GROUNDWATER DYNAMICS IN AN UNCONFINED COASTAL AQUIFER: GEOPHYSICAL INVESTIGATIONS AND MODELING

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GROUNDWATER DYNAMICS IN AN UNCONFINED COASTAL AQUIFER:
GEOPHYSICAL INVESTIGATIONS AND MODELING

BY

JEEBAN PANTHI

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
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Abstract

Coastal groundwater plays a vital role in the sustainability of coastal ecosystems and the freshwater supply for over one billion people worldwide. Understanding the movement and interaction of saltwater and freshwater in coastal aquifers is crucial for effective water resource management. The interface between these density contrast waters is highly dynamic and sensitive to natural drivers, such as sea level rise, storm surges, and drought, as well as anthropogenic factors, including groundwater over-pumping. In recent years, the drivers of coastal groundwater dynamics have been increasingly well understood and quantified. However, there is a need for data-model fusion, including integrating different forms of data, to address the challenges posed by the impacts of climate change and human activities.

Non-invasive geophysical techniques have emerged as effective tools for monitoring hydrological processes in coastal aquifers. For instance, the time-lapse electrical resistivity survey method, combined with groundwater monitoring, provides insight into saltwater-freshwater interaction in heterogeneous aquifers. I used this method along the southern coast of Rhode Island and developed baseline saltwater intrusion maps and demonstrated its applicability for rapidly estimating fresh groundwater discharge.

Bedrock topography delineation is essential for shallow groundwater mapping and can be achieved using non-invasive geophysical techniques, such as the Horizontal to Vertical Spectral Ratio (HVSR) seismic method. I developed a high-resolution bedrock topography map for the southern coast of Rhode Island using historical borehole lithological data and the HVSR technique. I validated the HVSR based bedrock topography mapping with
Electrical Resistivity Imaging technique. These non-invasive geophysical techniques provide a cost-effective and time-efficient approach to monitoring hydrological processes in coastal aquifers.

While the impacts of sea-level rise, storm surges, and over-pumping have been extensively studied, the impact of droughts on coastal aquifers remains largely under-investigated. I present a new approach for evaluating the fate of a freshwater lens during drought conditions by combining in-situ observations, geophysical measurements, and numerical modeling. Using a density-driven flow model and time-lapse electrical resistivity imaging, I examined the response of a shallow unconfined aquifer on a barrier island during the 2020 Northeastern United States drought. My results indicate an 11% reduction in the freshwater lens due to reduced recharge during the drought, returning to its normal position over the 2021 spring season. The impacts of drought on coastal aquifers are likely to become more significant with the predicted increase in the frequency and severity of droughts due to climate change. Therefore, my findings highlight the vulnerability of shallow unconfined coastal aquifers to droughts, emphasizing the importance of further in-depth studies.

In summary, the understanding of coastal groundwater dynamics is critical for sustainable water resource management in the face of potential climate change impacts. Non-invasive geophysical techniques, such as the time-lapse electrical resistivity survey method and the HVSR seismic method, have been demonstrated as effective tools for monitoring hydrological processes in coastal aquifers. In addition, there is a need for more monitoring networks and multi-disciplinary collaborations to address the challenges posed by the impacts of climate change and human activities on coastal aquifers.
Acknowledgment

“Science, my lad, is made up of mistakes, but they are mistakes which it is useful to make, because they lead little by little to the truth.” - Jules Verne

First and foremost, I offer my sincere gratitude to my major professor, Dr. Thomas B. Boving, for providing me with the opportunity, funding support, and necessary guidance to delve into the often-overlooked field of coastal hydrodynamics. I attribute much of the success of this work to your unwavering guidance during our in-person or online meetings, even while abroad and juggling other commitments, and constructive comments on my writings and thoughts. Your mentorship has imparted invaluable skillsets and insights that have been instrumental in my research journey. I am truly fortunate to have had you as my mentor. I extend my deepest appreciation to my committee member, Dr. Soni M. Pradhanang, for her exceptional support, both financially and otherwise, throughout my research. You taught me the fundamentals of research even before joining my Ph.D. program and have consistently been there for me, even after office hours. Your guidance has extended beyond cutting-edge scientific research and has been invaluable in many aspects of my personal life. Thank you for believing in me, even when I struggled to see the potential in myself that you saw. I am also grateful to my committee member, Dr. Ali S. Akanda, for his constructive feedback on my research. Your positive feedback and thought-provoking questions during my comprehensive examinations have been greatly appreciated. Every time I met you, either formally or casually, I gained a positive energy to work constantly on my research project.
I extend my deepest gratitude to the many individuals and institutions without them this dissertation would not be at this level. I would like to express my sincere appreciation to my fellow graduate students, Mamoon Ismail, Kyle Young, Brendan McCarron, Max Meadow, Ian Tulungen, and undergraduate students, Sophia Motta, Brandan Depanfilo, Janelle Kmetz, Yeriel Cruz, Kylie Plitt, Liz Niedermeyer, for their assistance with field sampling, as well as several other undergraduate students for their help with data transcribing. I will never forget my fellow graduate and undergraduate students who were with me in the field, rain or shine, day or night, and with hungry stomachs in some instances and dealing with lifting heavy weight and playing with dirt. I am also grateful to Professors Brian Savage, Ambarish Karmalkar, and Chris Russoniello for their expertise in writing code for numerical modeling and data visualization. Dr. Seogi Kang at Stanford University provided me with invaluable assistance by writing me a SimPEG python code on resistivity inversion. Professors Art Gold and Mark Stolt from the NRS department provided invaluable field logistics and tools. I would like to acknowledge the support of the RI state offices, including the Department of Environment Management, Water Resource Board, and Department of Health as well as local towns such as Charlestown (Matt Dowling) and South Kingstown, and residents for their generosity in providing access to their properties for field testing. My sincere appreciation goes to the US Geological Survey in Connecticut (Carole Johnson, Eric While, and John Lane) for providing me with training and access to their geophysical instruments. The research project was made possible by the generous funding provided by the Rhode Island Department of Planning, Community Development Block Grant Program (PIs: Pradhanang and Boving), the RI State-mapping project (PIs Savage, Boving, and Pradhanang), and the Geological Society of America through
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Dedication

