Seasonal Water Exchange between Aquifers and the Coastal Ocean

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Groundwater of both terrestrial and marine origin flows into coastal surface waters as submarine groundwater discharge and has been implicated as an important source of nutrients, contaminants, and trace elements to the coastal ocean1-3. However, the physical mechanisms controlling this discharge are poorly understood4. Large saline outflow from the subsurface has been observed by direct measurements1,5-7 and inferred from geochemical tracers8-10, yet sufficient seawater inflow has not been observed to balance this outflow, and understanding of the location and timing of inflow are speculative. Furthermore, discharge rates9,11 are out of phase with the seasonal oscillations of inland recharge and water table elevation that drive groundwater flow toward the coast. Here we explain large-scale saline discharge as the result of seasonal movement of the freshwater-saltwater interface within coastal aquifers that is driven by, but lags, seasonal recharge. Seawater is drawn into aquifers as the interface moves landward, and discharges back into coastal waters as the interface moves seaward. This explanation is supported by numerical modelling of the coastal groundwater system and direct observation of offshore fluxes and hydraulic potential gradients.

Studies that have directly measured submarine groundwater discharge (SGD) with networks of seepage meters show that much of the discharging water has salinity near that
of seawater\textsuperscript{5,7,12}, yet seawater inflow to coastal aquifers has not been observed in sufficient quantity to explain the large saline outflow. Moore\textsuperscript{9} used natural radium as a tracer of groundwater in coastal waters to estimate that SGD was \(\sim 100 \text{ m}^3/\text{d per m length of shoreline along the southeastern US coast, equivalent to 40\% of river discharge. From consideration of the regional freshwater balance, Moore}\textsuperscript{10} concludes that most of this discharge is seawater circulation, however natural tracers in seawater provide little insight into the groundwater dynamics that drive seawater circulation.

Several studies that infer SGD from natural radium measured in coastal waters reveal a seasonal pattern that is out of phase with the recharge cycle. Radium fluxes measured over several years along the South Atlantic Bight indicate that discharge is larger in the summer than the winter and spring\textsuperscript{14,15}, and monthly groundwater fluxes estimated from radium measurements in Rhode Island show a distinct pattern that peaks in the summer\textsuperscript{11}. Recharge is lowest in the summer where these studies were conducted along the eastern US coast because the strong seasonal oscillation of evapotranspiration that peaks in the summer dominates a weaker seasonal signal in precipitation. Along the Ganges Delta, SGD is also out of phase with the recharge cycle, although the seasonal pattern is shifted. There, radium fluxes are largest in the winter\textsuperscript{16}, but recharge is highest in the summer because it is dominated by the monsoon rather than evapotranspiration. The overwhelming majority of direct SGD measurements by seepage meters or hydraulic gradients previous to this study have been conducted in temperate regions during the summer only\textsuperscript{7,12,13}.

Both the missing source of saline groundwater to coastal aquifers and the seasonal pattern in total SGD can be explained by seasonal exchange of saline water between aquifers and coastal waters. In most coastal aquifers, freshwater discharge occurs year-round because the water table remains above sea level (Figure 1). In such aquifers, the
well-known Ghyben-Herzberg approximation\textsuperscript{17} predicts that the depth to the freshwater-saltwater interface below mean sea level is 40 times the water table elevation above sea level, a factor that results from the density difference between saltwater and freshwater. Thus, changes in the water table elevation may be amplified by a factor of 40 in the interface depth, potentially driving large fluxes of saline water between the subsurface and coastal waters. The Ghyben-Herzberg relation is only approximate for dynamic groundwater systems because it assumes hydrostatic conditions and no mixing of salt and freshwater, but it remains true that motion of the water table will result in interface movement. As recharge lifts the water table, the interface moves seaward, freshwater moves down to replace saltwater, more freshwater is drawn from inland, and saltwater discharges to the coastal ocean; the opposite set of fluxes occurs when the water table falls.

