainfall (throughfall) rate exceeds the infiltration capacity or the soil is fully saturated, overland runoff occurs (L. R. Gardner & Reeves, 2002; Xin et al., 2017). For example, the salinity of marsh creek water can decrease dramatically during rainfall events, with no obvious changes of groundwater salinity afterward (Cao et al., 2012; Carter et al., 2008; Xiao et al., 2017). Remarkable rainfall infiltration was also noted during a neap tidal period when there was comparatively more pore space available (C. E. Hughes et al., 1998). Thus, rainfall-induced marsh recharge depends on the antecedent conditions.

The volume and depth of rainfall-induced recharge is generally small compared to the volume of saline water stored in the marsh soil. Consequently, the effect of the rainfall is mainly limited to the surface layer of the marsh soil (Boorman, 2019) where it leads to a reduction of the porewater salinity (A. L. H. Hughes et al., 2012; L. R. Gardner, 1975; L. R. Gardner et al., 2002), the extent of which depends on infiltration capacity and drainage intensity (Hemond et al., 1984). It may also depend on dynamic vegetation response and plant freshwater uptake, perhaps in response to the rain event, but such a hypothetical phenomenon is not yet studied. Rainfall supplements and alters water flow across the marsh surface (L. R. Gardner & Reeves, 2002; Mwamba & Torres, 2002), transporting solutes to tidal creeks during periods of high runoff (Boorman, 2019). Rainfall events can induce watertable fluctuations, although both the peak and decay of the watertable response to rainfall lag behind the rainfall event (C. E. Hughes et al., 1998) due to the low permeability of marsh sediments, which limits the rate of infiltration and drainage.

Lateral subsurface freshwater discharge from adjacent uplands is also an important source of freshwater recharge to salt marshes. Steeper hydraulic gradients (i.e., the upland hydraulic head is much higher than that in the marsh) increase groundwater flow toward the lower-elevation tidal channel, enhancing discharge (Guimond, Yu, et al., 2020; Thibodeau et al., 1998). This lateral freshwater recharge from adjacent uplands induces upward flow and dilutes the porewater salinity in the marsh (A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015; Thibodeau et al., 1998). To the extent that lateral surface runoff from the upland to the marsh also occurs, it may have a similar but poorly understood effect. When the lateral freshwater recharge is sufficiently large, it can balance evapotranspiration loss and limit infiltration of saline water during tidal inundation. This upward seepage can transport solutes from within the marsh sediment to its surface (Nuttle & Harvey, 1995; Thibodeau et al., 1998).

2.5. Sea Level Changes

The response of salt marshes to a rise in mean sea level depends on the elevation of the marsh platform relative to mean high water level (A. M. Wilson & Morris, 2012; Morris et al., 2002). In the vertical direction, marsh survival depends on its ability to trigger sedimentary accretion at a rate comparable to that of mean sea level (Fagherazzi et al., 2012; Kirwan et al., 2010; Marani et al., 2007). If a marsh can not keep pace with the regional sea level rise due to sediment deficiency, land subsidence (sinking of marsh sediments) or a significant time lag in response to sea level rise (A. D'Alpaos et al., 2011; Allen, 2000; Kirwan & Murray, 2008), the proportion of the marsh platform inundated at high tide would expand and/or the hydroperiod would increase (i.e., relative sea level rise). Although prolonged inundation affects plant survival and might cause marsh collapse, the productivity of some marsh species tends to increase with moderate increases in inundation (Kirwan et al., 2010; Morris et al., 2002). Positive eco-geomorphic interactions between marsh plants and hydro-sedimentary processes allow marshes to adapt to sea level rise (Baustian et al., 2012; Kirwan et al., 2016; Marani et al., 2007; van de Koppel et al., 2005).

Once relative sea level rise occurs, it has the potential to substantially alter groundwater flow and soil conditions in salt marshes (A. M. Wilson, Evans, Moore, Schutte, & Joye, 2015; FitzGerald et al., 2008; Ursino et al., 2004). Simulations of idealized marshes with fixed configuration and subjected to monochromatic tides show that relative sea level rise causes groundwater exchange and hence nutrient export to decline (A. M. Wilson & Morris, 2012). This decline occurs because both the area of the creek bank exposed at low tide and the overall drainage time decline. Field observations in marshes in the southeastern USA took advantage of large (50 cm) seasonal variations in mean sea level to monitor the impact of sea level rise on groundwater exchange (A. L. H. Hughes et al., 2015; A. M. Wilson, Evans, Moore, Schutte, & Joye, 2015). These investigations confirmed that relative sea level rise causes groundwater exchange in the marsh to decline, even when the relative sea level rise causes the area of the intertidal zone to expand (A. M. Wilson, Evans, Moore, Schutte, & Joye, 2015). Presumably, these effects reverse as the seasonal sea level declines.

Xin et al. (2010) examined the influence of relative sea level rise on soil saturation in a creek-perpendicular marsh section subjected to spring-neap tides. The results show that, overall, soil aeration declines as relative sea level rise occurs. Relative sea level rise causes the area of poorly aerated interior zone (induced by neap non-inundating tides) to contract considerably. In contrast, the contraction of the well-aerated near-creek zone (induced by spring inundating tides) is inapparent. Further, lateral water fluxes decline in a 3D marsh system even though inland water levels rise, and the permanently inundated subtidal zone expands whereas the other zones (e.g., tidal near-creek zone) contract (Guimond, Yu, et al., 2020).

These modeling studies omitted geomorphological evolution (A. M. Wilson & Morris, 2012; Guimond, Yu, et al., 2020; Xin et al., 2010). Sediment deposition and elevation accretion allow salt marshes to combat long-term sea level rise (Kirwan & Murray, 2007; Nicholls & Cazenave, 2010). How and to what extent the combined sea level rise and geomorphological evolution affect groundwater flow and soil conditions in salt marshes remain to be examined.

2.6. Stratigraphy

Sediment deposition in salt marshes depends on a number of factors including the tidal regime, vegetation trapping and suspended sediment concentration (Temmerman et al., 2003). Tide- and wave-influenced sediment deposition can in some cases cause coarse-grained materials to accumulate in the lower and middle parts of the marsh, whereas fine-grained materials deposit over the upper parts (G. Evans, 1965; Z. Zhou et al., 2015). In many other cases, sediment trapping by marsh plants results in deposition of fine-grained sediments on top of existing coastal sand layers (L. R. Gardner & Porter, 2001; Wiegert & Freeman, 1990). Consequently, the stratigraphy of coastal and near-river marshes is commonly characterized by low-permeability mud and/or silt loam overlying sands or sandy loam (as illustrated in Figure 3a; C. E. Hughes et al., 1998; J. W. Harvey et al., 1987; L. R. Gardner & Porter, 2001; Wiegert & Freeman, 1990; Xin et al., 2009). In large estuarine embayments where most sediment supply (and coarse sediment deposition) is quite distant from or exported in a different direction from extensive low-energy marshy shorelines (e.g., the South San Francisco Estuary), the local salt marshes may be predominantly clay or silt-clay to substantial depths (Moffett et al., 2012).

Sediment layering influences groundwater flow patterns and discharge into creeks (Xin, Li, & Barry, 2013). Groundwater discharge is focused at the elevation of the tide in homogeneous marshes (L. R. Gardner, 2005a,

Figure 6. (a) Macropores (crab burrows) distributed in marsh soils. An *Uca arcuata* crab is shown at the lower left corner. (b) Morphology of a crab burrow taken using polyester resin casting (Chongming Dongtan wetland, China; images captured by P. Xin on 11 July 2007).

2005b). In layered marshes (e.g., mud overlying sandy loam), groundwater largely emerges from the channel bottom and below the fine-grained upper sediment layer (L. R. Gardner, 2007). The total discharge from the marsh to the creek also increases compared to a homogeneous mud marsh (A. M. Wilson & Morris, 2012; L. R. Gardner, 2007; Xin et al., 2012), with corresponding increases in solute exchange. The presence of lower or higher permeability units intersecting the tidal creek may also lead to variation in groundwater discharge along the creek length (Moffett et al., 2008, 2012).

The presence of a sandy layer underlying the marsh enhances soil aeration and decreases fluid residence time compared to homogeneous fine-grained marshes (Xin et al., 2011, 2012). Indeed, sandy sub-marsh aquifers act as conduits from uplands to tidal creeks, which also strongly influence the hydrology of the surficial mud layer. For example, an underlying sandy aquifer can support upward seepage through the marsh mud in the mid-marsh and upper marsh, likely affecting soil aeration and porewater salinity (A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015).

2.7. Macropores

In addition to plants, microbes, fish, birds and mammals, salt marshes provide habitats for many species of sediment-dwelling invertebrates. Bioturbation by invertebrates changes the sediment structure, which may increase the soil permeability and reduce capillarity. Among these invertebrates, crabs are one of the principal bioturbation agents in intertidal salt marshes (McCraith et al., 2003). Through feeding and digging burrows, crabs make visible macropores in marsh soils (Figure 6). The size and shape of burrows differ with species and location (Iribarne et al., 1997; Mouton & Felder, 1996). Iribarne et al. (1997) found that crab burrows in creeks and channels have large entrances, whereas those in the vegetated zone are deeper and have small entrances. Some burrows can reach up to 1-m depth with entrance diameters ranging from 2 to 7.5 cm, whereas in other regions of the world only smaller crab species with smaller burrows are present. The density of crab burrows is also spatially

and temporally variable (McCraith et al., 2003), for example, higher in summer and near creeks than in the marsh interior (Guimond, Seyfferth, et al., 2020). Macropores increase the total area of sediment air/water interfaces and substantially enhance the permeability of sediments and the flow caused by rainwater and tidal water infiltration (C. E. Hughes et al., 1998; Escapa et al., 2007; Guimond, Seyfferth, et al., 2020). In addition, after a tidal inundation, the tidal water in the macropores can act as a reservoir to resupply water lost by evapotranspiration, resulting in salt enrichment and higher salinity in the sediments (C. E. Hughes et al., 1998; J. W. Harvey & Nuttle, 1995).

For layered salt marshes, recent numerical studies considered macropores as independent super-permeability zones embedded in the mud layer (Xiao, Wilson, et al., 2019; Xin et al., 2009; Xu et al., 2021). These studies showed that macropores penetrating the upper mud soil layer act as preferential paths for tidal water flow to the underlying higher permeability soil layers. In essence, macropores drain the upper soil during the falling tide and recharge the aquifer during the rising tide. Hence, the tidally averaged soil saturation of the shallow soil layer is lowered by macropores, improving local soil aeration (Xin et al., 2009). In contrast to layered soils, macropores affect only slightly tide-induced water exchange in a homogeneous soil (Xin et al., 2009).

Macropores also increase soil saturation variability (Xiao, Wilson, et al., 2019; Xin et al., 2009). Effects of macropores on the groundwater flow and water exchange are intensified in stratified marshes and as the permeability contrast between the two soil layers increases (i.e., the permeability of the underlying layer increases; Xin et al., 2009). They can also increase evaporation rates in low-permeability soils with shallow watertables (T. Zhou et al., 2020). Macropores behave as a preferential flow path, delivering groundwater to the shallow soil, which results in higher soil saturations in the near-surface area (T. Zhou et al., 2020). Furthermore, most of the evaporated water bypasses the soil matrix and has only a little contact with soil grains.

2.8. Interacting Effects Among Forcing Factors

Natural salt marshes are affected by multiple forcing factors, for example, tides, rainfall and evapotranspiration (L. R. Gardner & Reeves, 2002; L. R. Gardner et al., 2002). Tidal flooding recharges marsh soils that have previously lost water by evapotranspiration and/or lateral drainage (L. R. Gardner & Reeves, 2002; Nuttle & Harvey, 1995). This also brings marine-sourced solutes (e.g., salt and sulfate) into the marsh systems. On the contrary, drainage intensity and evapotranspiration determine how much water and solutes are removed from the marsh soil (Bornman & Adams, 2010; Hemond et al., 1984; Moffett, Robinson, & Gorelick, 2010). In particular, tides dominate watertable variations and, together with sediment permeability, regulate the impact of rainfall and evapotranspiration on the watertable in near-creek zones (L. R. Gardner et al., 2002). In marsh interiors, the local water balance and watertable are mainly governed by vertical fluxes induced by rainfall and evapotranspiration, or by infiltration (Hemond & Fifield, 1982; H. Zhang, Moffett, et al., 2014; Marani et al., 2006). When rainfall-induced infiltration is less than evaporation, salt is concentrated continually in the sediment during marsh exposure, a process that continues until salt precipitation inhibits evaporation (C. Zhang, Li, & Lockington, 2014). When the groundwater is hypersaline (in excess of the tidal water salinity), tidal flooding could remove salt from the shallow soil layer via salt diffusion across the sediment surface (Moffett, Robinson, & Gorelick, 2010; Morris, 1995) as well as from salt fingers that penetrate into the deep sediments (Shen et al., 2016, 2018). The groundwater flow and salt movement could be further complicated by plant transpiration (Moffett & Gorelick, 2016b; Moffett et al., 2012) but the details of these interactions are not yet entirely illuminated. When evapotranspiration is low, the influx of fresh groundwater recharge (including regional groundwater input and rainfall-induced infiltration) can exceed water loss by evapotranspiration; then, groundwater drives a net upward flow of water and salt out across the sediment surface (Nuttle & Harvey, 1995). On the other hand, increased evapotranspiration causes more water loss and in turn enhances rainfall-induced infiltration (A. L. H. Hughes et al., 2012). These processes are expected to be regulated by marsh profiles (e.g., concave-up or concave down) and microtopography, but these interactions have not yet been examined.

The influences of rainfall, evapotranspiration and tides are spatially variable, resulting in variations in soil saturation and porewater salinity. Numerous field investigations show that porewater salinity gradually increases with marsh elevation, reaching a maximum around the high-high tidal level and then decreases beyond it (Figure 7; Adams, 1963; Pennings & Bertness, 1999; Pennings & Callaway, 1992; Xiao et al., 2017). This is due to the longer evaporation periods that accompany increases in marsh elevation. The increase in evaporation leads to more salt accumulation in surface soils affected by tidal inundation, whereas above the high-high tide level the marsh is rarely inundated and salt input is limited (inundation frequency decreases as the elevation decreases in
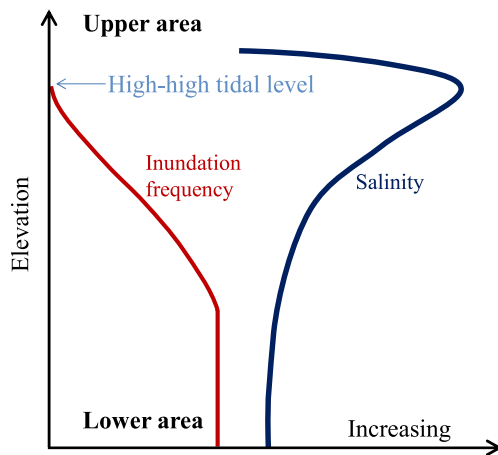
Figure 7. Typical porewater salinity distribution and inundation frequency along a marsh section with increasing elevation.

the upper marsh, illustrated in Figure 7). In this way, the soil salinity tends to decrease in the area above the high-high tide level (H. Wang et al., 2007; Shen et al., 2018).

Morris (1995) established a mass balance model considering gravity drainage, infiltration and evapotranspiration to investigate the processes of water and salt transport in salt marsh soil (Figure 3b). Their results show that sediment physical properties, evapotranspiration and elevation are important determinants of seasonal variation of soil salinity. As marsh elevation increases, losses of salt by drainage and diffusion decrease, and the range of salinity variation increases at higher elevations (Morris, 1995). Based on a similar model, H. Wang et al. (2007) confirmed that the high-high tidal level determines the location of the salinity maximum in a coastal salt marsh, which is commonly seen in field observations. They also pointed out that tidal fluctuations primarily control the width of the hypersaline zone. Evapotranspiration, temperature, hydraulic conductivity and salinity of incoming tidal water significantly affect the maximum salinity (H. Wang et al., 2007).

Moffett et al. (2012) developed a delineation based on marsh plant-water-sediment characteristics termed "ecohydrological zones." Marsh ecohydrological zones emerged from the spatial intersection of sediment heterogeneity and evapotranspiration heterogeneity, as they are the key controls on water movement throughout the marsh plain, which covers most of the marsh's area. They demonstrated that ecological (especially evapotranspiration) and hydrogeological heterogeneity can combine to create diverse hydraulic habitats even in a small salt marsh area. Guimond, Yu, et al. (2020) delineated similar hydrologic zones based on 3D studies. Such characteristic flow zones were confirmed in the field by other studies (A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015; Xiao et al., 2017).

Complementing such 2D-planar "ecohydrological zones" combining sediment and plant hydraulic controls on water infiltration and extraction, Xin et al. (2017) combined spring-neap tide, evaporation and rainfall considerations into 2D, creek-perpendicular "saturation-salinity zones" that also differed with (uniform) sediment permeability. A high permeability marsh exhibit low soil saturation and low salinity dominated by spring tides near the creek bank, high saturation and salinity dominated by evaporation and rainfall in the marsh interior (Hemond & Fifield, 1982; Marani et al., 2006), and a transitional zone between these extremes with gradually increasing soil saturation and salinity with distance from the creek (Figure 8). However, in a lower permeability marsh, soil saturation, and salinity are higher near the creek, due to evaporation and capillarity (Xin et al., 2017).