My heartfelt dedication goes to my mother, Ram Kumari, whose unwavering support and guidance have been instrumental throughout my research journey and instilled in me the values of being a good human being. Additionally, I humbly dedicate this dissertation to my beloved son, Aviskar, whose presence in my life has revealed and enriched parts of my heart that I never knew existed.
Preface

In crafting this dissertation, I have meticulously followed the format outlined by the University of Rhode Island Graduate School. I present this dissertation in manuscript format in accordance with the university guidelines. The dissertation comprises six chapters, each of which serves a distinct purpose in presenting my research findings.

The first chapter is a broad introduction to the topic, which aims to provide a comprehensive overview of the issues and challenges that my research seeks to address. The subsequent four chapters consist of manuscripts prepared according to the specific formatting guidelines of academic journals in the field of environmental science.

Manuscript I, *Saltwater intrusion into coastal aquifers in the contiguous United States — A systematic review of investigation approaches and monitoring networks*, has been published in the journal Science of the Total Environment in 2022. This manuscript represents a systematic review of existing research on saltwater intrusion in coastal aquifers, providing valuable insights into the approaches and monitoring networks employed by agencies in this field.

Manuscript II, *Time-lapse geophysical measurements for monitoring coastal groundwater dynamics in an unconfined aquifer*, is currently under review in the Groundwater Journal. This manuscript details the development and implementation of time-lapse geophysical measurements as a means of monitoring coastal hydrodynamics, offering a novel approach to addressing critical issues in this field.
Manuscript III, *Delineating Bedrock Topography with Geophysical Techniques: An Implication for Groundwater Mapping*, is currently under review in the journal Catena. This manuscript presents an innovative approach to delineating bedrock topography integrating geophysical techniques and explores its implications for groundwater mapping.

Finally, Manuscript IV, *Squeezing of freshwater lenses in barrier island: A combined geophysical and numerical analysis*, is being prepared for submission to the Journal of Hydrology. This manuscript presents a combined geophysical and numerical analysis of the squeezing of freshwater lenses in barrier islands, offering a unique perspective on this pressing issue.

Each of the manuscripts presented in this dissertation includes detailed information on authorship and publication, which is listed at the beginning of each manuscript. It is important to note that all the manuscripts presented here are as submitted to the journals, providing a transparent and accurate representation of my research findings.
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Chapter 1: Introduction

1.1 Motivation

Coastal groundwater is essential for many coastal communities across the globe. It plays a vital role in providing a source of fresh water for drinking and meeting the domestic, agricultural, and industry water needs. Coastal groundwater also supports the local economy for various purposes, including tourism. Furthermore, coastal groundwater is a critical component of coastal ecosystems, providing freshwater for wetlands, estuaries, and other coastal habitats. These habitats support unique plant and animal species and play an essential role in protecting coastal areas from storm surges and erosion. Overall, we must take steps to protect and conserve this valuable resource, which requires a better understanding of coastal groundwater dynamics (Michael et al. 2017).

Coastal groundwater dynamics is a complex process, as it is affected by various factors that act in a range of spatial and temporal scales. One of the critical issues in coastal groundwater dynamics is a saltwater intrusion, which occurs when saltwater from the ocean or estuaries intrudes into freshwater aquifers, contaminating the groundwater and reducing its quality. Saltwater intrusion can be caused by various factors, such as the over-pumping of groundwater (Jasechko et al. 2020), sea level rise (Langevin and Zygnerski 2013), and changes in land use (Barlow and Reichard 2010). The coastal zones are currently experiencing an escalation in water demands, primarily due to the rapid growth of the population in these areas. This trend is expected to persist in the future (Neumann et al. 2015; Nicholls and Cazenave 2010). As a result, a majority of the coastal aquifers are currently under significant stress (Michael et al. 2017; Post and Abarca 2010). Land subsidence due to groundwater mining can also increase the risk of saltwater intrusion (El
Shinawi et al. 2022), as it can cause the water table to drop, making it easier for saltwater to intrude into freshwater aquifers. Coastal groundwater change can also significantly impact coastal ecosystems. It can alter the distribution of freshwater and nutrients in these ecosystems, affecting the growth and survival of plants and animals (Sawyer et al. 2016).