Our numerical simulations of density-coupled groundwater flow and salt transport in coastal aquifers, forced by a seasonal cycle of recharge, predict that the freshwater-saltwater interface oscillates seasonally, driving a flux of saltwater in and out of the aquifer that can be of similar magnitude to the freshwater discharge (Figure 2). Furthermore, sensitivity analysis confirms that saline discharge lags inland fresh recharge under a variety of hydrogeologic conditions (Supplementary Figures 4 and 5). Aquifer head has been shown to lag recharge by 0-3 months\textsuperscript{18,19} because recharging water takes time to percolate through the unsaturated zone, and because the water table will continue to rise in response to elevated recharge past the time of peak recharge. At the coast, the system is further buffered by movement of the freshwater-saltwater interface. The change in freshwater hydraulic head is translated to the interface, inducing motion on a time scale that is a complex function of recharge forcing, aquifer parameters, and location within the aquifer. Our numerical simulations of the full dynamic system show that the net effect of these processes may be a lag of 1-4 months from peak recharge to peak saline discharge (Figure
Figure 2 illustrates the modelled fresh and saline discharge and inflow across the sea floor. The peak saline discharge for the parameter combinations in the six models was 19-100% of the peak fresh discharge, while total saline discharge ranged from 13-52% of total fresh discharge. Thus, simulated saline inflow during the winter can explain both a decrease in total wintertime discharge and the observed net saline discharge in the summer. Seasonal hydrologic cycles occur in many regions of the world\textsuperscript{20}, and seasonal seawater exchange is predicted for every set of parameters in our simulations (Supplementary Figure 2), indicating that seasonal saline exchange likely exists in a wide range of coastal systems.

To investigate this hypothesis of seasonal saltwater exchange, vertical hydraulic gradients were measured during the winter (February 2004) in the sediments beneath Waquoit Bay, Massachusetts, when ice cover enabled piezometer installation but prevented the use of seepage meters. Saline discharge (~70% of total discharge in August 2003) has been extensively characterized by direct seepage meter measurement during each summer from 1999-2001\textsuperscript{12} and in August 2002 and 2003 (this study) in the same location. The results (Figure 3) indicate that the direction of saline flow is reversed between the winter and summer. Net downward hydraulic gradients from the bay into the aquifer, over a complete tidal cycle, were measured at five locations corresponding to the band of highest saline discharge observed during the summer.

Both the fresh and saline discharge observed at Waquoit Bay in the summer are likely a result of flow in the unconfined aquifer, and seasonal interface motion explains the saline component of summer discharge as well as winter inflow. An analysis of meteorological data from Long Pond in Falmouth, MA\textsuperscript{21} indicates positive recharge during the study period (January 1999 to February 2004) from October to April, and greater
evapotranspiration than precipitation from late spring through early fall (Supplementary Figure 7). Well level data from varying depths in the upper aquifer within 6.5 km of the study site over the same period reveal a clear yearly cycle with a maximum head in April and a minimum in December. This lag between recharge and hydraulic head maxima is consistent with the numerical results.

Several mechanisms other than seasonal exchange have been hypothesized to drive saltwater circulation. However, these processes (labelled 1, 2 and 3 in Figure 1) appear to be of secondary importance at our field site. First (process 1 in Figure 1), tidal pumping drives water into the aquifer at high tide and out at low tide. Little change in elastic storage of submarine unconsolidated sediments can occur over the timescale of tides, so tidal pumping occurs only near enough to shore such that the water table can respond over a tidal cycle. In Waquoit Bay, tidal pumping occurs in a zone less than 15 m from shore, but net inflow has not been observed, and the magnitude of exchange is much less than the saline discharge farther from shore. Second (process 2 in Figure 1), both wave run-up and tides create inflow of seawater in the intertidal zone that discharges offshore. This circulation was investigated in Waquoit Bay with novel seepage meters that operate in the shallow intertidal zone, and also with a sodium bromide tracer injected near the high tide mark. The measured intertidal inflow was much less than that needed to balance saline outflow, and the tracer test results show that water that enters the sediment at high tide discharges between 2 and 3 m from the position of high tide, much closer to shore than the bulk of saline discharge (Supplementary Figure 6). Furthermore, bromide tracer was never detected at depths greater than 1.2 m, indicating that saline circulation due to tides and waves is confined to the intertidal zone. Third (process 3 in Figure 1), salt dispersion along the freshwater-saltwater interface can drive large-scale saltwater circulation. Kohout estimated this flux to be roughly 10% of the seaward flow of groundwater at a Florida field
site. In Waquoit Bay, however, net inflow seaward of the high discharge zone was observed in only one location during each August experiment, both within measurement error of zero flow, and most saline water discharges far offshore of the freshwater interface. In summary, all three of these potential mechanisms for saline circulation result in zero net saline outflow over a tidal cycle and none can explain the large net saline discharge observed in Waquoit Bay during the summer, leaving seasonally-induced saline cycling as the likely explanation for our observations.