Putting these two frameworks together (Moffett et al., 2012; Xin et al., 2017), the subsurface hydrology of a low-permeability marsh is dominated overall by the matrix, that is by the marsh sediments themselves and their embedded vegetation, as described by the "ecohydrological zones" of Moffett et al. (2012), which describe mostly vertical exchange. In contrast, a high-permeability marsh is dominated by the creeks, the tidal amplitude they experience, and presumably their size and density in dissecting the marsh, and the creeks control the "saturation-salinity zones" that extend away from them as by Xin et al. (2017). Together these perspectives frame a full spectrum of near-surface marsh hydraulics and dynamics that can span a variety of geological, physiographic, and coastline settings.

The interacting factors that influence groundwater flow in salt marshes also drive significant temporal variability in hydrologic conditions. As an
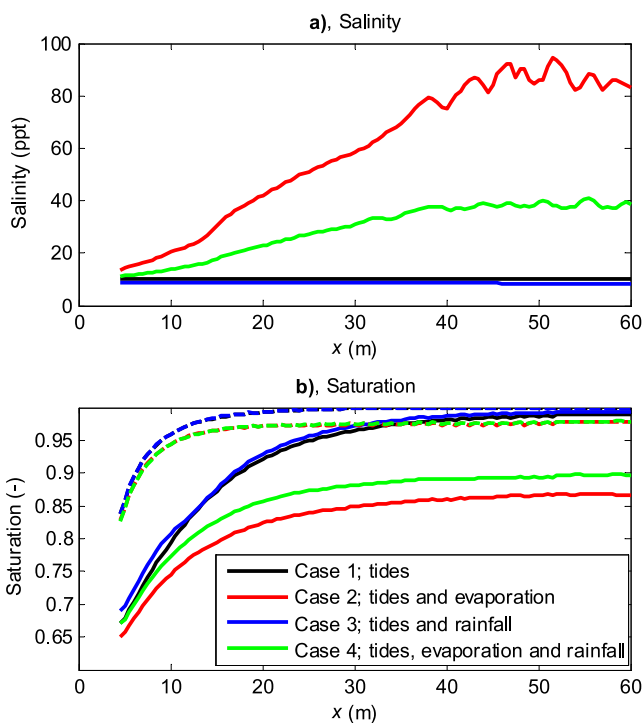
Figure 8. Annually averaged porewater salinity ((a), ppt means parts per thousand, i.e., grams of salt per kilogram of solution) and soil saturation (b) in the 0.3-m shallow soil layer of a creek-perpendicular and sandy-loam marsh section subjected to different combinations of spring-neap tides, evaporation and rainfall. For comparison, the annually averaged soil water saturation over the spring tidal cycle is also plotted (dashed lines in panel (b)). Forcing factors considered are given in the figure legend. The horizontal axis shows the distance to the creek (adapted from Xin et al. (2017)).

example, A. M. Wilson, Evans, Moore, Schutte, and Joye (2015) analyzed the groundwater discharge (lateral drainage) based on a year-long time series of hydraulic heads in a salt marsh. Although independent linear regressions suggested that neither spring-neap changes in tidal amplitude nor seasonal variations in mean water level caused more than a factor of two variation in groundwater discharge from the marsh on their own, calculations based on aggregate field observations indicated that interactions between these forcing factors caused groundwater discharge to vary by an order of magnitude over the course of the year.

3. Approaches for Determining Surface Water and Groundwater Interactions

3.1. Field Measurements

Although several methods qualitatively detect the occurrence of marsh SW-GW exchange, methods of direct empirical quantification of such interactions are challenging and few for any environment. In salt marshes, directly capturing transient rates of exchange are particularly complicated due to their oscillating flow directions, strongly 3D flow systems and periodic flooding. Many methods for quantifying intertidal SW-GW exchange are partially indirect, incorporating an element of modeling, such as a 1D gradient-flux model between measured hydraulic heads, or (bio)geochemical mixing models of end-member source waters (Burnett et al., 2006). This section reviews field methods for qualitative detection of SW-GW exchange, direct quantification of that exchange, and indirect or hybrid empirical-model quantification methods. Fully modeling-based methods for quantifying fluxes and higher dimensional flow dynamics are provided in Section 3.3. In this section, distinctions are not always made among aims to quantify vertical fluid exchanges across the periodically inundated marsh platform, or lateral exchange between marsh sediments through creek banks, or shoreline exchange and aquifer SGD, but caveats of each method's use in any of these settings are noted.

3.1.1. Qualitative Detection of Surface Water and Groundwater Mixing

In the simplest case, one can detect groundwater discharge from a salt marsh by visual inspection: one can observe a seepage face, discrete groundwater seeps, or draining macropores in the bank of a creek or on the shoreline of a marsh. However, even detailed inspection cannot distinguish discharging groundwater that infiltrated a short time ago or nearby, akin to bank storage or hyporheic flow, from discharge that traveled a long time and distance in the marsh or an underlying aquifer. Salinity of seepage, such as by a quick and inexpensive refractometer measurement, can help distinguish fresh terrestrial aquifer discharge from hypersaline, evaporation-concentrated marsh water, but will not conclusively distinguish most marsh groundwater from rapidly recirculated surface water; the same is generally true of spot-measurements of other chemical constituents of visually located seepage.

Temperature information (e.g., from in situ probes, cables, or thermal imaging) can distinguish marsh groundwater seepage that has spent sufficiently long in the marsh subsurface to cool during a warm season or warm during a cold season. On average, more characteristic marsh groundwater should be similar to mean annual air temperature, whereas rapidly recirculated surface water should be more similar to recent or diurnal mean air or surface water temperatures. Thermal methods cannot necessarily distinguish hypersaline marsh water and fresh aquifer water that have both spent ample time in the terrestrial subsurface, however. Fiber-optic distributed temperature sensing can be useful to locate areas of groundwater discharge that may not be visually apparent and may be subaqueous or periodically flooded, especially along linear channel features (Moffett et al., 2008). Airborne or above-water thermal imaging has been often used to locate plumes of cooler groundwater discharging into warmer surface water (Mulligan & Charette, 2006; Portnoy et al., 1998; T. K. Duarte et al., 2006), although interpretation can be complicated as cooler groundwater may plunge below warmer surface water if quite cold or saline and so not be detected, unless the surface water salinity is sufficiently high to overcome the groundwater's thermohaline density advantage. Using thermal imaging to detect warmer groundwater discharging into cooler surface water in the winter might be a more reliable method, but has rarely been used, perhaps due to impracticality in temperate locations where nearshore water freezes. Thermal imaging of a subaerial bank seepage face is also qualitative, but a more straightforward method to detect groundwater discharge (Deitchman & Loheide, 2009; Pfister et al., 2010).

An important note regarding qualitative detection of marsh groundwater discharge and of SW-GW exchange is that, although useful, such observations will likely be biased to note larger seepage areas in tidal creek banks at low tide, whereas smaller seeps, subaqueous discharges to tidal creeks or the coastal zone (SGD), filling of
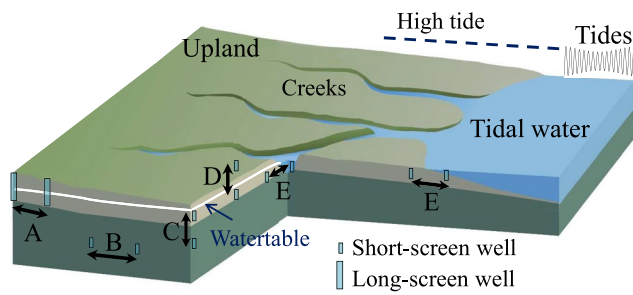
Figure 9. A conceptual diagram of using piezometer water levels to quantify hydraulic gradients and fluxes in salt marshes.

bank storage, and infiltration across the marsh plain will be more difficult to observe. More complete and quantitative methods are warranted for detailed studies.

3.1.2. Direct Measurement of Marsh-Surface Water Exchange Rates

To directly measure SW-GW exchange, a combination of methods is needed to quantify concomitant infiltration and discharge. Seepage meters are the only available tool to capture and quantify groundwater discharge. A seepage meter consists of a chamber inserted into the sediment surface, which has on its upper surface a port allowing free fluid outlet into a thin-walled water collection bag (Lee, 1977). The bag is pre-filled with a known mass and volume of water to reduce artifacts from bag-filling resistance, and in case the hydraulic gradient at the chamber should be in a condition to drive inflow rather than outflow at the start of measurements, or overall. The bag and chamber should remain submerged, as excessive noise is generated by a free water surface so as to make apparent seepage data unusable (D. Rosenberry & Morin, 2004). Either the discharged volume is recorded continuously (automated), or after a pre-determined monitoring time the bag is manually removed and weighed to quantify the discharged water (Taniguchi, Burnett, Cable, & Turner, 2003). Fluid electrical conductivity and temperature can be monitored both inside and outside the bag (Taniguchi et al., 2007).

The greatest limitation to the use of seepage meters in the marsh setting is the requirement to maintain the meter and bag submerged throughout deployment. This limits application to larger tidal channels that retain water at low tide, or to marsh-shoreline settings. Consequently, most coastal seepage meter studies have targeted offshore discharge of fresh aquifer water as SGD; many fewer studies have successfully applied seepage meters in salt marshes. Additional challenges with seepage meters are: (a) sediment, hydraulic gradient and discharge heterogeneity, which notoriously can yield significantly different measurements even at adjacent sampling locations (Taniguchi, Burnett, Smith, et al., 2003); (b) difficulty in avoiding artifacts, or positive bias in discharge measurements, due to anomalies caused by the bag or by fluid advected very shallowly through sediments and just under the meter edge due to overlying barometric, waves, or tide fluctuations (Bernoulli effect) rather than due to upland hydraulic heads driving "true discharge" (Cable et al., 2006; Corbett & Cable, 2003; Shinn et al., 2002, 2003), highly permeable sediments (D. O. Rosenberry, Nieto López, et al., 2020), meter or bag leakage, or other technical challenges; and (c) immense labor and deployment and measurement care required per point measurement in space and time, belied by the simple concept and low cost of the instrument (D. O. Rosenberry et al., 2008). In terms of collected water chemistry, a user should be aware that seepage meters and related devices (Sakamaki et al., 2006) most likely capture larger or macropore flow and do not necessarily sample fluids from smaller pores or matrix-fluid exchange (J. Harvey et al., 1995). Duque et al. (2020), D. O. Rosenberry, Duque, and Lee (2020), and D. O. Rosenberry et al. (2008) review the seepage meter method, explain of materials and techniques, and thoroughly discuss common sources of error and best practices.

3.2. Indirect Quantification of Surface Water and Groundwater Exchange

3.2.1. Hydraulic Gradient-Flux Approaches and Darcy's Law

The most common means to quantify SW-GW exchange in the coastal zone is to empirically model average groundwater flow. This approach uses well or piezometer water levels to quantify a hydraulic gradient, and Darcy's Law to translate the gradient to a flux (flux per unit discharging area). The horizontal sediment saturated hydraulic conductivity (K_{sat}) is the empirical scalar of gradient-flux proportionality. Three aspects of this approach (choice of model, measurement of hydraulic gradient, and estimation of hydraulic conductivity) warrant careful consideration in the tidal setting and require different user choices depending on if the flux to be quantified is: (A) upland, shallow, and unconfined groundwater exchange horizontally with the marsh, (B) terrestrial aquifer (potentially confined beneath marsh muds) horizontal flow beneath the marsh and/or SGD, (C) net vertical exchange between an underlying porous aquifer and overlying marsh sediments, (D) net vertical marsh groundwater flux between the marsh subsurface and surface water, or (E) lateral SW-GW exchange between marsh groundwater and tidal surface waters through tidal channel banks or the periodically exposed marsh shoreline at short timescales (illustrated in Figure 9). For marshes emplaced on sandy coastlines, (A) and (C) are likely inflows to

the marsh and (D) and (E) likely outflows; for marshes emplaced in low-permeability or bedrock locations, inputs from (A) and (C) may be zero or negligible and both inflow and outflow via (D) and (E) dominate.

For the flow paths of interest here, Darcy's Law can be written $q = -K_{sat}\Delta h/d$, where q is the net groundwater flux per unit cross-sectional area, Δh gives the difference of hydraulic head between two locations and d is the distance. The Darcy's Law approach is suited for cases (A), (B), and (C) above, but not (D) or (E). For case (A), this approach can diagnose the likely average groundwater exchange horizontally along a transect from the upland to the marsh (Tobias et al., 2001). In this application, monitoring wells should be shallow but fully screened, so as to capture the true watertable fluctuations but not penetrate into underlying sediment layers (which are likely of contrasting permeability). The distance between the wells should be minimized and directed perpendicularly across the upland-marsh ecotone to best target a short and direct upland-marsh flow line. The horizontal saturated hydraulic conductivity (not vertical) of sediments just below the watertable should be determined experimentally. Slug tests or bailing tests commonly used to estimate horizontal sediment hydraulic conductivity around a well may be adequate for the upland and marsh sediment estimates (if wells are shallow and screened only in the target near-surface layer near the watertable). If or when marsh water levels rise or fall rapidly with high or low tides, care should be taken in interpreting the apparently high-magnitude horizontal groundwater fluxes calculated by the method as the duration of this strong gradient may be too short to drive appreciable groundwater movement.

For case (B), a similar approach can be used to estimate terrestrial aquifer horizontal flow (potentially confined) beneath the marsh and/or SGD from an aquifer beneath the marsh to the coastal zone (A. M. Wilson, Evans, Moore, Schutte, & Joye, 2015). In this case, piezometers with short well screens should be installed fully through the upland, marsh, and/or offshore surface sediments. The screens should be solidly emplaced into the underlying aquifer material and the remainder of the casings should be solid above the screen and to above the maximum expected high water level. The casings should be well-sealed to the surrounding sediments and the piezometers should be thoroughly developed by pumping and purging to clear the screens of low-permeability material potentially entrained during installation or since last use. The screens should be of the maximum slot size possible to both prevent the entry of aquifer or screen-packing material into the casing but not add resistance to the free passage of the groundwater. Standard aquifer tests (e.g., slug test, pump test, and injection test) will serve to estimate the horizontal hydraulic conductivity. Over longer timescales, such as monthly or more, an even simpler method to quantify the discharge of a coastal aquifer, confined beneath a lower-permeability marsh, to a tidal creek or as SGD to the coastal zone can be developed with high fidelity as a multiple linear regression of SGD to tidal amplitude and 30-day average surface water level (A. M. Wilson, Evans, Moore, Schutte, & Joye, 2015).

For case (C), net vertical exchange between an underlying aquifer and overlying marsh sediments can be estimated by applying Darcy's Law in the vertical direction (J. W. Harvey & Odum, 1990), provided that the exchange being targeted across the subsurface layered-sediment interface that is the bottom surface of the marsh is fully beneath the watertable. This method is not suitable for application across the variably saturated sediments of the shallow marsh surface (case D). In this case, piezometers with short screens should be used, with all the same considerations as case (B) but installed to several target depths in a "nest" at a common location. Piezometers should be spaced closely enough to plausibly lie along the same or comparable vertical flow path but separated far enough apart that flow anomalies from installation or presence of one avoid affecting the other(s) as much as possible. A single casing with multiple independent head-measurement ports open to screens at different depths is ideal; alternatively, installation of several piezometers in an area of presumed spatially uniform vertical flow, but with perhaps a meter or two between piezometers, may be desirable. Care to use a vertical saturated hydraulic conductivity value is key in this case; typical slug-test methods to estimate K_{sat} provide a largely horizontal estimate, whereas vertical hydraulic conductivities (K_{sat_V}) are often an order of magnitude lower than horizontal ones (K_{sat_H}), such that one might estimate $K_{sat\_V} = 0.1\ K_{sat\_H}$ (Tobias et al., 2001).

Darcy's Law is not suited for estimating net vertical marsh groundwater flux between the marsh subsurface and surface water because this case (D) violates the assumption of fully saturated conditions and flow driven only by gravitational (neglecting capillary) forces between the two total head measurements. In the fine-textured sediments typical of many salt marshes, capillarity should be expected to be an important near-surface hydraulic processes, as well as the fluxes by soil surface evaporation and subsurface root water uptake (typically in the top 30 cm or so). For example, evapotranspiration and plant root rehydration may lower the marsh watertable continually between semidiurnal flooding tides, day and night (Dacey & Howes, 1984), but applying Darcy's Law to the head difference between the watertable and flood water would be mechanistically incorrect. Capillary rise

may maintain the smaller pores of a fine-textured marsh surface in a saturated condition (i.e., capillary fringe intersecting the marsh surface) even throughout low tides, whereas capillary suction may both enhance surface water infiltration in smaller pores and decrease overall hydraulic conductivity via air blocking by entrapped air accumulation in larger pores' preferential flow pathways (Princ et al., 2020; Snehota et al., 2015). This entrapment of an aerated layer, its pressures, gas composition, and compression dynamics during tidal flooding were documented decades ago by Chapman (1938) and Chapman (1940). Although likely broadly understood in theory by coastal and wetland hydrogeologists, the specific application to the case of empirical and modeling study of salt marshes of distinctions between true saturated hydraulic conductivity, quasi-saturated hydraulic conductivity with entrapped air, and unsaturated hydraulic conductivity with atmospherically connected air fluid phase (Princ et al., 2020; Sakaguchi et al., 2005) has yet to be thoroughly developed. Notwithstanding reinvigoration of shallow marsh plant-groundwater studies by Ursino et al. (2004), following earlier work (Dacey & Howes, 1984; Nuttle & Hemond, 1988), further investigation of relations among soil water and evapotranspiration in salt marshes is still warranted, as has been done in other wetlands (Schwärzel et al., 2006). Fruitful avenues of improved empirical flux analysis could expand on earlier methods that incorporated nuances into Darcy's Law calculations such as sediment compressibility and static (equilibrated) versus dynamic (un-equilibrated) head responses to flooding (Hemond et al., 1984), sediment dilation (Nuttle et al., 1990), and dynamic hydraulic conductivity-total head relationships (Nuttle, 1988b), and could valuably expand understanding of marsh plain surface and plant salt effluxes (Moffett, Robinson, & Gorelick, 2010; Nuttle & Hemond, 1988).