In coastal areas, groundwater can discharge to surface water bodies, such as rivers and coastal ponds. This discharge can significantly impact the water quality and ecology of these surface water bodies (Robinson et al. 2018) by adding and recycling nutrients and contaminants. However, the submarine groundwater discharges to estuaries and oceans are largely overlooked aspect of coastal groundwater dynamic. Thus, coastal aquifers are mixing grounds of contaminants transported from terrestrial environment and saline water from the ocean and estuaries (Michael et al. 2017). Several approaches are in practice to quantify the groundwater discharge into the coastal surface water bodies, such as radiometric isotope geochemistry (Tamborski et al. 2015), point scale field measurements (Michael et al. 2003), and modeling using water balance approach (Sawyer et al. 2016). However, application of geophysics in monitoring the groundwater is just emerging (Dimova et al. 2012) and this approach needs to be tested in detail in the coastal zone.

Despite an increased effort in monitoring coastal water bodies to understand the dynamics of these systems and assessing the potential impacts of human activities and natural events on them, there are some critical gaps to standardize monitoring effort. This issue is clearly identified in Chapter 2. Geophysical methods, such as seismic, electromagnetic induction, and ground penetrating radar, provide information on subsurface geology and hydrology, enabling the assessment of coastal aquifer systems and groundwater flow patterns. As
explicitly mentioned by Werner et al. (2013), there is a general lack of integrating diverse observation data in saltwater intrusion studies. This knowledge gap is addressed in Chapter 2.

Monitoring saltwater intrusion and submarine groundwater discharge processes is challenging. In Chapter 3, I focused on using non-invasive geophysical techniques, including time-lapse electrical resistivity surveys and groundwater monitoring, to demonstrate their applicability in monitoring coastal hydrodynamics at varying time scales and in different hydrogeological settings. In Chapter 4, the focus is on the mapping of shallow groundwater in coastal aquifers where thick surficial deposits cover the bedrock surface. The integration of non-invasive surface-based geophysical techniques is essential, but the presence of highly conductive fluid (saltwater) poses a unique challenge. The chapter proposes a methodological framework for addressing this challenge.

Saltwater intrusion modeling is a challenging task due to constantly shifting boundary, affected by a range of factors such as changes in sea level, tides, groundwater pumping, and rainfall patterns. Additionally, coastal groundwater systems are highly heterogeneous, with varying lithologies, and hydraulic conductivities. The numerical models used to simulate saltwater intrusion are highly complex and require a vast amount of data such as recharge and evapotranspiration rate, hydrodynamic dispersion to accurately represent the system. Particularly, calibrating the model is difficult (Konikow and Reilly 1999) mainly due to lack of high resolution and depth wise data to constrain the model (Sanford and Pope 2010). Despite these challenges, saltwater intrusion modeling remains a critical area of research due to the potential impact of saltwater intrusion on coastal communities, ecosystems, and economies (Michael et al. 2017). As such, continued research in this area
is essential to advance our understanding of the complex dynamics of coastal groundwater systems and to develop effective management strategies to mitigate the risks associated with saltwater intrusion. This topic is the focus of Chapter 5.

The primary aim of this dissertation is to contribute to the scientific community's understanding of the complex dynamics that govern coastal groundwater systems across a range of temporal scales. To achieve this goal, I have utilized a multi-faceted approach that incorporates a diverse range of data collection and analysis techniques. Specifically, I have employed historical lithological records obtained through crowdsourcing initiatives, established several observation wells at varying depths and distances from the ocean and estuaries, conducted three distinct geophysical surveys (including electrical resistivity imaging, ground penetrating radar, and passive seismic imaging), and developed a comprehensive numerical model capable of simulating the complex dynamics of coastal groundwater systems. Throughout the course of this dissertation, I have explored these research objectives in detail across four individual chapters. In doing so, I have generated an extensive baseline dataset related to groundwater levels, salinity, and overall lithological characteristics that has the potential to inform and guide future studies in this field. Moving forward, I will present a comprehensive overview of the conceptual framework that underpins this dissertation, including a detailed description of the specific data collection and analysis techniques employed in each of the individual chapters.
1.2 Dissertation overview

<table>
<thead>
<tr>
<th>CHAPTER 1</th>
<th>Introduction</th>
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<td>What are the current investigation approaches, and data availability for saltwater intrusion studies?</td>
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<th>CHAPTER 2</th>
<th>Systematic review</th>
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<td>What controls the interaction of fresh groundwater and saline ocean water in coastal aquifers?</td>
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<th>CHAPTER 3</th>
<th>CHAPTER 4</th>
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<td>Offshore and terrestrial controls of groundwater dynamics</td>
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<td>How to characterize the controls of freshwater availability in barrier islands?</td>
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<th>CHAPTER 5</th>
<th>CHAPTER 6</th>
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<tr>
<td>Freshwater lens squeezing and recovery in response to drought</td>
<td>Concluding remarks and way forwards</td>
</tr>
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**Figure 1-1: Chapter organization and their respective research questions.**