Seasonal inflow and outflow of seawater from coastal aquifers may account for a large component of SGD in areas of the world with significant seasonal cycles in net recharge, either driven by the seasonal cycle of evapotranspiration, as in many temperate areas, or by seasonal cycles in precipitation, as in regions with monsoonal climates. The nutrients and contaminants transported by SGD have been shown to greatly affect coastal ecosystems, and the saline component of discharge can be of equal or greater importance in this contribution than the freshwater discharge. In areas along the eastern US coast, the greatest saltwater discharge occurs in the summer when biological activity is maximum and river flow is minimum, so input of nutrients may be of particular importance. Seasonal exchange subjects coastal sediments to flushing by typically oxic seawater from above during inflow periods and often anoxic water from below during discharge periods. Furthermore, the chemistry of saltwater discharge may vary seasonally, because unlike other circulation processes, the initial saline water to discharge has most recently entered the aquifer, and the last saline water to discharge in the yearly cycle has had the longest subsurface residence time. This work demonstrates the connection between the inland seasonal hydrologic cycle and the saline groundwater systems in coastal aquifers, and suggests the potential for important seasonality in chemical loading to coastal waters.
Methods.

Numerical Simulations. Two-dimensional variable-density simulations were performed using the finite element model FEFLOW\textsuperscript{27}. The simulated domain extended 500 m landward and 200 m seaward from the shoreline for aquifers 20 m and 100 m thick (Supplementary Figure 1), and the number of elements ranged from 93,532 to 597,638. The aquifer was simulated as unconfined and spatially homogeneous, with zero flow and zero mass transport boundary conditions along the base and sides. The recharge boundary condition along the landward model top varied sinusoidally in time, with an average value of 0.002 m/d and amplitude of 0.0025 m/d. The seafloor boundary was a constant head with constant concentration where flow was inward, zero concentration gradient where flow was outward. Six model runs are compared in the supplementary material (Supplementary Table 1, Supplementary Figures 1-5), using a variety of aquifer parameters.

Seepage Meters. During August 2002 and 2003 twenty conventional seepage meters\textsuperscript{28}, enough to overcome local variability\textsuperscript{12}, were installed in Waquoit Bay. The meters were aligned in two transects, 1 m apart, and extended 50 m into Waquoit Bay perpendicular to the shoreline in the same location where 40 meters were used in 1999 and 2000\textsuperscript{12}. Eight novel intertidal seepage meters were placed in the nearshore zone of this transect, the locations varying with the position of the tide. These intertidal meters were not submerged, enabling measurement of groundwater inflow and outflow in very shallow water depths. The change in volume of water in the meters due to tidal fluctuations was recorded, as was the volume in an attached bag that maintained zero hydraulic gradient across the seepage meter wall. Seepage meter bags were collected at least every two hours over one tidal cycle. Discharge salinity was determined by conductivity probe measurements before and
after deployment in the conventional seepage meters, and by refractometer measurements from porewater samples 3 cm below the sediment surface next to each intertidal seepage meter. Groundwater flux and salinity measurements agree very well between adjacent conventional and intertidal seepage meters.