Darcy's Law is also not suited for estimating (E) lateral marsh SW-GW exchange between marsh groundwater and tidal surface water through tidal channel banks or through the periodically exposed marsh shoreline at short timescales. Darcy's Law can be used in confined aquifers that underlie many marshes, but unconfined systems violate the Law's assumption of an effectively horizontal watertable (potentiometric/piezometric surface) over the distance (d) between the two total head measurements. Fortunately in this setting the Dupuit Equation might instead be used, if its alternative assumptions of a constant-sloped watertable, flow parallel to the watertable slope, and entirely horizontal flow seem applicable (J. W. Harvey et al., 1987). Although the latter condition is typically violated by channel-bank and shoreline settings, this source of error is within the overall method uncertainty and its constituent parameterizations, especially of K (D. O. Rosenberry et al., 2008). In this case, the Dupuit Equation can be applied analytically (details in D. O. Rosenberry et al. (2008)). An alternative approach is to use a modified version of the Darcy Equation that appreciates tidal periodicity of the bank seepage face, sediment conductivity and compressibility, and integrates over a finite time such as a tidal cycle (Nuttle, 1988a), but at this point one is diverging from predominantly empirical, although indirect, quantification of groundwater discharge into 1D numerical modeling requiring calibration of multiple model parameters.

In any use of long-screen wells or short-screen piezometers to quantify head gradients or sediment hydraulic conductivities several additional methodological considerations must be appreciated. Wells with long screens, or fully screened from depth to near the surface, are called for if the intention is to monitor the full range of possibly large watertable fluctuations, but then the user must be aware that such long, fully screened pipes may serve as infiltration conduits or drains. Piezometers, or short-screened monitoring wells, should be carefully emplaced to the desired monitoring depth, but with awareness that they will likely only reflect a damped version of true watertable fluctuations even in an unconfined flow setting, and that vertical gradients from site-scale hydrological dynamics or subtle sediment layering may be present. Short-screened piezometers may also have their screens accidently located in local heterogeneity, requiring the replication of field measurements at multiple points or transects. Also, several such short-screened monitoring points intended for head gradient analysis along an assumed transect may or may not actually lie on a common flow path. Installation of sand completely around both long and short well screens is essential in salt marshes, and both well development and routine purging, as well as submerged datalogger cleaning, are warranted for prevention of clogging by sediments and/or biofilms. Well clogging is still possible, more likely with finer screens and in less hydrologically dynamic settings, and should be identified by over-damped data. Advice on additional issues of intertidal well installation, screening, and maintenance is provided by D. O. Rosenberry et al. (2008).

Finally, an important matter when working in rapidly and highly dynamic hydraulic environments are the well screen and instrumentation response times. Well-bore storage and response lag through the well screen add notable artifacts in water level records from wells of insufficiently small diameters and insufficiently long screens, particularly when sediment permeability is low or tidal amplitude is high (L. R. Gardner, 2009). Best practice is to

emplace a well's pressure transducer in the bottom of a slug or inner casing that displaces nearly the full wellbore storage volume, although this unavoidably induces the tradeoff that the support volume for the well's fluid exchange and hydraulic measurements is thereby reduced (L. R. Gardner, 2009). Consequently, the displacing slug should be removed and the well equilibrated to background water levels for at least several hours or a full tidal cycle prior to conducting any well tests, such as a slug or bailing test for hydraulic conductivity estimation. Related wellbore storage issues can be created for in-well salinity or temperature loggers if the storage allows either the vertical movement of stratified layers of different salinities or temperatures, or the overturning of such layers, which, in either case, can pass the sensors and result in seemingly highly interesting but actually useless "data."

3.2.2. Water Balance Approaches

It is also possible to indirectly estimate net SW-GW exchange by accounting for the other terms of the near-surface groundwater balance (Burnett et al., 2006; J. W. Harvey & Odum, 1990). This root zone water balance can be written as:

$$I_P + I_S + Q_{\text{in}} = \text{ET} + Q_{\text{out}} + \Delta S \ \text{ or } \ Q_{\text{net\_out}} = I_P + I_S - \text{ET} - \Delta S \quad (1)$$

for infiltrated precipitation (I_P), infiltrated surface water (I_S), total evapotranspiration (ET), groundwater inflow from upland (Q_{in}), groundwater discharge (Q_{out}), and change in soil water storage (ΔS), or rewritten in terms of net groundwater discharge as $Q_{\text{net\_out}}$. To solve for net groundwater discharge, the other terms must first be quantified.

Infiltration rates can be estimated with some careful considerations. Tidal water infiltration is gravity-driven, that is, standing-water pressure head gradients dominate the process, and in large pores. Its infiltration capacity is simulated by a ponded-head method such as a field permeameter or double-ring infiltrometer, or a laboratory permeameter or constant- or falling-head permeability test. It is not within the scope here to review these standard hydrogeological measurement methods and caveats; only to highlight that marsh surfaces are often low-permeability, except where macropores are present, and so proper instrument sealing to the sediments is crucial. The best approximation to I_S might be obtained on a day of particularly extended low-tide such that a long-duration falling-head infiltrometer measurement might be conducted as a simulated tidal flooding signal.

Precipitation infiltration, in contrast, will occur when the marsh surface is exposed and unsaturated. This infiltration is tension-driven, at least at first, that is, capillary forces in small pores dominate the process. It is to be emphasized that ponded-infiltration and tension-infiltration operate differently at the mechanistic level of which sediment pores fill, and how fast. Requiring unsaturated conditions for tension-infiltration is key, as fine-textured marsh sediments may remain largely tension-saturated between tides, that is, the capillary fringe above the watertable may remain at or very near the marsh surface even when the marsh is exposed. If precipitation infiltration can occur between flooding tides, then a reasonable simulation of the potential rate of I_P will be given by tension infiltrometer tests conducted with fresh water. (And if precipitation ponds, then by ponded infiltrometer tests, as above, but with shallow fresh water.) The infiltrometer measurements must then be scaled for the actual infiltration rate to decrease over time during the precipitation event, such as by a Green-Ampt type of approach. Alternatively, infiltration from floodwaters can be estimated as a function of the pressure drop between the flooding surface water and the as-yet un-equilibrated groundwater pressure, vertical saturated hydraulic conductivity, and compressibility (Hemond et al., 1984; Nuttle & Hemond, 1988). Precipitation occurring during a flooding tide, and during times of complete marsh sediment saturation, should be omitted from the marsh root zone water balance.

Quantifying ET and ΔS can be done directly with in situ lysimeter systems (Dacey & Howes, 1984; Hussey & Odum, 1992), but as lysimeters are designed specifically to prohibit subsurface exchange, they omit effects from groundwater drainage (Howes & Goehringer, 1994), and so the $Q_{\text{net\_out}}$ value desired is artificially zero. Eddy covariance systems can quantify ET (C. E. Hughes et al., 2001; Moffett, Wolf, et al., 2010), although they suffer from chronically incomplete mass balance closure and may not provide concurrent data on ΔS. The simple technology of the evaporating porous-cup atmometer might fruitfully be revived from early salt marsh studies (Johnson & York, 1915) for further modern use, or use of simple reference pan evaporation calibrated with a marsh "crop coefficient" (Moffett, Wolf, et al., 2010).

Change in sediment moisture storage, ΔS, is typically measured by one or more soil moisture sensors that quantify the dielectric constant of moist soil, which is strongly related to soil water content; other methods are under

development in agricultural settings (Hardie, 2020) but not yet translated to salt marsh studies. However, among the most common soil moisture sensors, soil moisture resistance sensors (e.g., gypsum block or similar) do not work in saline environments, tensiometers might become clogged by air or (bio)solids among rapidly oscillating flow directions. Electrical conductivity- or electromagnetic-based sensors or geophysical methods are distorted by salinity and clays and require data processing adjustments for high-salinity and consequently low signal penetration-settings (e.g., Moffett, Robinson, and Gorelick (2010)). Capacitance soil moisture sensors are preferred for the salt marsh setting, but are not without their uncertainties, and have a need to be calibrated to the specific soil-fluid mixture of the field setting. The common presence of entrapped air or other gas in the shallow marsh subsurface has long been recognized (Chapman, 1938), adding caution to the use and interpretation of soil moisture sensors as representative in the tidal setting. Finally, the user must carefully consider the depth of likely soil moisture dynamics and the large magnitude of its variations at different locations in the marsh: spanning 0–30 cm or the root zone in the marsh interior, but as much as the full channel bank depth near a tidal channel (e.g., 0–100 cm or more). Having estimated I_P, I_S, ET, and ΔS, however, one can then estimate the net groundwater discharge to surface water using the water-balance approach, as $Q_{\text{net\_out}} = I_P + I_S - \text{ET} - \Delta S$.

Water balance principles can alternatively be applied to estimate net marsh water loss (by drainage and evapotranspiration) over longer timescales of months to years, by pairing data on the mean marsh watertable drawdown per tidal cycle with data on the amount of infiltration required to (re-)raise the marsh watertable, which can be obtained via analysis of rainfall-well response data during low tides (L. R. Gardner & Gaines, 2008).

3.2.3. Tracer Approaches

A final group of methods to indirectly quantify SW-GW exchange is the use of natural or introduced tracers. Quantifying arrival times, dispersion, and dilution of introduced tracers via breakthrough curve analysis, such as after introducing bromide into groundwater (Tobias et al., 2001) or rhodamine dye into surface water (Blanton et al., 2010), is readily completed using standard hydrologic tracing approaches. Detecting SW-GW mixing has variously made use of strongly contrasting temperatures or concentrations of naturally occurring geochemical tracers, such as metal and salt ions, stable isotopes, radioactive nuclides, salinity, dissolved gasses such as methane, and anthropogenic tracers such as caffeine (Burnett et al., 2006). However, temperature-based methods effectively require fitting a heat transport model, and solute tracer methods require constraining all possible sources and assembling a source water mixing model. For the latter, choice or measurement of "pure" source water end-members may be fraught and require additional assumptions about transport and mixing, and sources and sinks, between available measurement locations and the specific marsh sections of interest (although this can be mitigated with a multi-tracer approach). The specific and highly useful application of radium isotopes in such mixing models for the indirect quantification of overall diffuse groundwater discharge volumes to saline surface water is detailed in Section 3.2.4.

Alternatively to source water mixing models, a full marsh-system mass balance of one of these tracers can be assembled by means similar to the construction of the marsh water balance, above (J. W. Harvey & Odum, 1990; Morris, 1995). The marsh salt balance is a common choice, constraining which can be paired with constraining a marsh water balance containing two unknowns, to enable solving the coupled system of two equations. In this application, however, there remain challenges quantifying marsh salt storage, especially during times with low flow and higher accumulation (Tobias et al., 2001) as well as specific rates and mechanisms of salt loss from the marsh plain surface and/or plant salt uptake (Moffett, Robinson, & Gorelick, 2010; Nuttle & Hemond, 1988).

3.2.4. Example: Marsh Groundwater Discharge Traced by Radon and Radium Isotopes

The utility of naturally occurring radionuclide tracers in SW-GW studies derives from the wide range of chemical properties and half-lives exhibited within the uranium (U) and thorium (Th) decay series. These decay series contain six thorium isotopes, four of which decay to the "Radium Quartet" (Rama & Moore, 1996) of isotopes ^{223}Ra, ^{224}Ra, ^{226}Ra, and ^{228}Ra. Thorium is extremely insoluble in most natural water, causing it to remain tightly bound to sediments. Thus, thorium decay provides a continuous source of Ra isotopes, which partition between the solid and dissolved phases. This partitioning is strongly dependent on salinity, with higher salinities releasing more dissolved Ra (Y. Li et al., 1977). Once Ra is removed from sediments, the isotopes regenerate on time scales controlled by their half-lives. For example, 1600 years are required to regenerate 50% of ^{226}Ra, but only 3.66 days to regenerate 50% of ^{224}Ra (Figure 10). Another radionuclide that is useful in salt marsh studies is the
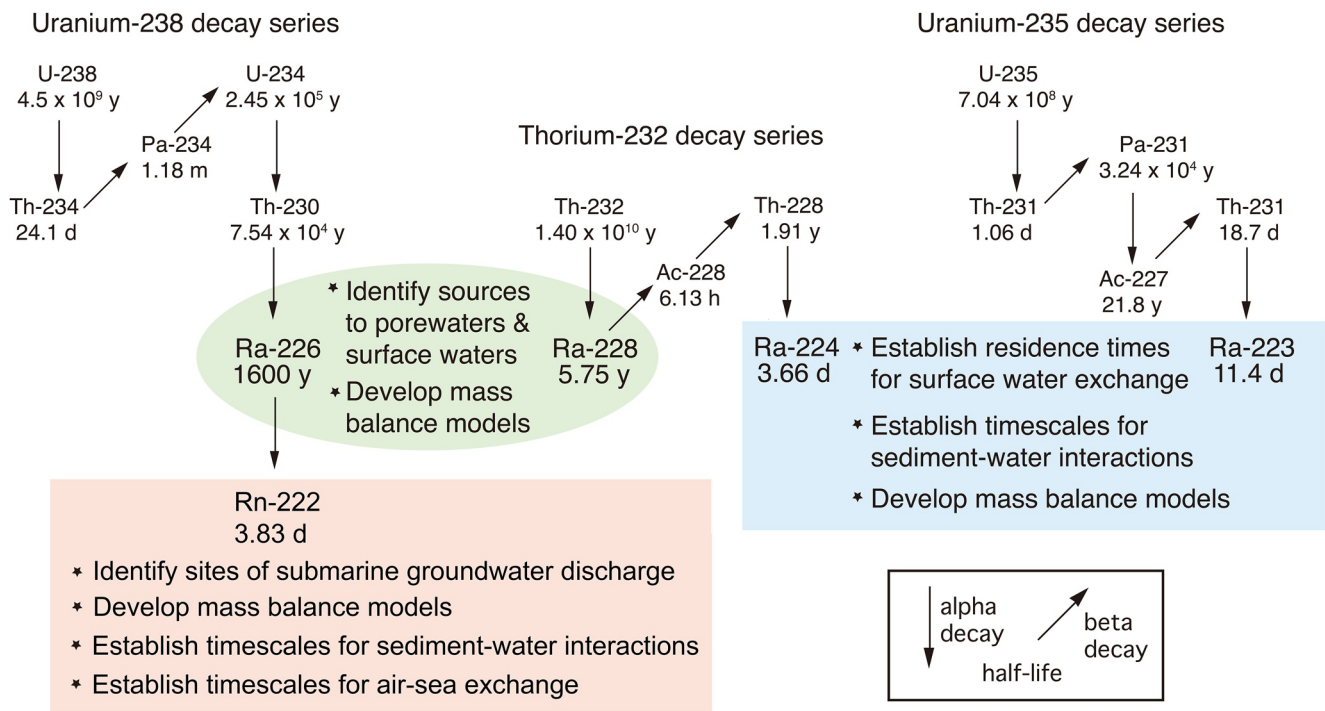
Figure 10. Schematic diagram of the natural uranium (U-238 and U-235) and thorium (Th-232) decay series. The decay type (alpha or beta) is shown by arrows; half-lives are given below each nuclide. Within the U and Th decay series are four radium isotopes derived from thorium decay plus radon-222. These isotopes have numerous applications in establishing timescales and sources of materials to salt marshes.

^{226}Ra daughter, the radon isotope ^{222}Rn. Like ^{224}Ra, ^{222}Rn regenerates quickly, and because it is a gas it has little affinity to adsorb to sediments and partitions almost completely to the dissolved phase.

The use of Ra and Rn in groundwater studies was established by Krishnaswami et al. (1982), who recognized that Ra could occur in groundwater by either mineral dissolution or by alpha recoil from thorium decay. They reasoned that the recoil component could be estimated by the activity of ^{222}Rn, assuming full dissolution of any Rn produced. In their studies of fresh groundwater, Krishnaswami et al. (1982) found ^{224}Ra activity was 3–4 orders of magnitude less than ^{222}Rn, indicating that most ^{224}Ra was adsorbed to solids after it was produced. However, Webster et al. (1995) found a strong dependence of Ra desorption on salinity and on sediment grain size such that fine-grained sediments in marine environments released much more Ra, although a significant fraction was bound inside crystal lattices where only slow diffusion or rock weathering would release it.