In this dissertation, I have included four manuscripts (Figure 1-1) which are in different stages of publications. Manuscript 1 (Chapter 2) reviews the progresses made in monitoring and modeling saltwater intrusion in the United States coasts and globally. I evaluated the use of investigation techniques, such as geochemical analysis, geophysical techniques and numerical simulations. Studying peer reviewed journal articles, I identified major drivers of saltwater intrusion and classified them in different coastal states. Also, I identified gaps
in saltwater intrusion monitoring and proposed increased application of geophysics to complement the sparse monitoring data, including artificial intelligence and machine learning tools. In manuscript 2 (Chapter 3), I explored the application of time-lapse electrical resistivity imaging technique in monitoring saltwater intrusion and submarine groundwater discharge in a heterogeneous aquifer in varying time scales. Saltwater intrusion and submarine groundwater discharge are interdependent hydrological processes in a coastal aquifer, therefore monitoring both processes at the same site gives a deeper insight. I highlighted the usefulness and criticisms of geophysical techniques in monitoring coastal groundwater dynamics. Manuscript 3 (Chapter 4) primarily focuses on developing a methodological framework to integrate two geophysical techniques (electrical resistivity imaging and passive seismic) for mapping bedrock topography in coastal area. Moreover, I demonstrated statistical techniques to refine the regression model for predicting the bedrock topography and we validated our approach taking independent data sets. Adding several data points on lithology, I developed an updated bedrock topography map for the southern coast of Rhode Island, a major contribution to the USGS data repository. Manuscript 4 (Chapter 5) investigates the geometry of freshwater lens in barrier islands. Integrating the in-situ observations, geophysical measurements, and numerical modeling, I developed freshwater lens devastation and recovery in different recharge and drought scenarios. I demonstrated a method to parameterize the saltwater intrusion model using geophysical measurements.

References


Chapter 2: Saltwater intrusion into coastal aquifers in the contiguous United States—a systematic review of investigation approaches and monitoring networks (Manuscript 1)

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Abstract

Saltwater intrusion (SWI) into coastal aquifers is a growing problem for the drinking water supply of coastal communities worldwide, including for the sustainability of coastal ecosystems depending on freshwater inflow. The interface between freshwater and seawater in coastal aquifers is highly dynamic and is sensitive to changes in the hydraulic gradient between the sea- and groundwater levels. Sea level rise, storm surges, and drought are natural drivers changing the hydrostatic equilibrium between fresh- and saltwater. Coastal aquifers are further stressed by groundwater over-pumping because of the
increasing needs of coastal populations. A systematic literature review and analysis of the current state of understanding the SWI drivers is presented, focusing on recent (1980 to 2020) investigations in the contiguous United States (CONUS). Results confirm that SWI is an active research area in CONUS. The drivers of SWI are increasingly better understood and quantified; however, the need for increased monitoring is also recognized. Our study shows that the number of monitoring sites have not increased significantly over the review period. Additionally, geophysical, and geochemical investigation techniques and numerical modeling tools are not utilized to their full potential, and data on SWI is not readily available from some sources. We conclude that there is a need for more SWI monitoring networks and closer multi-disciplinary collaboration, particularly between practitioners in the field and emerging modeling technique experts. Though we focus primarily on CONUS, our insights may be of value to the broader SWI research community and coastal water quality managers around the globe.

**Keywords:** Saltwater Intrusion, Monitoring Networks, Geophysical Techniques, Modeling, Review
1. Introduction

Groundwater is a crucial resource for society as well as for sustaining ecosystems. Globally, many cities depend on groundwater to some degree to meet their drinking water needs. That dependency is exceptionally high in coastal areas (Giordano, 2009), where many of the fastest-growing cities and communities are located. For example, along the 5000 km of coastline in the United States (Konikow and Reilly, 1999), coastal aquifers supply potable water to approximately 95 million people - that is 29% of the total U.S. population (U.S. Census, 2017). The coastal population in the U.S. has increased by 15% over the last two decades (U.S. Census, 2017), with continued increases expected in the following few decades (Neumann et al., 2015; Small and Nicholls, 2003). Freshwater aquifers in coastal zones are susceptible to degradation due to their proximity to seawater, the intensive water demands by the ever-increasing coastal population, and the rise in sea level (Williams, 2009). Consequently, most coastal aquifers are under stress (Cartwright et al., 2004; Michael et al., 2017; Post and Abarca, 2010; Werner et al., 2013). Under natural conditions, fresh groundwater flow towards the sea prevents the seawater from moving inland. If the natural balance is disturbed, seawater can migrate into coastal aquifers. This process, referred to as saltwater intrusion (SWI), presents a significant risk to coastal water supplies (Michael et al., 2017) in the U.S. and globally (Han et al., 2014). SWI increases the concentration of mainly sodium and chloride, making the groundwater unsuitable for drinking purposes once the groundwater contains 3% seawater (Jasechko et al., 2020). The U.S. Environmental Protection Agency (EPA) has set chloride (Cl-) as a secondary standard and limits its concentration at 250 mg/L for drinking water (US EPA, 2020). The average chloride concentration in seawater is 19,000 mg/L (Barlow, 2003). Water with
total dissolved solids (TDS) concentration below 1,000 mg/L is *freshwater*, between 1,000 to 10,000 mg/L is *brackish* water, and an excess of 10,000 mg/L is considered *saline* water (NGWA, 2017). In some cases, the freshwater contamination by SWI can be localized with little to no impact on the regional water supply. In other cases, its effects are noticed over large areas, such as in southern Florida (Barlow and Reichard, 2010; Prinos, 2019). Virtually all coastal aquifers worldwide are inherently susceptible to saltwater contamination to some degree (Williams, 2009), and the interaction between saline water and fresh groundwater in the coastal environment controls many biogeochemical processes and affects coastal ecosystems (Sawyer et al., 2016; Tully et al., 2019). For instance, the inflow of saline irrigation water can potentially alter saltmarshes (Li et al., 2018; Neubauer et al., 2013), impact microbial communities in soils and decrease crop yields (Guimond and Michael, 2021), and can change the soil- and groundwater chemistry such as an increase in ionic strength and alkalization (Darwish et al., 2005; Neubauer et al., 2019).