**Piezometers.** In February 2004, 11 piezometers were placed through bay ice along the summer seepage meter transect to depths of 0.6 m to 0.9 m. The water levels in each piezometer and the bay were measured with an electronic water level meter approximately every 1-2 hours during daylight hours over several days, along with the salinity profile measured every few hours with a conductivity probe. The average vertical hydraulic gradient over a tidal cycle from two days of data with similar tidal variation was calculated by correcting for density differences in the bay and piezometer water columns. Slug test data from October 20, 2000 and August 28, 2002 along this transect was converted to estimates of hydraulic conductivity using the Hvorslev method\(^{29}\) (Supplementary Figure 8). The density-corrected hydraulic gradients and interpolated hydraulic conductivities have been converted to flux estimates by Darcy’s Law.

Measured seepage rates, both inflow and outflow, from seepage meters correlate closely to hydraulic gradients measured in adjacent piezometers, so these summer seepage meter results and winter piezometer results can be compared directly.

**Tracer Test.** On August 27, 2001, a sodium bromide solution was injected into the beach at the head of Waquoit Bay near the high tide mark, during high tide. The 0.243 M injection solution was designed to have the same density as seawater, 1.025 Kg/L, in order to track the movement of the infiltrating baywater. Twenty-one piezometers were driven to depths of 0.3 to 1.4 m, nested in groups placed approximately every 0.6 m to a distance 4.3 m bayward from the injection point. Small-volume porewater samples were extracted every 1-
2 hours during daylight for four consecutive days, 32 sample times in all. A conductivity probe and a multi-meter connected to a bromide electrode were used to measure the total conductivity (in mS/cm) and Br⁻ concentration (in mV calibrated to moles/L) of each sample.

1. Simmons, G. M. Importance of submarine groundwater discharge (Sgwd) and seawater cycling to material flux across sediment water interfaces in marine environments. *Marine Ecology-Progress Series* 84, 173-184 (1992).


**Supplementary Information** accompanies the paper on *Nature*’s website (http://www.nature.com).

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Figure 1. Cross-section of coastal groundwater flow (not to scale). In the Waquoit Bay watershed, as in many temperate regions, recharge peaks in winter and early spring and is lowest in summer. The water table cycle is similar but lags the recharge, driving movement of the freshwater-saltwater interface that is potentially amplified by a factor of 40 over the change in hydraulic head, although such amplification decreases as dynamic equilibrium is established. Seasonal interface motion induces inflow and outflow of saline water at the sea floor (4) that is several months out of phase with the recharge cycle. Other mechanisms of saline circulation include tidal pumping (1), intertidal circulation (2), and dispersive entrainment (3).
Figure 2. Simulated total fresh discharge, saline discharge, and saline inflow over the sea floor per meter length of shoreline. Seasons are approximate for a typical yearly recharge cycle within the United States. Results are presented for selected days over one simulated year for a model parameter set with an aquifer thickness of 100 m and (a.) lower hydraulic conductivity ($K=1\times10^{-4}$ m/s) and longitudinal and transverse dispersivity ($D_l=0.1$ m, $D_t=0.005$ m), and (b.) higher parameter values ($K=5\times10^{-4}$ m/s, $D_l=2$ m, $D_t=0.1$ m). Increasing either hydraulic conductivity or aquifer thickness results in an increase in both seasonal outflow and dispersive entrainment but reduces lag, and increasing dispersivity increases dispersive circulation, slightly decreases seasonal outflow, and has no effect on lag (see Supplementary Figure 2).
Figure 3. Submarine groundwater discharge into Waquoit Bay, MA. August seepage meter measurements reveal primarily saline discharge over a tidal cycle in both 2002 and 2003. In 2003, net saline discharge was 7.5 m$^3$/d, and net fresh discharge was 3.5 m$^3$/d per m length of shoreline. The freshwater measurement is greater than a water balance estimate of 1.7 m$^3$/d/m$^{30}$, but consistent because the measurements were taken under a topographic high and at only one time point. February hydraulic gradient measurements converted to flow rates indicate saline inflow at the location where peak outflow was measured during summer. An upward hydraulic gradient and fresh porewater were observed beyond 50 m from shore beneath a low-permeability muck cap and are discussed in the Supplementary Material (Supplementary Figure 9).
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Supplementary Material

This supplement provides further numerical modeling results and additional data and analysis from the field site at Waquoit Bay, Massachusetts. The material is organized as one table and nine figures: six for the numerical modeling (Supplementary Table 1 and Supplementary Figures 1-5), and four for the field site (Supplementary Figures 6-9).