Below we briefly outline some applications and insights with Ra and Rn in salt marsh studies, which typically follow one of two approaches. (a) Box models are built on the steady state assumption, that what enters a control volume per unit time equals what exits. By evaluating removal processes and measuring most inputs, other inputs, specifically the discharging groundwater's contribution, may be estimated (Peterson et al., 2019). (b) Another approach is to use differences in activity ratios (AR's) between isotopes with different half-lives to reveal a deficiency of one isotope, implying a flux from one system to another. Radium AR's are also used to estimate the time a water mass has spent circulating in a coastal bay, that is, its water-mass age. These models can be based on either (a) a point or line input, after which the water is isolated from the source (W. S. Moore, 2000), or (b) a continuous input, for systems where the input is widespread over the study area (W. S. Moore, 2006), but may be insufficiently sensitive for water ages shorter than about 5 days, compared to typical uncertainty in the data (Knee et al., 2011). Radium studies have been used to differentiate among submarine groundwater sources (W. S. Moore, 2003; W. S. Moore et al., 2006; Schutte et al., 2020) and to test hydrodynamic models and the results of other tracers' studies (Breier et al., 2009; Crotwell & Moore, 2003; Rapaglia et al., 2010). More complete reviews of radionuclide techniques, models, and applications can be found in Burnett et al. (2006), Swarzenski (2007), M. Charette et al. (2008), P. Cai et al. (2012), Rodellas et al. (2017), Jiao and Post (2019), Taniguchi et al. (2019), Shi et al. (2019), and Guimond and Tamborski (2021).

In the first application of Ra isotopes to salt marshes, Bollinger and Moore (1984) measured three times more ^{224}Ra at the low tide compared to the high tide in the creek water surrounded by a salt marsh in North Inlet Estuary, South Carolina, USA. Porewater revealed a deficiency of ^{224}Ra relative to ^{228}Th in the upper 5 cm, implying a flux into the overlying water that a box model suggested would require 2–3 days to supply. Fresh, low-Ra groundwater input was omitted from the initial study by Bollinger and Moore (1984), but Rama and Moore (1996) subsequently found salty groundwater and high Ra activities in wells 1–3 m deep within the marsh. They concluded that the primary source of dissolved Ra in the tidal creek water was salty groundwater discharging directly into marsh creeks that had cut into the underlying saline aquifer. A model of the marsh ^{228}Ra and ^{226}Ra mass balances revealed that this saline aquifer discharge provided the major nutrient source to the marsh, and estimated the flux to possibly rival total riverine nutrient fluxes to the coastal ocean for the region (Krest et al., 2000).

Ties between saline salt marsh groundwater discharge and other constituent fluxes have also been estimated via Ra-constrained water budgets, for example, DIN (dissolved inorganic nitrogen) outwelling (M. A. Charette et al., 2001), also DIP (dissolved inorganic phosphorus), DIC (dissolved inorganic carbon), and DOC (dissolved organic carbon; W. S. Moore et al., 2006; Porubsky et al., 2014); vanadium and chromium discharge but molybdenum and uranium sinks (O'Connor et al., 2015); methylmercury (Shi et al., 2018) and methane outwelling (Schutte et al., 2020); and ^{226}Ra, DIC, and NH_4^+ release by oxidation of organic matter within the aquifer using SO_4^{2-} as the electron acceptor (W. S. Moore & Joye, 2021). Ra tracing has additionally proven useful, not only to detect and estimate saline groundwater and associated geochemical discharge volumes, but also to help illuminate drivers of such saline SGD, such as enhanced seawater and sediment interactions and groundwater discharges under drought conditions (M. A. Charette et al., 2003) or, conversely, high sea level during the summer caused by ocean heating (W. S. Moore, 2006, 2007, 2010; W. S. Moore & Joye, 2021). On the other hand, passage of a moderate-strength cyclone caused a pulse of Ra- and nutrient-poor creek water into the marsh aquifer to depths of at least 5 m, before pre-storm concentrations recovered in the aquifer within a few tidal cycles (A. M. Wilson et al., 2011).

However, the heterogeneity of marsh sediments and a range of salinity and residence times therein produce a wide variety of radium activities within even a small portion of a marsh or coastal aquifer groundwater system (Duque et al., 2019; Michael et al., 2011). Thus, selecting the appropriate "source aquifer" endmember for Ra-based coastal groundwater discharge models is problematic. One can select the highest measured activities to produce a conservative estimate, a statistical mean or median if the data set is robust, or solve multiple equations using different radionuclides or other tracers (Sanial et al., 2021). The task is simplified if there are correlations between Ra isotopes and other dissolved constituents; sadly, this is not always the case. Most models rest on the steady state assumption noted above and used in most of the above studies, which is tested only rarely.

3.3. Numerical Models

Early studies on salt marsh hydrology mostly utilized field measurements, when numerical models were simple, mainly being 0D or 1D and steady-state or with simple time-varying inputs. Later, 2D models prevailed and today still account for most of salt marsh groundwater simulations, typically with transient simulation. More recently, numerical models have been extended to 3D to explore the hydrological processes in salt marshes at a higher level of complexity (Table 1). Early simulation studies, and even most studies still today, focused on the effects of transient tidal fluctuations (simple sinusoidal tides, spring-neap tides or real conditions) on groundwater flow in marsh sediments (uniform, layered soils or with macropores). The hydraulic conductivities used in these numerical investigations range from 8.1×10^{-9} to 6.9×10^{-4} m s^{-1}. These values cover the typical soil property ranging from low-permeability clay loam to high-permeability sand (according to the statistics of Carsel and Parrish (1988)). Most models commonly omit freshwater recharge and evapotranspiration or consider only one of these two factors, as discussed below.

The various model types reviewed below all aspire to water balance closure, to match the change in water storage to the sum of the fluxes across boundaries. The variations of water storage in an unconfined groundwater system (e.g., salt marsh) are accompanied by water drainage via two physical and sometimes coexisting mechanisms: volumetric water change due to watertable fluctuations and elastic water release due to the deformation of the soil skeleton (i.e., soil compressibility; Figure 11). As the watertable falls, water in soil pores seeps out and de-saturation occurs to the extent described by effective porosity. Then air enters and occupies portions of the soil pores (Figure 11a). Elastic water release depends on the loading of the soil skeleton by overlying water (or barometric

Table 1
Summary of Used Numerical Models for Simulating Hydrology in Salt Marshes

Dimension	Tidal regime	Soil geology	Saturation regime	Time domain	Recharge	Evapotranspiration	Density-dependent flow	References
0D	Real	Uniform	Saturated	Transient	Precipitation	✓	None	Nuttle and Harvey (1995) and Morris (1995)
1D-X	Real	Uniform	Saturated	Transient	Precipitation	✓	None	Montalto et al. (2007) and Nuttle (1988a)
	Real	Uniform	Saturated	Transient	None	None	None	J. W. Harvey et al. (1987)
	Real	Uniform	Saturated	Steady-state	Precipitation	✓	None	Tobias et al. (2001)
	Spring-neap	Uniform	Saturated	Transient	None	None	None	Xin et al. (2010)
1D-Z	Real	Uniform	Saturated	Transient	None	None	None	Hemond et al. (1984)
	Real	Uniform	Saturated	Steady-state	Inland freshwater + precipitation	✓	✓	J. W. Harvey and Odum (1990)
2D-XZ	Real	Uniform	Unsaturated	Transient	None	✓	None	Hemond and Fifield (1982)
	None	Uniform + macropores	Unsaturated	Steady-state	None	✓	✓	T. Zhou et al. (2020)
	Real	Layered + macropores	Unsaturated	Transient	Precipitation	✓	None	C. E. Hughes et al. (1998)
	Real	Uniform	Unsaturated	Steady-state	Precipitation	None	✓	Carter et al. (2008)
	Real	Layered	Unsaturated	Transient	None	None	None	Xia and Li (2012)
	Real	Layered	Unsaturated	Transient	Precipitation	✓	None	Xiao et al. (2017)
	Sinusoidal	Uniform	Saturated	Transient	None	None	None	L. R. Gardner (2005a), L. R. Gardner (2005b), and L. R. Gardner and Wilson (2006)
	Sinusoidal	Uniform	Unsaturated	Transient	None	✓	None	Ursino et al. (2004) and Marani et al. (2006)
	Sinusoidal	Uniform	Unsaturated	Transient	None	None	None	L. R. Gardner and Wilson (2006), A. M. Wilson and Gardner (2005), Wilson and Gardner (2006), and A. M. Wilson and Morris (2012)
	Sinusoidal	Uniform	Unsaturated	Transient	None	None	✓	Shen et al. (2015) and Shen et al. (2016)
	Sinusoidal	Layered	Unsaturated	Transient	None	None	None	Gardner (2007) and Wilson and Morris (2012)
	Sinusoidal	Layered	Unsaturated	Transient	Inland freshwater + precipitation	✓	✓	Xiao, Wilson, et al. (2019)
	Sinusoidal	Layered	Unsaturated	Transient	Inland freshwater	None	✓	Xiao, Li, et al. (2019)
	Sinusoidal	Uniform	Unsaturated + Air	Transient	None	✓	None	H. Li et al. (2005) and Cola et al. (2008)
	Sinusoidal	Uniform	Unsaturated	Transient	Precipitation	✓	✓	H. Wang et al. (2007)
	Spring-neap	Uniform	Unsaturated	Transient	None	None	None	Xin et al. (2010)
	Spring-neap	Uniform	Unsaturated	Transient	Precipitation	✓	✓	Xin et al. (2017)
	Spring-neap	Uniform	Unsaturated	Transient	None	✓	✓	Shen et al. (2018)

Table 1
Continued

Dimension	Tidal regime	Saturation regime	Soil geology	Time domain	Recharge	Evapotranspiration	Density-dependent flow	References
3D-XYZ	Sinusoidal	Unsaturated	Layered + macropores	Transient	None	None	None	Xin et al. (2009)
	Sinusoidal	Unsaturated	Layered + macropores	Transient	None	✓	None	Xu et al. (2021)
	Real	Unsaturated	Uniform	Transient	None	✓	None	Moffett et al. (2012)
	Real	Unsaturated	Layered	Transient	None	✓	None	Moffett et al. (2012) and H. Zhang, Moffett, et al. (2014)
	Real	Unsaturated	Layered	Transient	Precipitation	✓	None	Y. Zhang et al. (2018) and Guimond, Yu, et al. (2020)
	Sinusoidal	Unsaturated	Uniform	Transient	None	None	None	Xin et al. (2011)
	Sinusoidal	Unsaturated	Layered	Transient	None	None	None	Xin et al. (2012)
	Spring-neap	Unsaturated	Layered	Transient	None	None	None	Xin, Li, and Barry (2013)

Note. X: direction perpendicular to tidal creek, Y: direction parallel to tidal creek, Z: vertical direction.

pressure (Figure 11b). Overall, the amount of water release via volumetric drainage is 3–4 orders of magnitude higher than that via elastic release (Jacob, 1950). Earlier salt marsh models considered only pore drainage and did not include tidal loading and elastic water release; for example, this mechanism was added to a popular SUTRA model in the year 2000 (Reeves et al., 2000).

3.3.1. 0D Models

0D groundwater models consider a discrete volume of marsh sediment. Variations in the hydraulic head and changes in water storage are linked to fluxes across the boundaries. Generally, there are three components of a water budget for a salt marsh (Equation A1), including (a) net flux across the marsh surface by water infiltration and/or evapotranspiration, (b) horizontal flux between the marsh sediments and tidal creeks (drains, boundary conditions) or adjacent sediment units (e.g., uplands), and (c) groundwater exchange between a targeted marsh sediment layer and an underlying aquifer (Equation A1; Morris, 1995; Nuttle & Harvey, 1995). 0D models were mainly used to investigate the water and solute fluxes in marsh sediments at a regional scale. However, due to the highly simplified dimensionality and soil geometry, 0D models cannot represent the spatial variability and multiscale hydrological processes in salt marshes (Moffett et al., 2012).

3.3.2. 1D Models

1D groundwater models are also simple but can account for some spatial variability of salt marsh hydrology. They are developed based on either the classic Darcy's Law, for example, for saturated flow in the vertical direction (Hemond et al., 1984; J. W. Harvey & Odum, 1990; Tobias et al., 2001), or the Boussinesq equation in the horizontal direction with the Dupuit-Forchheimer assumption (Equations A2 and A3; J. W. Harvey & Odum, 1990; Montalto et al., 2007; Nuttle, 1988a; Xin et al., 2010). Generally, the Dupuit-Forchheimer assumption means that in a shallow unconfined aquifer (e.g., a salt marsh), a hydraulic gradient drives groundwater to flow horizontally and the drainage is proportional to the saturated aquifer thickness (i.e., the distance between the local watertable and impermeable base). Xin et al. (2010) later developed a modified-1D Boussinesq model with added vertical flow effects and/or capillary effects for situations in which the Dupuit-Forchheimer assumption does not apply.

According to the research purposes, the 1D models can be divided into two types (vertical and horizontal). The first type concerns vertical infiltration through the marsh surface (Hemond et al., 1984; Hemond & Fifield, 1982; J. W. Harvey & Odum, 1990). The second type simulates the horizontal propagation of the tidal signal away from the tidal creeks and drainage toward the creeks during ebb tides (J. W. Harvey et al., 1987; Montalto et al., 2007; Nuttle, 1988a; Tobias et al., 2001; Xin et al., 2010). For 1D models, the groundwater exchange induced by watertable fluctuations significantly depends on the effective porosity.

Most investigations using 1D models consider only simplified conditions, for example, fully saturated flow, homogeneous marsh sediment (Table 1). Early models' assumptions of saturated flow processes are particularly rough estimates. When the tidal level falls below the marsh surface, a vadose zone (including unsaturated zone and capillary fringe) develops, which later disappears during the next tidal inundation (Figure 11a). The vadose zone stores a considerable amount of water, or may even be fully tension-saturated in fine-textured soils (i.e., capillary fringe), and affects the watertable dynamics and overall behavior of the shallow unconfined aquifer through water exchange with the underlying aquifer (Kong et al., 2015). Moreover, most salt marshes are layered (L. R. Gardner, 2007) and/or have many macropores (e.g., crab burrows; Xin et al., 2009). These aspects can not be captured by 1D models.
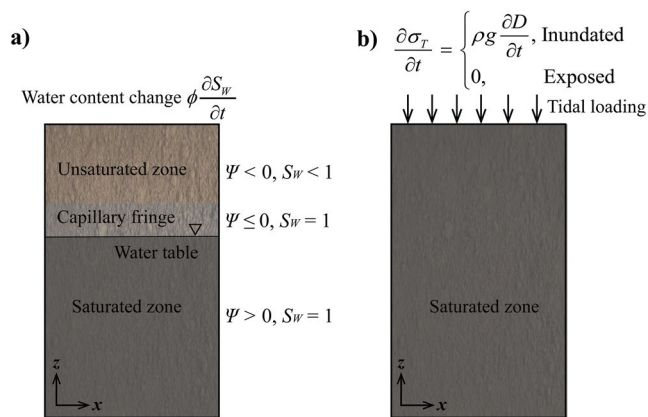
a)

b)
$$\frac{\partial \sigma_T}{\partial t} = \begin{cases} \rho g \dfrac{\partial D}{\partial t}, & \text{Inundated} \\ 0, & \text{Exposed} \end{cases}$$

Figure 11. Schematic diagram of water content change (a) and tidal loading (b) in salt marsh sediments. Symbols: ψ is the pressure head, S_W is the soil water saturation, ϕ is the porosity, σ_T is the total stress, g is the magnitude of gravitational acceleration, ρ is the fluid density, D is the local surface water depth, and t is the time.

3.3.3. 2D Models

Most of 2D marsh models focus on vertical sections perpendicular to tidal creeks (A. M. Wilson & Gardner, 2006; L. R. Gardner, 2005a, 2005b; Shen et al., 2015; Ursino et al., 2004; Xin et al., 2009). This is still the most common type of model domain in this research area today. Most of these 2D models are based on the Richards' equation and so are capable of simulating the transient and variably saturated flow (A. M. Wilson & Gardner, 2005, 2006; Xin et al., 2010), exceptions being the models of L. R. Gardner (2005b), L. R. Gardner (2005a), and L. R. Gardner and Wilson (2006), which only consider fully saturated flow and are based on the boundary integral method and, alternatively, the fully two-phase model of H. Li et al. (2005) that considers coupled and separate air and water movement in the sediment pores. Apart from variably saturated flow, some 2D models include more complexities, for example, sediment stratification and/or macropores (A. M. Wilson & Morris, 2012; C. E. Hughes et al., 1998; Xiao, Wilson, et al., 2019; Xin et al., 2009), evapotranspiration (H. Wang et al., 2007; Marani et al., 2006; Shen et al., 2018; T. Zhou et al., 2020; Ursino et al., 2004; Xin et al., 2017), precipitation (Xin et al., 2017), inland freshwater input (Xiao, Li, et al., 2019), or density-dependent flow (Shen et al., 2015, 2016, 2018; Xiao, Li, et al., 2019; Xin et al., 2017). Water retention condition in the unsaturated zone is described by the van Genuchten (1980) or the W. R. Gardner (1958) functions, which quantify the pressure-dependent soil hydraulic conductivity and soil saturation.