The equilibrium between saltwater and freshwater bodies in coastal aquifers is delicate and controlled by the dynamic interactions between hydraulic heads of the groundwater and sea level. The extent of saltwater intrusion depends on environmental characteristics of the groundwater system, including topography (Chen et al., 2019; Yu et al., 2016) and geology (Chang and Yeh, 2010; Goebel et al., 2017), and on natural and anthropogenically induced processes and events, such as land subsidence due to groundwater extraction (Yu and Michael, 2019), compaction of geologic deposits, altered drainage patterns (Barlow and Reichard, 2010; Rasmussen et al., 2013), drought (Rodrigues et al., 2019), over-pumping
Broadly, there are two types of saltwater intrusion problems caused by sea level rise: (1) head-controlled or topography limited, and (2) flux-controlled or recharge limited (Befus et al., 2020; Michael et al., 2013; Werner and Simmons, 2009). The inland groundwater level is fixed in the head-controlled intrusion scenario, but groundwater discharge to the ocean varies. In contrast, in the flux-controlled intrusion scenario, the rise in groundwater level is commensurate with the sea level rise, assuming an unsaturated zone is adequate to accommodate additional groundwater. The head-controlled system is more vulnerable to saltwater intrusion because the terrestrial groundwater level does not respond to the rising sea level; thus, the hydraulic gradient between fresh and saltwater is imbalanced, leading to an influx of saltwater landward. The lateral saltwater intrusion is almost irreversible on a human time scale. This is because of the generally slow flow of groundwater; therefore, it may take decades to flush out saltwater, even if the freshwater flow increases (Falkenmark et al., 2003; Kumar and Bithin, 2018).

The rise in sea level, together with the increased frequency and intensity of storm events and droughts, are expected to intensify in the future (IPCC, 2021; Knutson et al., 2020), making coastal aquifers more vulnerable to saltwater intrusion (Goebel et al., 2017). For reference, the global sea level is predicted to rise by 0.6 m on average by 2100 (Klassen and Allen, 2017; Sweet et al., 2022). However, the rate by which the sea level rises varies locally and depends on ocean currents, geologic settings (Goddard et al., 2015; Yin et al.,
and anthropogenic forcings, such as land subsidence induced by oil and groundwater abstraction (Nicholls and Cazenave, 2010; Yu and Michael, 2019). This variability can be illustrated by the Northeastern region of the U.S., where the sea level is rising by 2.5 mm per year, compared to the Mid-Atlantic section where rates are higher (3.5 mm per year) and the Southeastern Atlantic where rates are lower (2.3 mm per year) over the last century (Goddard et al., 2015). An important driver for these local differences along the U.S. East coast is the isostatic rebound in formerly glaciated areas. Similar effects are reported in parts of northern Europe (Lambeck et al., 1998). Independent of the cause, the rise in sea level can reduce the hydraulic gradient in coastal aquifers (Michael et al., 2013), creating an enabling environment for saltwater intrusion. Sea level rise in concert with high tides has substantially increased nuisance flooding along sections of the U.S. coast over the last decades (Moftakhari et al., 2015), such as in Miami in Florida (Hauer et al., 2021), coastal zones in Alabama (Murgulet and Tick, 2008) and in New Orleans, Louisiana (Gotham et al., 2018). Though these nuisance floods are largely non-destructive, at least for the time being, they have the potential to contaminate coastal aquifers with seawater. It is projected that the high tide flooding will increase rapidly along the contiguous United States (CONUS) coastlines in the coming decades, with the highest rates expected for the Atlantic coasts (Thompson et al., 2021).

Severe hurricanes and winter storms, called nor’easter in the Northeastern U.S., result in more catastrophic storm surges causing land erosion, coastal flooding, and vertical infiltration of saline water into aquifers. Storm events cause longer-lasting inundation periods of coastal aquifers but also increases rates in precipitation and then freshwater flow.
towards the coast. In addition, the decay rate of hurricanes after landfall in CONUS has been decreasing in recent years in comparison to historical data (Li and Chakraborty, 2020). Saltwater contamination by inundation during storm events is of local to regional effect. The magnitude of the impact depends on the strength and extent of the storm, the morphology of the shoreline, such as topographic depressions (Yu et al., 2016), the topography of the seafloor, and the shape of the coastline (Debernard and Røed, 2008). In contrast, the inundation of coastal areas caused by tsunamis can be more widespread and severe, as exemplified by the devastating tsunami in the Indian Ocean basin in 2004 (Chinnasamy and Sunde, 2016). However, tsunami events are much less frequent than coastal storms.

Over-pumping of freshwater can exacerbate SWI as pumping reduces the natural hydraulic gradient to the sea which can cause the saltwater-freshwater interface to move inland (Klassen and Allen, 2017). Also, over-pumping can lead to upconing that permits saltwater intrusion from the bottom of the well. Over-pumping can lead to localized saltwater intrusion, as observed in the Biscayne karst aquifer Florida (USGS, 2020). This aquifer, which is the primary source of water for Miami and associated southern towns in Florida, is an example of SWI caused mainly by large-scale groundwater withdrawal and historical sea level rise (Barlow and Reichard, 2010; Langevin and Zygnerski, 2013; Tully et al., 2019).