The simulation domain is shown in Supplementary Figure 1 and the six sets of parameter values used for the sensitivity analysis are listed in Supplementary Table 1. Results of the different simulations are summarized in Supplementary Figures 2-5. First, the simulated fresh and saline groundwater fluxes across the sea floor are plotted over one year (Supplementary Figure 2), and the total simulated saline discharge results are plotted as a function of aquifer parameters (Supplementary Figure 3). Second, normalized model input (recharge) and output (aquifer head, interface salinity, and fresh and saline velocity) are plotted over one year (Supplementary Figure 4), and the corresponding effects of parameter variation on the time lag between peak recharge and peak aquifer head and velocity are displayed in Supplementary Figure 5.

Data from Waquoit Bay are shown in Supplementary Figures 6-9 and include results from the sodium bromide tracer test (Supplementary Figure 6), evapotranspiration calculations (Supplementary Figure 7), and field measurements of groundwater seepage, hydraulic gradient, hydraulic conductivity (Supplementary Figure 8), and salinity (Supplementary Figure 9) with distance into Waquoit Bay.
**Supplementary Table 1.** Aquifer model simulation parameters. Two values of aquifer thickness and dispersivity and three values of hydraulic conductivity were used in the simulations to analyze model sensitivity.

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Supplementary Figure 1. Model schematic: flow and transport boundary conditions, initial concentration profile, and dimensions. Seasonal recharge varies sinusoidally every 365 days with a mean value of 0.002 m/d and an amplitude of 0.0025 m/d. The model mesh is triangular, with increasing density near the shoreline where flow is high and spatially variable. The small arrows represent the direction and magnitude of simulated groundwater flow at each node during a simulation. The color represents salt concentration: blue is fresh (C=0 mg/L) and red is saline (C=30,000 mg/L). Freshwater flow converges above the freshwater-saltwater interface to discharge within the first 4 m from the shoreline. Models were run to pseudo-steady-state and then analyzed over one simulation year.
Supplementary Figure 2. Total fresh discharge, saline discharge, and saline inflow over the sea floor per meter length of shoreline on selected days throughout a simulated year for each model run. Models 1 and 5 correspond to (a.) and (b.) of Figure 2. Dispersive circulation occurs throughout the year and can be seen as saline outflow during times of net saline inflow. Saline water flows into the aquifer away from the shore, the magnitude decreasing monotonically with distance, and out of the aquifer on the seaward edge of the freshwater discharge, with inflow and outflow occurring simultaneously but in different places along the boundary.

Seasonal saline circulation is evident as saline inflow during part of the year and saline outflow following the period of high recharge. The seasonal component of saline flow also decreases monotonically with distance, but the direction is either in or out, depending on the time of year rather than the position along the boundary.
Supplementary Figure 3. The effect of model hydraulic conductivity (K), longitudinal dispersivity (D_l), and thickness (b) on saline discharge. Total saline circulation and peak saline discharge as a percentage of peak fresh discharge are plotted against parameter values. Total saline circulation and the proportion of saline discharge increase with both hydraulic conductivity and aquifer thickness. Increasing the dispersivity slightly increases the amount of total saline circulation, but decreases the seasonality of the system: the proportion of peak saline discharge compared to peak fresh discharge.
Supplementary Figure 4. Normalized variation in recharge, aquifer hydraulic head, interface position, and velocity at fresh and saline points of the bay floor over one simulation year for each of the six model runs. Actual values were normalized by their maximum and minimum over the year to reveal the phase of the annual cycles for each variable and the relative time lag. Hydraulic head is reported for a point 50 m landward of the shoreline at sea level. Concentration, or salinity, at a point 20 m landward of the shoreline within the freshwater-saltwater interface indicates interface movement: highest concentration coincides with the extent of landward interface motion, and lowest concentration coincides with the seaward extent. Freshwater velocity at the shoreline and saline velocity on the seafloor 20 m from the coast indicate discharge variation throughout the year. Seasons are
approximate for a typical yearly recharge cycle within the United States. Model characteristics are listed in Supplementary Table 1. The lack of dependence on dispersivity is clear in the nearly identical patterns in models 1 and 4 and models 2 and 3. In the thin and high K models, saline velocity does not track head and fresh velocity exactly; instead it exhibits a slightly lower lag from the maximum recharge, possibly because unlike fresh velocity, it is a result of the interface velocity rather than the aquifer head gradient. The salt concentration indicates that the interface begins to move seaward 0-30 d after the aquifer head begins to rise.