For 2D models based on Richards' Equation (e.g., Voss & Provost, 2008), one noteworthy point is the need for a tidal loading modification. Richards' Equation assumes constant total stress (σ_T) on the porous medium (Figure 11b). Under this assumption, during inundation, fluctuations of groundwater pressure (P) with varying depths of overlying tidal water would result in changes of effective stress (σ_e), thereby generating numerically an water release through the elastic marsh sediment. In fact, the total stress on marsh sediment is not constant and likely changes in a similar way to the groundwater pressure, so the effective stress is likely invariant. Therefore, as suggested by Reeves et al. (2000) and A. M. Wilson and Gardner (2006), a tidal loading term should be incorporated into the Richards' Equation to account for the total stress change. However, the numerical tests of Xin et al. (2009) showed that once the saturated hydraulic conductivity of marsh sediment is greater than 1.0×10^{-6} m s^{-1}, water release via volumetric drainage is much larger than that via elastic release. Under this condition, an alternative approach for dealing with the total stress variation is to ignore the compressibility terms in the modified Richards' Equation (see Equations A4 and A5 for more information).

When simulating groundwater flow in a creek-perpendicular marsh section, tidal fluctuations lead to a varying sediment-water interface (e.g., moving boundary). During falling tide, the watertable in the marsh soil becomes decoupled from the tidal level. Consequently, a seepage face will form and develop along the sloping marsh-creek bed interface (A. M. Wilson & Gardner, 2006). To deal with such a moving boundary, below the creek water surface submerged boundary nodes are prescribed by hydrostatic pressure according to the local water depth given by the tidal level. For exposed boundary nodes above the tidal level, three conditions are considered depending on the aim and structure of the model: (a) if the nodes were saturated in the previous time step, they may be treated as seepage face nodes and assigned an atmospheric pressure, (b) if the nodes were unsaturated in the last time step, they may be prescribed to be no-flow (A. M. Wilson & Gardner, 2006), and (c) or the exposed nodes may be prescribed a specified flux to represent an infiltration or evapotranspiration rate.

3.3.4. 3D Models

Most 2D models assume a flat marsh surface or a slight slope toward the tidal creek and ignore other topographic variations. Natural salt marshes commonly exhibit strong three-dimensionality, with a meandering and complex tidal creek network (B. R. Evans et al., 2019) embedded in a 3D spatially heterogeneous marsh system. Compared to modeling of lower dimensions (e.g., 0D, 1D, and 2D), 3D numerical investigations of salt marsh hydrology are relatively rare being more technically challenging and computationally expensive.

3D simulations of the hydrological processes in salt marshes require the coupling of a surface hydrodynamic model and a groundwater model to be able to predict where the surface water wetting front will be at a given time, as it cannot be simply prescribed over complex terrain. These two models are coupled at the common SW-GW interface and solved in a combined fashion. SW-GW exchange is driven by the hydraulic head difference at the sediment-water interface. The surface water component of the 3D models is usually described using models based on the depth-averaged shallow water equations, which consist of continuity and momentum equations. A typical surface hydrodynamic model is ELCIRC, which uses a finite-volume/finite-difference, Eulerian-Lagrangian method to solve the 3D shallow water equations (Y. Zhang & Baptista, 2004). Regarding the groundwater component of the 3D coupled models, there are two common types. The first type is based on the Boussinesq equation, which describes the saturated flow only and is so incapable of simulating the complex porewater flow in partially saturated marsh soils. The second type (e.g., SUTRA (Voss & Provost, 2008)) is built upon Richards' Equation for variably saturated flow; as this is more realistic, this review only includes 3D models that couple the shallow water equations for surface water flow to the Richards' Equation for variably saturated subsurface flow. Some examples of such 3D models are ELCIRC-SUTRA (Yuan et al., 2011) and HydroGeoSphere (Therrien et al., 2006).

Other researchers further extended the above-mentioned 3D models by developing regional-scale, process-based, distributed hydrological models, such as PIHM-Wetland (Y. Zhang et al., 2018). However, most studies using the aforementioned 3D models have so far only considered the hydrodynamics and neglected solute transport processes, which would multiply the degrees of freedom being solved for.

4. Impact of Hydrology and Competition on Plant Ecology

The zonation of salt marsh plants is influenced by a variety of biotic and abiotic factors (Armstrong et al., 1985; Vince & Snow, 1984; Wyatt, 1992). Based on a global review (Figure 12), overland tidal flow (flooding period and frequency), topography and subsurface abiotic factors (soil moisture, oxygen availability, groundwater salinity and nutrient status) along environmental gradients have a major role in the zonation of salt marsh plants
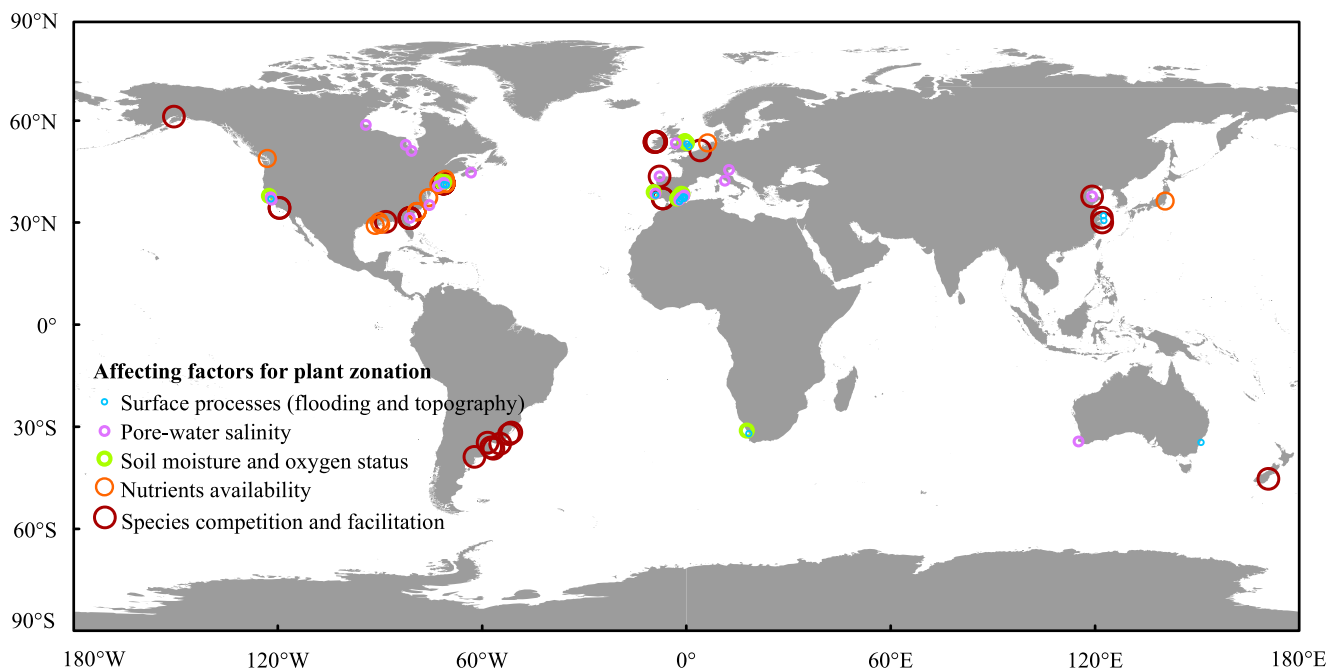
Figure 12. Global case studies on abiotic and biotic factors affecting salt marsh plant zonation. The circles mean the primary control factor of plant zonation pattern for most species. We conducted a broad search of peer-reviewed articles collected from the Web of Science core collection (1963–2020) using the following keywords: TS = brackish marsh or salt marsh and TS = plant zonation or plant distribution. These keywords generated 3,263 papers. To avoid any bias in publication selection, the following basic criteria were followed to select the relevant studies: (1) The selected experiments were conducted in the field, (2) study duration was at least one growing season, and (3) the data provided in selected papers include at least one of the soil physico-chemical characteristic, biotic index or water characteristic. Finally, a total of 71 papers (with 104 monitoring sites) were used in the analysis.

(Marani et al., 2013; Moffett et al., 2012; Wyatt, 1992). In contrast, biotic factors such as interspecific competition, transpiration and nutrient uptake mostly have local effects on plant zonation in the given marsh environments (Pennings et al., 2005). However, potential sampling bias in this distribution cannot be ruled out, and may even be likely due to the global geographic distribution of selected research sites in a given marsh-related field (e.g., hydrology vs. ecology) or relative number of publications from certain regions or field areas.

4.1. Effects of Tidal Flow and Topography

Salt marshes are regularly flooded by tidal water and their marsh plant communities may be arrayed in topographic zonation patterns that are submerged to varying extents along the main topographic gradient (Bertness & Ellison, 1987; D'Odorico et al., 2013; Moffett et al., 2012; Xie et al., 2020). Salt marsh plants help to promote sediment deposition and geomorphological stabilization to mitigate the intertidal disturbance for other species to reside on the raised mudflat surface (Caçador et al., 2007). Numerous studies have aimed to determine which hydrological factors affect plant zonation (Alvarez-Rogel et al., 2007; A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015; Bertness & Ellison, 1987; Hladik & Alber, 2014; Moffett, Robinson, & Gorelick, 2010; Sanchez et al., 1996; Silvestri et al., 2005; Snow & Vince, 1984; Zedler et al., 1999). Flooding and elevation of the tidal range are identified as the primary factors that affect plant survival and consequently as the physical basis of the distinctive zonation of marsh plant communities (Bockelmann et al., 2002; Johnson & York, 1915; Sanchez et al., 1996; Silvestri et al., 2005).

Flooding establishes a prominent abiotic stress for the salt marsh plants even if they are salt-adapted, with soil anoxia and waterlogging gradually decreasing from the mudflat edge to the terrestrial upland (Cui et al., 2011; Mendelssohn et al., 1981; Woerner & Hackney, 1997; Zedler et al., 1999). Flooding generally regulates the amount of vegetation cover by limiting the plant growth, inducing the anaerobiosis of substrate, or triggering the toxic effects of salt. This effect is generally species-specific; for instance, *Sedges* perform best in water-saturated soil and optimal growth occurs in half-flooded soil, whereas other species (e.g., *Phragmites*) often prefer well-drained conditions (Ge et al., 2015; Pennings & Callaway, 1992; Silvestri et al., 2005). A global meta-analysis by Balke et al. (2016) showed that the height difference between the lowest elevation of salt marsh pioneer vegetation and mean high water increases logarithmically with tidal range when including macrotidal salt marshes. The potentially vegetated section of the tidal frame below mean high water does not proportionally increase with tidal range. Hence, flooding frequency (also tidal range) is extensively recognized as the more important factor to define the elevational limit of marsh plants, relative to inundation duration (Álvarez Rogel et al., 2001; Armstrong et al., 1985; Bertness & Ellison, 1987; Bockelmann et al., 2002; Sanchez et al., 1996).

Flooding frequency changes with marsh elevation such that minor changes in marsh elevation could lead to major changes in flooding frequency. Particularly within low-elevation marshes, small differences in elevation are sufficient to alter the intertidal environment (Brereton, 1971; Davy et al., 2011; Zedler et al., 1999). An increase in marsh elevation often causes a decrease in flooding frequency and duration and results in increases in redox potential (due to oxidation) and soil salinity (due to evaporation). However, the interaction between flooding frequence and salinity mediating plant zonation in salt marshes may change geographically. In some cases, spatial variation in geology and hydrology driving spatial variation in flooding and salinity rather than variation in the general mechanisms affecting plant distribution and growth (Pennings et al., 2005).

Tidal creeks in intertidal habitat also fundamentally regulate plant zonation (Sanderson et al., 2001). Tidal channel networks strongly control flooding flow and drainage, and thereby structure the vegetation distribution and community composition in salt marshes (Morzaria-Luna et al., 2004; Sanderson et al., 2000). Tidal channels also facilitate seed dispersal and seedling establishment and affect biodiversity of species close to channel banks (Rand, 2000).

4.2. Effects of Soil Moisture and Oxygen Availability

Relationships between salt marsh plant zonation and soil moisture have been well studied (Bornman et al., 2008; Gonzalez-Alcaraz et al., 2014; Moffett et al., 2012; Rogel et al., 2000; St. Omer, 2004). For instance, Bornman et al. (2008) concluded that soil moisture had the greatest influence on the distribution of the dominant salt marsh species along the arid west coast of southern Africa. Here, the low rainfall and irregular occurrence of advected sea fog increase the importance of a shallow saline watertable in the floodplain as a source of moisture during

dry periods. In a semiarid Mediterranean climate, soil moisture and soil ionic composition (e.g., K^-/Na^-, Ca^{2-}/Na^-, and Ca^{2-}/Mg^{2-} balances) were the most important edaphic gradients related to plant zonation in the salt marshes (Rogel et al., 2000). Soil moisture could regulate the spatio-temporal variations in cation adsorption and ionic balance due to degrees of soil's exchange complex and salt solubility within different types of clay mineral in salt marshes.

Gradients in oxygen availability are among the most important environmental parameters influencing zonation in salt marsh communities. Wijte and Gallagher (1996) tested the effects of oxygen availability on the germination of two widely distributed salt marsh grasses, *Spartina alterniflora* and *Phragmites australis*. The germination rates of both species were reduced at low oxygen. When a low-elevation mudflat is colonized by vegetation, the germination responses of the marsh species to hypoxia are important in establishing their initial zonation.

Sediment redox potential also affects plant growth and the consequent zonation patterns in salt marshes (Caçador et al., 2007). Redox potential generally decreases with increasing tidal flooding and decreasing tidal flat elevation, and low redox potential could inhibit the growth of some marsh halophytes (Armstrong et al., 1985; Bertness & Ellison, 1987; Davy et al., 2011; Howes et al., 1981; Mendelssohn et al., 1981). Nevertheless, some marsh halophytes have the ability to oxygenate their rhizospheres (via aerenchyma) and to tolerate low redox potential (Colmer & Flowers, 2008). For instance, Davy et al. (2011) reported that on the North Norfolk coast (UK) the effects of sediment redox potential were distinguishable from the marsh species, showing that *Suaeda maritima* can tolerate a wide range of elevations but was absent from areas with low redox potential; and *Puccinellia maritima* favored low redox potentials independently of elevation. In contrast, *Salicornia europaea* occurred predominantly at lower elevation and was unsensitive to redox potential; *Atriplex portulacoides* was more averse to low redox potential than to low elevation; and *Elytrigia atherica* was restricted to both high redox potential and high elevation.

4.3. Effects of Groundwater Salinity

Understanding the role of groundwater salinity in determining plant zonation patterns is particularly important in salt marshes (Pennings et al., 2005). Geographic, tidal, and climatic variations may induce different salinity patterns at different marsh elevations or at the same marsh elevations in different global regions (Cui et al., 2011; Pennings et al., 2005). Most studies show that porewater salinity and flooding gradients are not parallel along the marsh elevation (Álvarez Rogel et al., 2001; Angiolini et al., 2013; Pennings et al., 2005; Sanchez et al., 1996; Silvestri et al., 2005).

High porewater salinity correlates with an increase in concentration of sulfate ions, which can be converted to hydrogen sulfide in low-redox marsh sediments (Alldred et al., 2017). The toxic effects of sulfide on plants lead to expected decline in root growth as salinity increases in anoxic marsh sediments. Salinity stress probably plays a much more important role in mediating plant zonation patterns at lower latitudes (Pennings et al., 2005). Some marsh plants can tolerate a wide range of salinities and colonize areas from high marsh flats to low-elevation zones (Bertness et al., 1992; Caçador et al., 2007; Emery et al., 2001; Engels et al., 2011; Pennings et al., 2005; Sanchez et al., 1996). The most extreme evidence of geographical variation on the role of salinity in mediating plant patterns lies in the occurrence of salt pans in low-latitude marshes, where soil salinity exceeds levels that plants tolerate (Pennings et al., 2005; Pielou & Routledge, 1976).

Pennings et al. (2005) suggested that the nature of ecological interactions is likely to vary geographically because of variations in the physical environment, and these variations must be taken into account in order to successfully generalize the results of field studies across geographical scales. Moffett, Robinson, and Gorelick (2010) highlighted that the combined effects of porewater salinity, soil moisture, and sediment texture on plant zonation should be examined as intertidal soil moisture and salinity change significantly over the short tidal cycle time. The roles of spatial patterns of soil salinity, moisture and topography on plant zonation shows that multiple abiotic variables exhibit different relationships with vegetation, and combinations of soil variables regulate the vegetation zones and habitats of individual plant species (Moffett, Robinson, & Gorelick, 2010). Additional field observations showed that salt marsh plants such as *S. alterniflora* grow taller near tidal creeks than in the interior areas, as near-creek zones are commonly characterized by low water moisture and porewater salinity (see Section 2; A. M. Wilson, Evans, Moore, Schutte, Joye, et al., 2015; Howes & Goehringer, 1994; Howes et al., 1981; Mendelssohn et al., 1981; Silvestri et al., 2005).

Furthermore, Bertness and Pennings (2000) interpreted the spatial variation in process and pattern in salt marsh plant communities in eastern North American, showing that the plant zonation was strongly influenced by both climate and salinity. In low-latitude salt marshes, a hot climate leads to elevated salinities at upper/middle elevations, and marsh zonation patterns are strongly driven by soil salinity patterns. In contrast, in high-latitude salt marshes, the climate is much cooler, and thus porewater salinity plays only a minor role in maintaining marsh plant distributional patterns. The difference in porewater salinity between low- and high-latitude salt marshes may generate variations in plant organization and zonation.