There are many excellent reviews of SWI research studies at a global scale. For instance, Werner (2010) reviewed SWI in Australia, focusing on investigation and management
approaches, including recommendations toward developing modeling and field assessment tools for managing SWI in the future. Werner et al. (2013) focused their review of SWI processes on measurement, prediction, and management methods on the global level. They concluded that SWI is an 'active' research area but highlighted a need for intensive field measurements to validate the growing number of numerical models. Ketabchi et al. (2016) reviewed the relation of sea level rise to SWI and identified that the SWI controlling factors need to be part of an integrated modeling framework. More recently, Hussain et al. (2019) reviewed SWI management options by focusing on hydraulic and physical management strategies. They recommend developing a network of monitoring wells screened at different depths from the surface. Agoubi (2021) reviewed the extent of saltwater intrusion in North Africa and identified the challenges of inadequate data and the need for a robust data inventory.

Reviews that focus on SWI studies in CONUS are limited in number and scope. Barlow (2003), for instance, presented a detailed review of SWI studies but focused on the Atlantic coast only. Barlow and Rechard (2010) summarized the extent of SWI in North America at the continental scale and described investigation approaches and tools. But since then, field methods have improved, and a lot of new data has been generated. Jasechko et al. (2020) highlighted the lack of monitoring data as one of the biggest challenges for SWI studies in CONUS. In summary, many of these prior reviews identified a need for more extensive monitoring and more data to meet the demand of SWI investigation and management approaches.
Building on these previous reviews, we systematically reviewed recent publications on SWI in CONUS and synthesized the current state of knowledge, critically discussed investigation approaches, and examined the drivers of SWI in light of the new data. In addition, we looked at the availability and distribution of SWI monitoring well networks across CONUS. We then conducted a critical analysis of data availability and collection approaches relevant to the study of SWI in CONUS, specifically the efficacy and limitations of geophysical, geochemical, and modeling methods including hybrid and data-driven methods. Though we focus primarily on the contiguous United States, our insights may benefit the broader SWI research community and coastal water quality managers around the globe.

2. Method

![Flow diagram of the systematic literature review on SWI in CONUS.](image)

We conducted a keyword-based literature search using two scientific search engines, namely *Scopus* and *Web of Science* (Figure 2-2). The keywords were identified based on a subset of seed literature and included single terms to capture relevant publications and...
Boolean combinations to narrow the search and screen out unrelated literature. The search was limited to studies in CONUS from 1980 to 2020 and only included articles published in the English language (Table S1, Supplementary Information). Results of the literature search returned more than 200 references; however, these references did not include reports published by the United States Geological Survey (USGS) because these were not covered by the two search engines. Therefore, we manually selected relevant publications from the USGS repository (n=12). After the initial keyword search, the screening of the scientific papers was based on titles, abstracts, and publication dates. Since our focus was SWI case studies, drivers, and investigation approaches, we did not include studies that deal with the SWI impacts on the ecosystem, agriculture, and SWI restoration. SWI case studies and studies focusing on process understanding in CONUS field sites were more heavily weighted than laboratory studies. After this screening step, 98 publications remained (Table S1, Supplementary Information). These were divided into three major categories based on the SWI investigation methods: geochemical (chemical and isotope tracers), geophysical, and numerical modeling studies. Metadata, such as location or hydrogeologic setting, investigation method/model, driver of SWI, were also documented. We complemented current research methods identified for CONUS with global literature on this topic, when needed. Lastly, to report the current status of SWI monitoring networks in CONUS, we implemented two approaches:

1) The Water Quality Portal (WQP) is a joint effort of the USGS and EPA, and we examined WQP data [https://www.waterqualitydata.us/] toward studying SWI in coastal states. To do this, we identified wells based on their distance from the shoreline and frequency of electrical conductivity (EC) or total dissolved solids (TDS) sampling. The
detailed schema of monitoring well selection from the portal is described in Figure S2 (Supplementary Information).

(2) Further, since the USGS is the leading agency in monitoring groundwater in the U.S., we contacted twelve USGS Water Science Centers, including state and local government agencies, following Barlow et al (2002) and as detailed in Table S3 (Supplementary Information), to identify additional monitoring wells that are not part of the WQP.

3. Results

Our results are organized as follows: Section 3.1 focuses on the evolution of SWI research in CONUS, followed by a discussion of select SWI case studies. Section 3.2 summarizes the state of knowledge about investigative approaches to SWI processes based on: numerical modeling, geochemical, and geophysical techniques. Lastly, Section 3.3 present the result of our investigation of SWI monitoring networks in CONUS.