**Supplementary Figure 5.** The effect of model hydraulic conductivity, dispersivity, and thickness on time lag. The number of days between peak recharge and peak aquifer head 50 m landward of the shoreline, freshwater velocity at the shoreline, and saline velocity 20 m offshore are plotted against parameter values. Hydraulic conductivity and thickness generally vary inversely with time lag, while aquifer dispersivity has no effect.
Supplementary Figure 6. Interpretation of NaBr tracer test data. Contours of natural salinity are shown as grayscale, contours of injected bromide are shown as solid lines. Dashed contours are inferred, dashed piezometers indicate screen location and length. Salinity is approximated by electrical conductivity measurements in mS/cm and bromide concentration is in moles/L. The injection point is the approximate extent of water at high tide, and piezometers are placed along a transect perpendicular to the shoreline. (a) Experimental set-up and salinity
profile. (b)-(h) Approximate subsurface bromide contours for selected sample times. The plume appears to travel downward initially and then circulate upward, the center moving roughly 1 m/d. The bayward edge of the plume appears to begin to flow into the bay approximately 40 hours after injection, discharging between 2 and 3 m from the position of high tide and the injection point. Bromide and saline porewater are never detected in piezometers driven to depths greater than 1.2 m, supporting the assertion that saline circulation due to tides and waves in Waquoit Bay is confined to the first few meters into the bay, closer to shore than most of the discharging fresh water.

Supplementary Figure 7. Monthly recharge (precipitation – runoff – evapotranspiration) of water to the subsurface estimated from average monthly rainfall and temperature data\textsuperscript{1} near Waquoit Bay using the Thornthwaite\textsuperscript{2} method. Monthly precipitation varies widely throughout the year, but the seasonal cycle in incoming solar radiation results in a temporal pattern in evapotranspiration that dominates precipitation.
Supplementary Figure 8. Comparison of hydraulic gradient and discharge profiles for summer and winter investigations along a transect perpendicular to the shoreline in Waquoit Bay. (a.) Hydraulic conductivity estimates from slug tests and interpolated values (left axis), and winter hydraulic gradient (right axis). Interpolated values are used to calculate groundwater discharge from gradient measurements using Darcy’s Law. The conductivity estimate 70 m from shore is extrapolated from the measured data. (b.) Summer and winter submarine groundwater discharge. Flow of baywater into the aquifer is observed where maximum offshore outflow was measured during the summer. Saline discharge is minimal in the February experiment.
Supplementary Figure 9. Discharge salinity for summer 2002 and 2003 seepage meter studies and porewater salinity for winter 2004 piezometer study are displayed on the left axis. Horizontal lines represent average baywater salinity. Winter 2003 hydraulic gradient is shown on the right axis. Winter gradient measurements were made beyond 50 meters offshore where summer flux measurements were impossible because of a thick layer of organic muck that prevents seepage meter placement. The large upward gradient corresponding to fresh porewater far from shore is likely a result of upwelling from a confined aquifer. Geological analysis\textsuperscript{3,4} and well logs reveal three geologic layers under Waquoit Bay: an upper coarse to medium sandy aquifer 11 m thick and a lower fine-sand aquifer separated by a confining unit of silt, clay, and very fine sand. The organic layer thickens with distance and likely wedges into the top layer, breaching the confining unit, and leading to the observed upward gradient and fresh porewater from the lower aquifer. We postulate that this gradient drives minimal flow due to the very low permeability of the mucky layer. The downward winter gradient corresponds to saline porewater equal to that of baywater.