4.4. Effects of Nutrient Status

Many salt marshes are nutrient-limited, especially in the pristine or undisturbed systems, because mineral sediments provide little nutrition to support plant growth (Kachi & Hirose, 1983). However, compared to the pre-industrial era, human activities have remarkably enhanced the nutrient loadings to coastal areas. During recent decades, coastal ecosystems have received excessive nutrients from acid precipitation, sewage waste, and fertilizer runoff with nitrogen (N) and phosphorus (P) loads. With nutrient input from terrestrial groundwater and riverine supply, salt marshes became the most nutrient enriched ecosystem around the world (Pardo et al., 2011; Seitzinger et al., 2005).

Confusingly, investigations based on gradient-exposure studies and on fertilization experiments have reported conflicting responses of salt marshes to nutrient enrichment. More nutrients can disproportionately provide positive benefits in coarse mineral soils (Wigand et al., 2015). Plants generally allocate a lower proportion of their growth to belowground biomass because they require fewer roots to acquire the nutrients for growth of photosynthetic tissues aboveground (Ericsson, 1995). Marsh biomass growth, elevation change, and sediment cohesive strength play key roles in determining salt marsh resilience to harsh tidal environment or future sea level rise. Reductions in root mass are often accompanied by a decrease in sediment stability, as observed under high nutrient enrichment (Deegan et al., 2012; Kearney et al., 2011; Turner et al., 2009; Watson et al., 2014). However, many experiments and physiological models show that moderate nutrient enrichment can enhance growth of both aboveground and belowground biomass of marsh plants, resulting in increased sediment capture due to greater biomass and stem density (Anisfeld & Hill, 2012; Darby & Turner, 2008; Fox et al., 2012; Graham & Mendelssohn, 2014; Morris et al., 2002). Morris, Shaffer, and Nyman (2013) also reasoned that an increase in belowground biomass with nitrogen fertilization is expected if the increase in total growth compensates for lower relative allocation of growth to roots and rhizomes, because increasing nutrient availability is associated with a decrease in the root-to-shoot ratio. Increased nutrient loads will also alter salt marsh structure and trophic level and interact strongly with other abiotic factors in tidal environment. It is difficult to determine conclusively whether nutrient enrichment positively or negatively impacts salt marshes (Mozdzer et al., 2020).

The uncertain function of nutrient enrichment might relate to the varying roles of nutrient form, sediment type, soil structure and porosity, and redox and salinity gradients in affecting salt marshes (Mozdzer et al., 2020). For instance, different N species (e.g., nitrate vs. ammonium) and N-P-K (nitrogen-phosphorus-potassium) ratios may induce different responses in salt marshes (Mozdzer et al., 2011). The wide range of mineral components in tidal sediments with varying permeability, bulk density and porosity, and redox status are likely to affect differently nutrient maintaining and plant growth (Mozdzer et al., 2020). The response of salt marshes to nutrient enrichment is also related to salinity regimes and nutrient biogeochemical processes. Salinity appears to inhibit uptake of ammonium by marsh plants and facilitates release of adsorbed ammonium from sediment surfaces by ion exchange (Bradley & Morris, 1990; Giblin et al., 2010), resulting in an accumulation of ammonium in porewater. A buildup of ammonium or higher N inputs may inhibit root development and growth of plants, thereby destabilizing coastal wetlands. However, Alldred et al. (2017) found that belowground biomass of marsh plants was related positively to salinity and negatively to extractable nitrogen content in sediments, and no significant interaction between salinity and nitrogen was found. This indicated that salinity and nutrients were independently related to salt marsh plants.

4.5. Species Competition and Facilitation Regulated by Environmental Conditions

The distribution and zonation of salt marsh plants are not controlled solely by physiological restrictions under geomorphological and geochemical conditions in intertidal environments. Biotic interactions of competition and

facilitation are of particular importance in plant zonation (D. Wang et al., 2021; Sanchez et al., 1996). Competition is to preempt limiting resources such as light, water and nutrients, that determine rates of plant growth and colonization. Facilitation favors neighboring plants to enhance the survival, growth, and/or reproduction. The importance of biotic interactions in plant zonation along the tidal flooding gradient have been well illustrated (Castillo et al., 2000; Costa et al., 2003; Levine et al., 1998; Pennings & Callaway, 1992; Snow & Vince, 1984). Plant-plant competition is often observed to mediate the spatial distribution of plants, and facilitation is important in mediating plant performance and diversity in marshes with varying flooding and salinity stress (Bertness & Ellison, 1987; Bertness & Hacker, 1994; Callaway et al., 1990; Engels et al., 2011; He et al., 2011; Pennings & Callaway, 1992). Sometimes, extremely stressful environments can surpass the physiological stress tolerance range of salt marsh plants (Holmgren & Scheffer, 2010).

Tilman's Resource-Ratio hypothesis states that competition for nutrients occurs when there is a low nutrient supply belowground. Whereas aboveground, competition for light might be more important (S. D. Wilson & Keddy, 1985; S. D. Wilson & Tilman, 1993). The nutrient-dependent competitive hierarchy leads to trade-offs between belowground and aboveground competitive abilities, and thus nutrient supply may significantly affect the competitive dynamics between salt marsh perennials and their resultant zonation across an environmental gradient (S. D. Wilson & Keddy, 1985; S. D. Wilson & Tilman, 1993). On the other hand, the competitive ability of a species is also determined by its biomass allocation strategy according to resource availability along a productivity gradient.

The Competition-to-Stress hypothesis (Guo & Pennings, 2012) suggests that some marsh species are excluded from high-salinity regions because of abiotic stress, or are excluded from low-salinity regions by competition (Crain et al., 2004; Engels & Jensen, 2010). Transplantation experiments showed that salt marsh plants performed well in brackish marshes if neighbors were absent, whereas brackish marsh plants died when transplanted into higher salinity habitats. The Stress-Gradient hypothesis predicts that the importance or intensity of competition and facilitation would change along abiotic stress gradients, with facilitation being more important in high abiotic stress conditions (Bertness & Hacker, 1994). At lower levels of abiotic stress, competition and consumer pressure become increasingly important determinants of species distribution. Although the Stress-Gradient hypothesis predicts well latitudinal patterns of facilitation, it omits differences in intraspecies and interspecies salt tolerance among plants (Pennings et al., 2003). Pennings et al. (2003) noted that intraspecific adaptations and community structure and processes are likely to vary geographically. Generally, the salt-sensitive species are likely to be facilitated by neighbors, whereas the salt-tolerant species are unlikely to benefit substantially from neighbor amelioration of soil salinities.

The trade-off between stress tolerance and nutrient competition is identified as one of the major drivers of salt marsh plant zonation (Bertness & Ellison, 1987; Daleo et al., 2008; Guo & Pennings, 2012). According to Bertness et al. (1992) and Pennings et al. (2001), the competitive dominants are typically unable to survive in physically harsh conditions (e.g., frequent flooding and high salinity), while competitively subordinate and stress-tolerant plants are seen only in more stressful habitats because they are displaced from benign habitats by dominant competitors. If nutrient supply also controls competitive dominance across physical gradients in salt marshes, it may dictate a trade-off between competitive ability and stress tolerance underlying plant zonation (Bertness et al., 1992; Guo & Pennings, 2012; Levine et al., 1998; Snow & Vince, 1984).

4.6. Feedback From Marsh Plants on Hydrology and Biogeochemistry

Salt marshes with high aboveground and belowground plant biomass notably modify the hydrological and morphodynamic conditions in the coastal area (Bouma et al., 2005; C. D'Alpaos et al., 2006; Temmerman et al., 2007). The effect of plants on hydro-morphodynamics in salt marshes is well known on the individual and patch scales (Järvelä, 2002; Siniscalchi et al., 2012). Submerged plants slow tidal flow and velocity, increase the bottom shear stress and dampen wave energy, thus promoting deposition of suspended sediments (Bouma et al., 2005; Neumeier & Amos, 2006). The wave attenuation capacity of plants depends on individual morphological properties and horizontal and vertical spatial canopy structure (Möller et al., 2014; Rupprecht et al., 2017; Ysebaert et al., 2011). In this way, the existence of salt marshes is beneficial to maintain tidal geomorphology and guarantee the safety of the coastal ecosystems (Augustin et al., 2009; Möller et al., 2014; Osorio-Cano et al., 2019). Plant roots can regulate soil properties (permeability, porosity, and heterogeneity) of the shallow sediments. While a summary of feedback among plant hydrology and morphodynamics is beyond the scope of this review, we highlight here

that these linked processes also have the potential to affect SW-GW interactions in salt marshes, but these effects are poorly understood.

Salt marshes are important sources or sinks of biologically important nutrients such as carbon (C) and nitrogen (N) in coastal ecosystems, influencing biogeochemical cycling (Tobias & Neubauer, 2019). Marsh macrophytes contribute autochthonous organic matter to the ecosystem, and marsh geomorphology and hydrodynamics lead to trapping of allochthonous organic and mineral particulates. CO_2 uptake (or primary production) by marsh plants generally exceeds respiration (CO_2 emission), leading to their capacity for producing excess organic C and acting as CO_2 sinks (C. M. Duarte et al., 2013; Mcleod et al., 2011). The ability of vegetated coastal habitats to trap particles from the water column and store them in the soil results in a high burial rate of organic C in salt marshes, which exceeds corresponding rates of terrestrial forest soils by 30–50 fold (Mcleod et al., 2011). However, Vázquez-Lule and Vargas (2021) recently indicated that CO_2 and CH_4 emissions during senescence and dormancy of a coastal marsh overshadow annual carbon uptakes, showing an overall net C source from the marsh to the atmosphere.

N pollution is one of the dominant drivers of global change affecting ecosystems. Due to application of synthetic fertilizer in agriculture, increased runoff, and human population growth in coastal areas, N pollution has increased in recent decades and poses one of the greatest threats to coastal marshes (Cloern, 2001; Howarth & Marino, 2006). N enrichment of coastal and estuarine systems alters biogeochemical cycles, resulting in disruptive or harmful blooms of phytoplankton and macroalgae, threating food webs and biodiversity (Nelson & Zavaleta, 2012). Coastal salt marshes can efficiently intercept watershed-derived N and other pollutants before they reach the ocean (Brin et al., 2010; Nelson & Zavaleta, 2012), with main pathways of plant uptake into tissue, denitrification by microbial communities, and burial in sediments (Bianchi, 2007; Seitzinger & Kroeze, 1998). N uptake by marsh plants is one of the larger N flux terms ranging from 1 to 33 g N m^{-2} y^{-1}, based on a wide global research data (Hopkinson & Giblin, 2008; Rozema et al., 2000; Tobias & Neubauer, 2019).

5. Impact of Surface Water and Groundwater Interaction on Coastal Biogeochemistry

SW-GW interactions facilitate biogeochemical cycling of carbon and nutrients in salt marshes. Organic matter in marsh sediments can be degraded and transformed by complex biogeochemical processes such as oxic respiration, denitrification, manganese or iron reduction, sulfate reduction and methanogenesis. Then groundwater flow can release soil carbon and nutrients to surface water and eventually to the sea (i.e., outwelling). These large carbon and nutrient fluxes have a significant impact on coastal biogeochemistry.

5.1. Carbon Outwelling

Salt marshes are primary production hotspots that accumulate sediment and carbon at high rates (Alongi, 2020; D. Wang et al., 2021). Part of the carbon accumulated in salt marsh sediments can be flushed out via shallow porewater exchange and deep groundwater flow, eventually reaching the coastal sea (Figure 13). The lateral transport of carbon and nutrients from salt marshes and other coastal ecosystems is referred to as outwelling. Discussions of the impact of carbon and nitrogen outwelling from salt marshes on primary productivity in coastal waters began in the 1960s based on observations on the US East Coast (Odum, 1968, 1980; Teal, 1962). Growing interest in salt marshes as carbon sequestration hotspots is reflected in investigations of carbon budgets (Santos, Burdige, et al., 2021). Outwelling functions as a long-term carbon sequestration mechanism that enables the storage of both particulate and dissolved carbon in the ocean. For example, DIC outwelling can store carbon as dissolved bicarbonate in the oceans.

Most investigations of salt marsh carbon fluxes, including outwelling, have focused on the US East Coast (Neubauer & Anderson, 2003; Raymond et al., 2000; Tamborski et al., 2021; Z. A. Wang & Cai, 2004) with a few examples from Canada's West Coast (Diggle et al., 2019), Australia (Santos et al., 2019; Webb et al., 2019) and China (Liu et al., 2021). Quantifying carbon and nutrient outwelling requires an understanding of water transport in and out of salt marshes or an assessment of water flushing times (W. S. Moore et al., 2006). For example, outwelling was quantified using high-frequency current measurements (Bogard et al., 2020; Chu et al., 2018) and natural tracers such as radium and radon isotopes (Santos et al., 2019; Tamborski et al., 2021). There are multiple
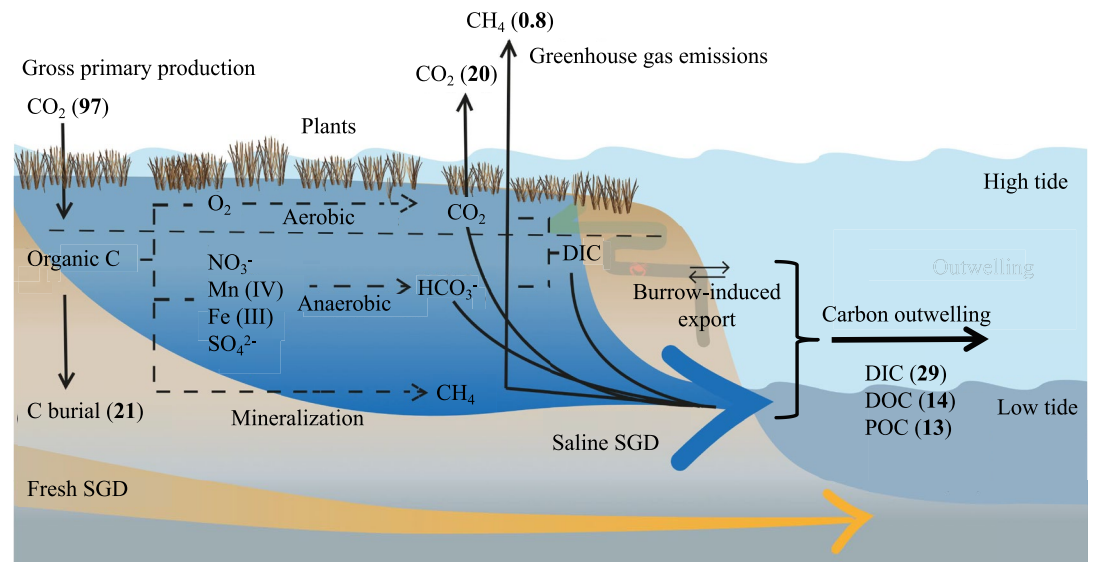
Figure 13. A summary of hydrological and carbon cycle pathways in salt marshes. Submarine groundwater discharge is submarine groundwater discharge. The hydrological background was modified from Santos, Chen, et al. (2021), carbon fluxes upscaled to global salt marshes in units of Tg C y^{-1} are obtained from Alongi (2020), and simplified biogeochemical reactions are obtained from Burdige (2007). Globally, the gross primary production of salt marshes is 97 Tg C y^{-1}. The outwelling fluxes of dissolved inorganic carbon, dissolved organic carbon, and particulate organic carbon are 29, 14, and 13 Tg C y^{-1}, respectively, which are comparable to the global carbon burial (21 Tg C y^{-1}) in salt marshes. The remaining carbon is released into the atmosphere in the form of CO_2 (20 Tg C y^{-1}) and CH_4 (0.8 Tg C y^{-1}). There are unquantified but likely large uncertainties in all of those fluxes.

approaches employed to assess flows on different spatial and temporal scales, so it is unclear when and how the results from different studies can be compared (Tamborski et al., 2021).

The few estimates of carbon outwelling in salt marshes remain spatially and temporally limited, preventing robust attempts to build global-scale budgets (Alongi, 2020). A synthesis of observations in salt marshes on the US East Coast revealed that about 80% of the carbon taken in by salt marsh vegetation is eventually outwelled to nearby estuaries (Najjar et al., 2018). In nearly all investigations, salt marshes seem to release disproportionally high carbon fluxes that often exceed both river fluxes draining much larger catchments and local carbon burial in sediments (Liu et al., 2021; Najjar et al., 2018).

The speciation of carbon determines whether salt marsh carbon will be stored in the ocean or returned to the atmosphere. Salt marshes export detritus or particulate organic carbon (POC), DOC resulting mostly from leaching of vegetation and soil organic matter, and DIC resulting from biogeochemical reactions within sediments. Regional and global scale estimates indicate that salt marsh DIC outwelling rates are about double DOC outwelling, and one order of magnitude greater than POC (Alongi, 2020; Najjar et al., 2018; Figure 14). DIC is made up of CO_{2aq}, HCO_3^- (bicarbonate) and CO_3^{2-} (carbonate). Salt marshes are enriched in all of these DIC species. At high pH approaching seawater values >8, most of the DIC will be present as carbonate alkalinity (HCO_3^- and CO_3^{2-}) that can be stored in the ocean over long time scales. At low pH (<6.5), most of the DIC will be present as CO_2 that returns quickly to the atmosphere. Therefore, the pH of coastal waters receiving salt marsh outwelling will control if DIC is stored in the ocean over long time scales (Sippo et al., 2019).