3.1 Locations, Evolution and Focus of SWI Research

The number of scientific papers on SWI in CONUS has been increasing since the 1980s (Figure 2-3a). Most articles were published on numerical modeling (n=48), followed by geochemical studies (n=36). In comparison, few articles focused on geophysical methods (n=14). The number of published papers on geochemical research and numerical modeling doubled after 2000, while the rate of publications in the geophysical category stagnated over the same period. The number of SWI case studies also varied by region. The largest number of case studies are from Florida (n=33) followed by California (n=19), while our literature search and selection procedure yielded no published studies from Maine, New
Hampshire, Rhode Island, and Connecticut along the Atlantic coast, and Washington and Oregon along the Pacific coast. Further, we divided the number of studies published on SWI in a given state by the length of the state’s coastline to normalize the amount of information produced relative to the size of each state and region (Figure 2-3b). According to this metric, studies from California, Alabama, and Florida outnumber other states in CONUS.

The most frequently studied drivers of saltwater intrusion in CONUS are sea level rise (SLR), over-pumping, storm surge, and drought (Figure 2-4). The importance of these drivers varies regionally. Over-pumping was identified as the most important factor in California whereas SLR was of relative importance in Florida. Specific SLR rates and their variation by region are also shown in Figure 2-4. The rate of SLR is highest along the Gulf of Mexico Coast, followed by the Atlantic Coast, while the SLR rate along the Pacific Coast is generally lower.
Figure 2-3: Saltwater intrusion studies in CONUS. (a) Year-wise publication record of research articles in refereed journals classified by geochemical, geophysical, or modeling methods, (b) State (state code in X-axis) and region-wise studies normalized by the length of shoreline.
Figure 2-4: SWI cases in the United States [Data: Jasechko et al. (2020)]. Also shown are sea level rise (SLR) rates to illustrate regionally variable trends [Data: NOAA]. The upper inset graph shows the main SWI drivers investigated in the articles reviewed.

Along the Atlantic coast, SWI research focuses mainly on over-pumping and SLR. For instance, over-pumping on Long Island, New York, and SWI have been observed since the early 20th century (Stumm et al., 2020). On nearby Manhattan Island, New York, chloride concentration rose to 15,000 mg/L in 2004 in a freshwater aquifer affected by over-pumping (Stumm and Como, 2017). Over-pumping was also reported as the primary driver of SWI in Brunswick County in southern Georgia (Krause and Clarke, 2001) and, similarly, on Hilton Head Island in South Carolina. At that location, over-pumping was linked to increased population, tourism, and industrial activities since the 1970s (Payne, 2010). In a
geochemical investigation, saltwater upconing as a result of over-pumping was reported in the Yorktown aquifer on the southern coast of North Carolina, where the salinity in a well that feeds a reverse osmosis plant has increased from 1,000 to 2,500 mg/L since the late 1980s to 2010 (Vinson et al., 2011). In East Shore, Virginia, increasing groundwater salinization, both in shallow and deep aquifers, is associated with over-pumping and aquifer stratigraphy (McFarland and Beach, 2019). That is, a buried paleochannel is presumed to draw in saltwater landward from the Chesapeake Bay and the Atlantic ocean (Nowroozi et al., 1999). Gaswirth et al. (2002) described a similar influence of aquifer stratigraphy in Raritan Bay in New Jersey. Several pumping wells were shut down due to SWI along a paleochannel. An investigation on the Albemarle Peninsula in North Carolina showed that a drought between 2007 and 2009 induced SWI. Here, TDS concentration in groundwater rose to 12 mg/L from less than 7 mg/L in the pre-drought years (Ardón et al., 2013).

Along the Gulf of Mexico coast, storm surges, besides over-pumping, were cited as major drivers of SWI. Referring to storm surge impacts, Van Biersel et al. (2007) reported that the coastal aquifer in Louisiana was devastated after Hurricane Katrina in August and Hurricane Rita in September 2005, respectively. Here, for at least a year after the storms, the key indicators of SWI (specific conductance, chloride concentration) remained elevated in groundwater. Elevated levels of chloride and TDS were also reported from the coastal aquifer in Baldwin County, Alabama. The main drivers were saltwater infiltration by storm surge events, over-pumping, and direct infiltration from the inter-coastal waterways (Murgulet and Tick, 2008). The dissolution of geologic features, such as salt domes, and irrigation return flows, have increased TDS concentrations in groundwater along the Texas
Gulf Coast (Chaudhuri and Ale, 2014). The highest number of SWI research studies was conducted in Florida. Over-pumping and rising sea levels were the most often cited drivers of SWI in this state, such as in Pompano Beach in southern Florida (Langevin and Zygnerski, 2013) and Biscayne aquifer (Guha and Panday, 2012; Langevin and Zygnerski, 2013). Further, in a modeling study in east-central Florida, Xiao et al. (2019) found that storm surges can significantly salinize groundwater.

On the Pacific coast, as reported from Monterey Bay in California (Goebel et al., 2017), locally high rates of groundwater pumping rates are presumed to be the underlying cause for SWI along much of the west coast. In the Monterey Bay region, saltwater was observed in monitoring wells up to 16 km inland (Goebel et al., 2019, 2017). An ancient seawater intrusion was detected in San Diego in California in a study that analyzed the chemical and isotopic tracers (Anders et al., 2014). Similarly, due to over-pumping, extensive groundwater level decline and seawater intrusion are reported in the coastal aquifer in Los Angeles (Nishikawa et al., 2009). Though a set of barrier wells exists there to reduce the intrusion, saltwater is still occurring in some areas such as the Dominguez Gap (Land et al., 2004). We did not find any article or scientific reports from Washington and Oregon in our review.