When oxic seawater reacts with organic-rich salt marsh sediments, the available dissolved oxygen is rapidly consumed. As a result, DIC production in salt marshes is often related primarily to sulfate reduction in organic-rich sediments fed by a nearly constant supply of sulfate from seawater infiltrating sediments (X. Hu & Cai, 2011). The sulfate in seawater (~28 mM) becomes the energy source for microbes consuming organic matter in sediments (Berelson et al., 1996). During sulfate reduction, pyrite (FeS_2) and iron monosulfide (FeS) burial occur (Berner et al., 1970; Reithmaier et al., 2021) along with release of bicarbonate (alkalinity) to groundwater. Understanding the ratios between DIC and alkalinity thus provides insights into whether salt marsh outwelling can
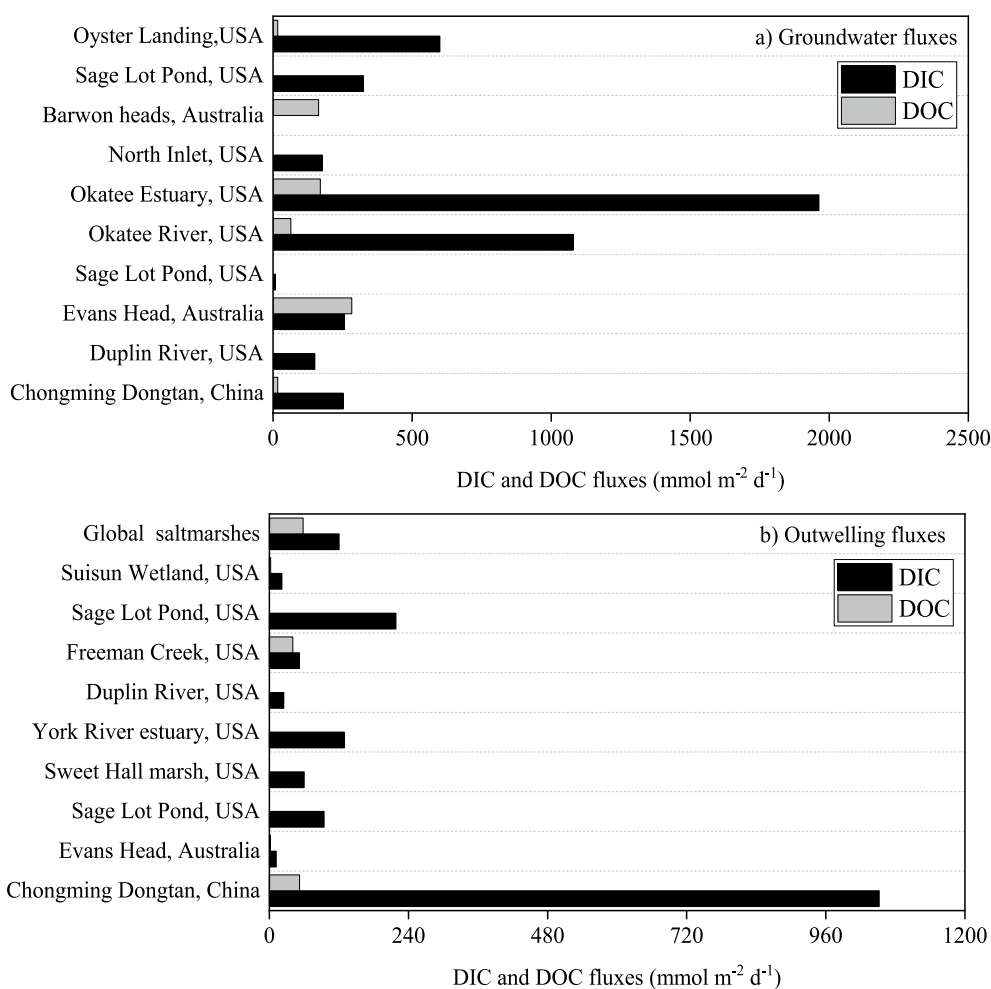
Figure 14. Summary of dissolved inorganic carbon (DIC) and dissolved organic carbon (DOC) fluxes in salt marshes including groundwater to surface water fluxes (a) and outwelling of carbon from marsh to the coastal sea (b). Note that groundwater and outwelling fluxes are for per unit surface water area (i.e., inundated area) and per unit catchment area, respectively. DIC exceeds DOC fluxes in most salt marsh study cases. DIC and DOC speciation are not available in some cases. Original data can be found in Raymond et al. (2000), W. Cai et al. (2003), Neubauer and Anderson (2003), Z. A. Wang and Cai (2004), W. S. Moore et al. (2006), Porubsky et al. (2014), Z. A. Wang et al. (2016), Chu et al. (2018), Peterson et al. (2019), Santos et al. (2019), Webb et al. (2019), Alongi (2020), Bogard et al. (2020), Czapla et al. (2020), Correa et al. (2021), Liu et al. (2021), and Tamborski et al. (2021).

be a net sink for carbon and whether oxic or anoxic biogeochemical processes dominate sediment organic matter consumption (Cyronak et al., 2018).

In salt marshes, sediment-generated DIC and alkalinity are exported to coastal waters by groundwater exchange driven by tides or terrestrial hydraulic gradients (A. M. Wilson & Morris, 2012). Tidally driven groundwater exchange usually flushes the organic rich upper meter of intertidal marsh sediments (Figure 13) and tends to exceed the deeper, fresh groundwater flow as a source of carbon by an order of magnitude (Czapla et al., 2020; Tamborski et al., 2021). Conduits created by roots and burrows in salt marshes accelerate groundwater exchange and biogeochemical flows (T. C. Stieglitz et al., 2013; Xin et al., 2009) by increasing sediment permeability (Guimond, Seyfferth, et al., 2020). As a result, burrows enable the transport of oxygen and sulfate deep into salt marsh sediments, increasing the area where biogeochemical reactions can take place as also observed in mangrove sediments (Agusto et al., 2020; T. Stieglitz et al., 2000; Xiao et al., 2020). When coupled to biogeochemical reactions within organic-rich sediments, groundwater exchange releases byproducts of soil organic matter degradation to the coastal ocean effectively enabling outwelling (Santos, Burdige, et al., 2021; W. Cai, 2011).

Enhanced groundwater flushing in burrows can thus minimize the potential for carbon burial in salt marsh sediments (Thomson et al., 2019; Xiao et al., 2020).

Salt marsh carbon outwelling varies seasonally with the highest rates in the summer when both primary productivity and microbial respiration processes are highest (Z. A. Wang et al., 2016). High carbon outwelling rates are expected during episodic events such as situations of low sea level and high groundwater flow that disproportionally contribute to annual fluxes (Chu et al., 2018). Due to the highly dynamic nature of salt marshes, short term observations relying on discrete sampling over one or a few tidal cycles are unlikely to resolve carbon outwelling (Chu et al., 2018). Nutrient enrichment can enhance marsh productivity to some extent and ultimately carbon outwelling (Czapla et al., 2020). As observed in intertidal mangrove wetlands, tidal ranges, species composition, and the 3D creek-bank structure are expected to drive groundwater exchange and biogeochemical processes within sediments (Chen, Santos, et al., 2021). While no large-scale comparative data sets are available to assess the role of physical and geomorphological drivers in salt marsh carbon cycling, erosive marshes are subject to greater loss of POC and DOC (Ganju et al., 2019).

5.2. Greenhouse Gas Emissions

CH_4, CO_2, and N_2O are often supersaturated in salt marsh surface waters, making salt marshes a source of greenhouse gases to the atmosphere with highly variable emission rates (Chmura et al., 2016; Seyfferth et al., 2020). The high organic matter content, microbial activity, and rapid groundwater exchange create excellent conditions for greenhouse gas production in salt marsh sediments (Trifunovic et al., 2020). Methane and nitrous oxide have higher global warming potential than carbon dioxide and can partially offset the carbon sequestration ability of salt marshes via burial or outwelling (Al-Haj & Fulweiler, 2020; Kroeger et al., 2017; Rosentreter et al., 2021).

Tides are a major driver of greenhouse gas emissions from salt marsh water as revealed by high-resolution time-series (Trifunovic et al., 2020) and geochemical tracer observations (Porubsky et al., 2014; Santos et al., 2019). The periodic inundation and exposure of intertidal marsh sediments create cyclic aerobic conditions that promote greenhouse gas production and emissions to the atmosphere, usually at low tide (Rosentreter et al., 2021). Indeed, tidal cycles seem to be more important than day-night cycles for driving CO_2 dynamics (Tong et al., 2013). Day-night dissolved CO_2 cycles would imply dominant photosynthesis-respiration cycles in the water column, which does not seem the case in salt marshes. Salt marsh groundwater is highly enriched in greenhouse gases relative to surface water. However, only a few studies have measured groundwater-derived greenhouse gas fluxes in salt marshes (Porubsky et al., 2014; Santos et al., 2019; Schutte et al., 2020). Crab burrows play an essential role releasing greenhouse gases from salt marsh sediments by creating channels, expanding the area available for sediment-water exchange, and trapping gases from crab and sediment respiration (Xiao et al., 2020).

The global average water-air CO_2 flux from salt marshes is 6.8 mmol m^{-2} d^{-1} (Alongi, 2020). CO_2 emissions are supported by microbial decomposition of sediment organic carbon that enhance groundwater DIC and CO_2. CO_2 partial pressures in groundwater can exceed 50,000 µatm and drive high groundwater-derived CO_2 fluxes in the range of 58–120 mmol m^{-2} d^{-1} (Santos et al., 2019). In some cases, CO_2 emissions from salt marshes can even exceed regional river DIC fluxes, as observed on the marsh-dominated US East Coast (W. Cai & Wang, 1998) that tend to have the highest CO_2 fluxes in the summer when soil respiration rates are highest (W. Cai et al., 1999; Z. A. Wang & Cai, 2004). CO_2 fluxes from water-atmosphere interface can be twice as high as those from sediment-atmosphere interface, yet only a few studies compared sediment and water emissions (Neubauer & Anderson, 2003; Trifunovic et al., 2020; Z. A. Wang & Cai, 2004). While the water-air CO_2 flux of salt marshes is comparable to mangroves, average global CH_4 emissions from salt marshes are about nine times greater than from mangrove water (Alongi, 2020).

The average global CH_4 water-air flux in salt marshes is 0.24 mmol m^{-2} d^{-1} (Alongi, 2020) and can range from −0.09 to 94.4 mmol m^{-2} d^{-1} when including both water and sediment fluxes (Rosentreter et al., 2021). Methane is generated during microbial decomposition of organic matter usually under anaerobic conditions. CH_4 is then transported from sediments to the water column, and ultimately the atmosphere through diffusion, bubble ebullition and plants. Salinity controls CH_4 production, with higher CH_4 fluxes associated with low salinity (and thus low sulfate) marshes (Bartlett et al., 1987; Poffenbarger et al., 2011). The presence of high concentrations of sulfate in high salinity systems makes sulfate reduction inhibit methanogenesis. However, the effect of salinity on

CH_4 is still debatable because methylotrophic methanogenesis could produce some CH_4 even with the presence of sulfate-reducing bacteria (Seyfferth et al., 2020).

During tidal inundation, periodic anoxic conditions enhance anaerobic respiration and CH_4 production (Seyfferth et al., 2020). CH_4 is thus trapped in sediments and groundwater, escaping to the atmosphere at low tide (Hirota et al., 2007). Some of the methane is laterally exported during the ebb tide and escapes to the atmosphere off the marsh once it reaches the coastal ocean or is oxidized to CO_2 (Trifunovic et al., 2020). Ebullition also contributes to water-air CH_4 emissions but remains poorly understood in salt marshes (Al-Haj & Fulweiler, 2020; W. Yang et al., 2017). Seasonal cycles in groundwater levels can drive CH_4 fluxes in salt marshes (Vázquez-Lule & Vargas, 2021). CH_4 can range from 2 to 1,000 $\mu mol\ l^{-1}$ in groundwaters and exceed surface water by two orders of magnitude (Porubsky et al., 2014; Schutte et al., 2020; W. Yang et al., 2017). CH_4 in surface water was correlated with the groundwater tracer radon (Santos et al., 2019). Since radon is an unambiguous tracer of water in contact with sediments, these correlations indicate that groundwater flow or porewater exchange in salt marshes is an important source of CH_4. Groundwater-derived CH_4 fluxes in salt marshes were quantified only on three occasions and range from 0.2 to 1.2 $mmol\ m^{-2}\ d^{-1}$ (Porubsky et al., 2014; Santos et al., 2019; Schutte et al., 2020). These values are comparable to the global average of 0.24 $mmol\ m^{-2}\ d^{-1}$ emitted from salt marsh surface water (Alongi, 2020).

Salt marshes emit relatively small amounts of N_2O compared to CO_2 and CH_4. Water-air N_2O fluxes are highly variable, ranging from 0.001 to 0.04 $mmol\ m^{-2}\ d^{-1}$ (Ferrón et al., 2007; Sun et al., 2014; Tong et al., 2013). N_2O is produced during microbial respiration of soil organic matter as a byproduct of denitrification and nitrification. Water depth and oxygen availability control N_2O fluxes. While net N_2O emissions occur during anoxic submerged periods, N_2O uptake was observed when sediments were emerged and thus more exposed to oxygen that minimizes denitrification (Tong et al., 2013; W. Yang et al., 2017). The presence of vegetation in the marsh will also modify nitrification and denitrification by transporting oxygen through roots (Hirota et al., 2007). High nitrogen availability in the marsh enhances denitrification and drives high N_2O emissions (Sun et al., 2014). As for CO_2 and CH_4, groundwater and porewater flow releases N_2O from intertidal sediments. Groundwater-derived N_2O fluxes in salt marshes were reported as 8.8 $\mu mol\ N\ m^{-1}\ d^{-1}$ in US (Porubsky et al., 2014) and 170 $\mu mol\ N\ m^{-1}\ d^{-1}$ in Australia (Wong et al., 2013). In the Australian case, groundwater-derived N_2O fluxes exceeded local riverine inputs by an order of magnitude. In both cases, salt marshes were influenced by large anthropogenic nitrogen inputs. There are no estimates of N_2O emissions from pristine salt marshes though they are likely to uptake N_2O as observed in mangroves with no nitrogen inputs (Maher et al., 2016). Additional data are needed to evaluate whether salt marshes act as a source or sink of N_2O and build global estimates.

5.3. Nitrogen and Phosphorus Loads

Nutrients are naturally enriched in salt marsh groundwater relative to surface water due to the mineralization of sediment organic matter and aquifer sources (Guimond & Tamborski, 2021; Howes & Goehringer, 1994; Santos, Chen, et al., 2021). When combined with rapid flow, high nutrient concentrations cause large groundwater-derived fluxes to the coastal water (Charette, 2007; Chen, Du, et al., 2021; Krest et al., 2000; Valiela et al., 1978). Indeed, groundwater-derived nutrient fluxes can rival regional riverine nutrient fluxes as observed in South Carolina salt marshes (Krest et al., 2000). Salt marsh groundwater generally has a much higher N/P ratio than the Redfield ratio of 16. As a result, the large nutrient fluxes with high N/P ratios can drive primary productivity and modify the nutrient structure and composition of coastal water (Santos, Chen, et al., 2021; W. W. Wang et al., 2011).

Nitrogen cycling in salt marshes is variable and complex (Tobias & Neubauer, 2019). Salt marsh vegetation can assimilate nitrogen directly from groundwater or from nitrogen-fixing microorganisms. Sediment organic matter and POC can be degraded to NH_4^+ (i.e., ammonification) in anaerobic sediments (White & Howes, 1994). Denitrification is the primary route for losses of gaseous nitrogen from salt marshes and removes about 30% of the available nitrogen (Velinsky et al., 2017). Salt marsh NO_3^- also can be converted to NH_4^+ by DNRA (dissimilatory nitrate reduction to ammonium) in groundwater, which can occur through a fermentative pathway and/or chemoautotrophy coupling hydrogen sulfide oxidation to nitrate reduction (Brunet & Garcia-Gil, 1996; Giblin et al., 2013; Megonigal et al., 2003). The large amounts of NH_4^+ and dissolved organic nitrogen (DON) leaching from soil organic matter will be exported eventually to the coastal sea through porewater exchange and groundwater discharge (Santos, Chen, et al., 2021).

As a result of reducing conditions often observed in salt marsh groundwater, DON and NH_4^+ concentrations are the dominant species (generally >100 μmol L^{-1}), exceeding the oxic species NO_3^- (<10 μmol L^{-1}) by at least one order of magnitude (Tobias & Neubauer, 2019). For example, the groundwater concentrations of NO_2^-, NO_3^-, and NH_4^+ in a salt marsh of the Yangtze Estuary were 0.3, 1.5, and 280 μmol L^{-1}, respectively, making NH_4^+ the dominant DIN species (Chen, Du, et al., 2021). Groundwater-derived nitrogen (mainly ammonium; 9.72×10^6 mol d^{-1}) in salt marshes from Chongming Dongtan, China, were equivalent to ∼12% of the Yangtze River input (mainly nitrate; Chen, Du, et al., 2021). The high secondary permeability of sediments created by burrows and root structures facilitate water and nutrient exchange between SW-GW (Santos, Chen, et al., 2021).

Part of the phosphorus in salt marshes precipitates as hydrated metal phosphate minerals such as $Fe_3(PO_4)_2 \cdot 8H_2O$ due to the alkaline environment (van de Velde et al., 2021). A smaller fraction of phosphorus is dissolved in groundwater as DIP due to high ionic strength (Paludan & Morris, 1999). The DIP is then released to the coastal water through groundwater exchange (Krest et al., 2000; W. S. Moore et al., 2006). Due to the high reactivity of phosphorus with salt marsh sediments, DIP in seawater can be imported into some salt marshes. For example, DIP fluxes in salt marshes of the St. Lawrence Estuary oscillated between negative (import) and positive (export) between −3.73 to 6.34 mg P m^{-2} d^{-1} (Poulin et al., 2009). Greater tidal amplitudes are expected to drive a release of DIP in some salt marshes or shift the system from export to import depending on local conditions (Childers et al., 2002). These opposite patterns (i.e., DIP import or export) may represent the trade-offs at higher tides between enhanced tidal drainage and an expanded oxidized unsaturated zone favorable to DIP sorption to iron oxides (Tobias & Neubauer, 2019). In addition to the tidal factors, phosphorus uptake by salt marsh vegetation and mineralization may also affect the DIP concentration in groundwater and the net import or export (Lillebø et al., 2007).

6. Knowledge Gaps and Research Needs

Despite considerable research efforts made to advance our understanding of SW-GW interactions in salt marshes and associated plant ecology and coastal biogeochemistry, major knowledge gaps remain to be filled.

One primary gap stems from persistent difficulties quantifying SW-GW exchange at large-scales and over long periods. Currently, flow patterns in 2D creek-perpendicular marsh sections are reasonably well established, but 3D behaviors at regional scales and under coupling of multiple forcing factors are poorly understood. This calls for advanced approaches for monitoring hydrological and chemical transport in natural marshes and high-efficiency models for simulating coupled surface water, groundwater and solute transport. At the same time, tracer-based and biogeochemical studies typically have been limited to relatively short monitoring periods, owing to the intensity of effort needed to collect and analyze samples. Long-term field investigations and numerical simulations will be necessary to characterize salt marsh processes, because soil conditions, SW-GW exchange, biogeochemical processes, and chemical exchange between marsh and coastal sea vary at different time scales. Studies that span annual time scales (e.g., A. M. Wilson, Evans, Moore, Schutte, and Joye (2015)) will be particularly important.

Prior studies have generally failed to quantify the impact of temperature on SW-GW exchange and marsh outwelling. Numerical simulations have assumed isothermal conditions (i.e., no temperature difference), and in field studies it can be very difficult to isolate the effect of temperature from among other parameters that vary daily and seasonally. Significant temperature differences can develop between the marsh surface and subsurface; between rainwater, groundwater and surface water (including creek water, river water, and seawater); and between inundated and exposed portions of the marsh surface. Marsh temperatures also fluctuate over multiple time scales, influenced by storms, tides, daily and seasonal fluctuations and climate change. Variations of these temperatures affect groundwater flow, evapotranspiration, soil conditions, plant productivity, and biogeochemical processes. The full impact of changes in temperature on real marsh systems, as opposed to isolated growth experiments and incubations, requires additional study.

Another major knowledge gap centers around bioturbation from invertebrate and vertebrate populations (e.g., crabs, crayfish, clams, mussels, worms, muskrats, nutria, and feral hogs), which modifies sediment geometry and properties. This issue is also linked to long-standing difficulties in characterizing solute transport through macropores and permeable pathways in porous media (Xiao, Wilson, et al., 2019). Current methods must be expanded to quantify the resulting changes in sediment properties, which affect chemical exchange between salt

marshes, coastal seas, and the atmosphere (Guimond, Seyfferth, et al., 2020; Xiao et al., 2020). We also note that invertebrate bioturbation intensity varies seasonally, with more activity in the summer, and may show clear spatial trends in density and size. Within the context of these trends, such features as crab burrows are randomly distributed, and their morphology is irregular. This is much more complex than the highly idealized crab burrows considered in numerical models thus far until now. Future studies are needed to explore linkages among bioturbation intensity, soil conditions and chemical exchange and to quantify the effects for plant growth/zonation and marsh outwelling.

The impact of human activities on salt marshes is also a fertile area for study. Intensive human activities in coastal zones also affect salt marshes from the terrestrial and marine sides. Land use, river management, agriculture, aquaculture and urbanization affect regional hydrological and morphological processes in intertidal zones. Groundwater flow and transport through salt marshes is also commonly altered by trenching, levees and dikes, and dredging, which affects tidal dynamics. With regard to water quality, sewage discharge and fertilizer use affect the nutrient status in salt marshes and plant growth/zonation and chemical cycles. Preliminary studies suggest that development may affect the composition of groundwater below salt marshes, with important implications for chemical exchange between salt marshes and estuaries, but wide marshes may be able to buffer salt marsh groundwater from the effects of development (A. M. Wilson et al., 2021). Understanding the links between human activities and salt marsh systems will be necessary to support decision-making for protecting and restoring salt marshes.

Additional knowledge gaps stem from a significant geographical bias in salt marsh investigations, with most of this research performed on the US East Coast. New investigations in other marsh-dominated coasts are essential to refine global budgets and resolve the hydrological drivers of salt marsh biogeochemical cycles. Marshes in different regions vary tremendously in tidal regime, stratigraphy, sediment supply, fauna, and plant species. Moreover, while a large portion of salt marshes are found in subpolar zones, field investigations in and numerical simulations on salt marshes in seasonally frozen regions are sparse. This significantly inhibits a global analysis of plant zonation and outwelling in salt marshes. This region is sensitive to global climate change and how these salt marshes respond to a warming climate deserve detailed studies.

We must also work to place SW-GW exchange and biogeochemical processing in the context of marsh systems as a whole. A fuller understanding of salt marsh productivity, zonation, and biogeochemical budgets will depend on our ability to link geomorphological and geochemical features, hydro-sedimentary processes, and intraspecific and interspecific biological mechanisms. For example, plant biomass (both above- and below-ground) and morphology (height, stem diameter, and density) change seasonally. These changes in turn affect surface water flow, sediment trapping, evapotranspiration, soil conditions and biogeochemical cycling, including carbon sequestration, carbon export, and nutrient outwelling. This larger context will be particularly important as we consider the effects of climate change. The current mechanistic understanding of linked ecological, hydrologic, geomorphic, and biogeochemical processes is not comprehensive enough to enable assessments of how sea level rise, climate change, or related processes like widespread mangrove expansion into salt marshes (Kelleway et al., 2016) will modify groundwater flows and eventually biogeochemical budgets. Understanding the biogeochemical processes driving carbon outwelling is a prominent example of this need. Separating processes producing alkalinity from those producing carbon dioxide is required to resolve how outwelling will contribute to salt marsh carbon sequestration, because CO_2 returns to the atmosphere and alkalinity remains in the ocean (Santos, Burdige, et al., 2021). While carbon budget calculations (e.g., Najjar et al., 2018) reveal the general behavior of salt marsh carbon budgets, they generally do not examine alkalinity fluxes as part of the overall analysis, and no comprehensive seasonal assessments of alkalinity outwelling are yet available for salt marshes. Similar gaps exist in key nutrient budgets and in budgets for other constituents, including mercury (Das et al., 2013) and emerging contaminants, are essentially unknown.

Sea level rise, as a component of climate change, will alter the coastal landscape and affect marshes dramatically. Raising sea level by millimeters per year on coastal plains translates to horizontal displacements of meters per year. This will force saltwater further into marshes and salinize some freshwater marshes. Storm surges associated with intense cyclones temporarily raise sea level by meters per hour, affecting wide swaths of coastlines. Saltwater forced ashore during these events can rapidly infiltrate fresher marshes to significant depths through burrows and salt fingering. Once in place the denser, saltier water requires months to years to wash away. The ecological effects of increasing the ionic strength of groundwaters are well known, although predicting the timing

and type of disturbance that triggers migration of the forest-marsh boundary remains difficult. Less recognized is the effect of sulfate ion. Increasing sulfate brings a powerful oxidizing agent into the groundwater, increasing the oxidation capacity by one to two orders of magnitude compared to dissolved oxygen. Oxidation of biogenic carbon by sulfate initiates a series of reactions that release soluble nutrients, sulfide, metals and alkalinity. Soil carbon oxidation reduces the marsh's ability to sequester CO_2 and increases outwelling of nutrients. In addition to the direct effect of storm surges on salinity, their propagation into salt marshes affects sediment transport and distribution and alters groundwater flow. Additional effects of climate change such as warming temperatures and changes in precipitation and wind patterns will affect salt marshes in other, largely unknown, ways. Current models do not adequately address the effects of climate change on salt marshes.

Finally, and perhaps most importantly, hydrological, ecological and geomorphological processes are intertwined and conjointly affect marsh structures and functions (Figure 2b). Unfortunately, prior studies typically focused on individual processes, commonly due to disciplinary boundaries. In natural marshes, multiple and interacting processes lead to dynamic abiotic (e.g., marsh topography, creek configuration/distribution, and heterogeneous sediment distribution) and biotic (e.g., plant biomass/distribution and evapotranspiration) conditions. These real systems are much more complex than simplified marsh geometry and sediment geology considered in most prior modeling studies. Interactions among hydrology, sediment transport, geomorphology, and vegetation in response to different boundary disturbances remain open questions. Coastal zones and salt marshes are now squeezed under unprecedented pressure of global climate change and human activities (Michael et al., 2017). We need to strengthen multidisciplinary collaboration among hydrology, environmental science, ecology, geomorphology, oceanography and climatology to advance our understanding of hydrological, biogeochemical and morphological processes in salt marshes and substantially enhance our ability to protect and restore these wetlands.

7. Conclusions

Based on our review of the current state of knowledge regarding SW-GW interactions and their impact on plant ecology and coastal biogeochemistry in salt marshes, the following main conclusions can be drawn.

SW-GW interactions in salt marshes have been examined extensively by modeling, field measurements and tracer studies in the last 20 years. Modeling studies were mainly based on numerical simulations and assumed simple boundary conditions and sediment geology. Most of them focused on 2D creek-perpendicular marsh sections and demonstrated the importance of near-creek water circulations driven by tidal fluctuations. These circulations lead to marked water and chemical exchange between SW-GW. They likely form low-saturation, well aerated and low-salinity zones, particularly in salt marshes with high soil permeability. The extent of near-creek water circulation depends on soil properties and tidal regimes. For example, soil stratigraphy and the distribution of macropores place important controls on groundwater flow and transport. Spring-neap tides and other multi-scale variations in the tidal signal strongly influence lateral drainage to creeks. Studies of 3D marsh systems identified distinct ecohydrologic zones related to creek networks. In the marsh interior away from creeks, lateral drainage to the creeks is weak. In this way, inundation events (frequency and hydroperiod), rainfall and evapotranspiration jointly affect water and chemical exchange and soil conditions.

A range of field techniques are available for determining SW-GW interactions in salt marshes. Seepage meters and hydraulic-gradient approaches are generally used for small spatial scales, whereas mass balance approach and radionuclide analyses are applied to larger scales. Overall, methods for measuring effluxes (e.g., SGD) are relatively mature. In contrast, determining influxes is still challenging. 2D and 3D numerical models, considering both saturated and unsaturated groundwater flow, have been well adopted and integrated with various boundary conditions (e.g., rainfall-induced infiltration and evapotranspiration) and sediment properties (heterogeneity and macropores). Selection of model dimension depends on the research scale. 3D models are more suitable for investigating the marsh hydrological processes at a system scale, whereas 2D models are sufficient for research focusing on areas not affected by complex creek network, for example, creek-perpendicular marsh sections away from other creeks.

The zonation of salt marsh plants is influenced by both abiotic and biotic factors. Overland tidal flow and subsurface abiotic conditions (salinity, nutrient status, soil type, and oxygen availability) are the major factors determining the zonation of salt marsh plants. These factors are modified by tidal fluctuations (inundation frequency and

hydroperiod), topography and soil property. The biotic factors in terms of species competition and facilitation are also closely related to soil moisture, groundwater salinity and oxygen, and nutrient availability.

Large carbon fluxes out of salt marshes ("outwelling") triggered by SW-GW exchange have now been observed at a few sites. Salt marshes remove carbon from the atmosphere, accumulate it in their sediments and eventually discharge carbon, greenhouse gases and nutrients into the adjacent sea via tidal flushing. Salt marsh outwelling not only supports coastal ocean productivity, but also contributes to marine and atmospheric carbon budgets. The extent of outwelling depends on marsh production, soil properties, perturbation and tidal fluctuations and can be highly variable seasonally and spatially.

Appendix A: Governing Equations of Different Subsurface Flow Models

A1. 0D Models

$$V_B S_S \frac{dh}{dt} = \oint_{V_B} - (q_s + q_v + q_h)\, ds \tag{A1}$$

where h is the hydraulic head (L); t is the time (T); V_B is the volume (L^3); S_s is the specific storage of the sediment (L^{-1}); q_s is the net water flux across the sediment surface (LT^{-1}); q_v is the vertical flux from an underlying aquifer (LT^{-1}); q_h is the horizontal flux within the wetland sediment (LT^{-1}); and ds is the differential surface area (L^2) (Nuttle & Harvey, 1995).

A2. 1D Models

A2.1. Linear Boussinesq Model

Under the Dupuit-Forchheimer assumption of hydrostatic pressure, the linear Boussinesq equation is often used to simulate subsurface flow in unconfined aquifers such as salt marshes (Bear, 1972):

$$n_e \frac{\partial H}{\partial t} = K_S L \frac{\partial^2 H}{\partial x^2} \tag{A2}$$

where H is the watertable height above the impermeable base (L); x is the distance (L); K_S is the saturated hydraulic conductivity (LT^{-1}); L is the marsh sediment thickness (L); and n_e is the effective porosity (–).

A2.2. Nonlinear Boussinesq Model

When tidal amplitude is relatively large in comparison to the aquifer thickness, the nonlinear effects should be considered (Parlange et al., 1984):

$$n_e \frac{\partial H}{\partial t} = K_S \frac{\partial}{\partial x}\left(H \frac{\partial H}{\partial x}\right) \tag{A3}$$

A3. 2D Models

When accounting the change of total stress, the groundwater flow governing equation for 2D models is given as (Reeves et al., 2000):

$$\frac{\partial (\phi S_W)}{\partial t} + S_W S_S \frac{\partial \Phi}{\partial t} = \nabla [K(\psi)\nabla \Phi] + Q_p + S_W (1 + \phi)\alpha \frac{\partial \sigma_T}{\partial t} \tag{A4a}$$

$$\text{with } \Phi = \frac{P}{\rho g} + z \tag{A4b}$$

where $K(\psi)$ is the hydraulic conductivity (LT^{-1}), which equals to the saturated conductivity K_S (LT^{-1}) in the saturated zone and relies on the pressure head ψ (L) in the unsaturated zone; S_W is the soil saturation (-); ϕ is the porosity (–); Φ is the hydraulic head (L); P is the groundwater pressure (ML^{-1}T^{-2}); g is the magnitude of gravitational acceleration (LT^{-2}); ρ is the fluid density (ML^{-3}); S_S is the specific storage coefficient (L^{-1}),

$S_S = \rho g[(1 - \phi)\alpha + \phi\beta]$; and α (M^{-1}LT^2) and β (M^{-1}LT^2) are the compressibility coefficients of soil matrix and water, respectively. $S_W(1 + \phi)\alpha\partial\sigma_T/\partial t$ in Equation A4a accounts for the tidal loading, σ_T is the total stress (ML^{-1}T^{-2}), and $\partial\sigma_T/\partial t = \rho g\partial D/\partial t$, with D being the local surface water depth (L). If D is smaller than zero, the local marsh surface is exposed and the tidal loading term is no longer needed. Q_p is the sink or source term (T^{-1}) which can be used to setup evapotranspiration or infiltration.

Following the suggestion of Xin et al. (2009), when K_S is larger than 1.0×10^{-6} m s^{-1}, an alternative approach for dealing with the total stress variation is to ignore the compressibility term. In this way, Equation A4a can be simplified as:

$$\frac{\partial(\phi S_W)}{\partial t} = \nabla[K(\psi)\nabla\Phi] + Q_p \tag{A5}$$

Glossary

Capillarity: Ability of water to move upwards through soil pores due to surface tension.

Groundwater: Fresh or saline water found in porous media (here, soils and sediments) below the land surface or below the sediment-water interface. Includes unsaturated and saturated systems.

Groundwater flow: Movement of water through porous media (here, soils and sediments) below the land surface or below the sediment-water interface.

Hydraulic conductivity: A measure of the capacity of a porous media to transmit water (or other liquid). It depends on the intrinsic permeability of soils, soil saturation, density and viscosity of the water.

Permeability: Intrinsic property of a porous medium that quantifies its ability to transmit fluids

Porewater: Groundwater in the unsaturated zone

Porewater salinity, groundwater salinity: Amount of dissolved salt in a given mass or volume of water in the subsurface.

Soil saturation: Extent to which soil pores are filled with water.

Submarine groundwater discharge (SGD): Total flux of groundwater from coastal aquifers to the sea.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

Data for the global distribution of salt marshes (Figure 1) can be downloaded from https://bdj.pensoft.net/articles.php?id=11764. The details of field sites used for plant ecology (Figure 12) is given in Supporting Information.

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