3.2 Current SWI Investigation Approaches

The articles and reports considered in this study were subdivided into three categories based on the investigation method(s) employed by the authors: (1) Modeling methods: studies based mainly on numerical/computer models and a few data-driven models, (2) Geochemical methods: studies using natural tracers, and (3) Geophysical methods: studies
collecting data via resistivity imaging, electromagnetic (EM) waves, seismic reflection, or airborne sensing of temperature, besides others.

3.2.1 Numerical Modeling

The physical process of SWI can be explained and modeled in two ways: the sharp-interface approach and the variable-density approach (Dokou and Karatzas, 2012). For the sharp interface approach, the relationship between the hydraulic head, salt concentration gradient, and the depth to the interface between the immiscible fluids, saltwater and freshwater, is expressed by a hydrostatic approximation described by Ghyben (1889), Herzberg (1901), and Cooper (1959).

\[ z = \frac{\rho_f}{\rho_s - \rho_f} \Delta h_0 \]  

where \( z \) is the depth of the interface from the sea level, \( \Delta h_0 \) is the hydraulic head between the sea level and groundwater level, \( \rho_f \) and \( \rho_s \) are the densities of freshwater and saltwater, respectively. The parameter \( z \) is typically unknown, and \( h_0 \) is a function of how much groundwater is extracted or recharged and the change in sea level, which is variable in both time and space. Because of its simplicity and low computational cost, the sharp-interface analytical model is widely employed for seawater intrusion investigations. However, this type of analytical model lacks flexibility because of the simplifying assumptions on isotropy, homogeneity, and geometry. Further, these models neglect the mixing of freshwater and saltwater across a transition zone that is a typical characteristic of SWI.
Several studies are based on the fundamental analytical solutions of Equation 1. For example, Henry (1964) developed a semi-analytical solution to define the shape of the fresh/salt water interface using seaward flux of freshwater. Similarly, Strack (1976) used the Ghyben-Herzberg relation and Dupuit hydraulic assumption and considered a sharp interface approximation to derive a two-dimensional model for saltwater intrusion. Later, Mantoglau (2003) and Stratis et al. (2016) developed an analytical model based on the sharp interface approximation and Ghyben-Herzberg relation and used it to optimize freshwater pumping stochastically.

Based mostly on variably density approach, numerical models are applied to simulate the saltwater intrusion and the mixing zone between salt- and freshwater. These models are based on a combination of mathematical equations solved by computers to find an approximate solution to the underlying physical problem. As such, numerical modeling has been recognized as an efficient tool for understanding the magnitude and complexity of SWI, both at spatial and temporal scales. The application of numerical modeling to U.S. coastal aquifer has been increasing between 1980 and 2020 (Figure 2-3a) and a great variety of modeling approaches and codes are in use. The most frequent ones, together with examples of studies in which these models were employed, are listed in Table 2-1. Of all the numerical models, SEAWAT (n=13) is the most widely used code in CONUS, followed by SUTRA (n=5), MODFLOW (n=4), FEMWATER (n=3), FEFLOW (n=1), DSTRAM (n=1), MT3DMS (n=1), MODHMS (n=1), HydroGeoSphere (n=1) and MODFLOW-USG (n=1).
Table 2-1: Saltwater intrusion numerical codes most frequently used in the CONUS.

<table>
<thead>
<tr>
<th>SWI. Model System</th>
<th>Model Features</th>
<th>CONUS Applications</th>
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<tbody>
<tr>
<td>SEAWAT (Langevin et al., 2008)</td>
<td>Finite Difference, 3D</td>
<td>Mulligan et al., 2007; Lin et al., 2009; Sanford and Pope, 2010; Langevin and Zygnerski, 2013; Masterson et al., 2014; Xiao et al., 2016; Abd-Elaty et al., 2016; Xu et al., 2019</td>
</tr>
<tr>
<td>MODFLOW (Bakker et al., 2013)</td>
<td>Finite Difference, 3D</td>
<td>Guvanasen et al., 2000; Uddameri and Kuchanur, 2007; Reichard et al., 2010;</td>
</tr>
<tr>
<td>SUTRA (Voss and Provost, 2010)</td>
<td>Finite Element/Finite Difference, 2D/3D</td>
<td>Bush, 1988; Nishikawa, 1997; Misut and Voss, 2007; Nishikawa et al., 2009</td>
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<tr>
<td>FEFLOW (Trefry and Muffels, 2007)</td>
<td>Finite Element</td>
<td>Loáiciga et al., 2012</td>
</tr>
<tr>
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<td>Type</td>
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<td>HydroGeoSphere (Therrien et al., 2006)</td>
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<td>Yu et al., 2016</td>
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<tr>
<td>TRANSDENS (Hidalgo et al., 2005)</td>
<td>Finite Element, 1D/2D/3D</td>
<td>Abarca et al., 2013</td>
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<td>FEMWATER (Lin et al., 1997)</td>
<td>Finite Element, 3D</td>
<td>Bray et al., 2007; S. and W.-G., 2008; Xiao et al., 2018</td>
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<td>MT3DMS (Bedekar et al., 2016)</td>
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<td>Tsai, 2010</td>
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<td>MODHMS (HydroGeoLogic, 2002)</td>
<td>Finite Difference, 3D</td>
<td>Guha and Panday, 2012</td>
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<td>MODFLOW – USG (Panday et al., 2017)</td>
<td>Finite Difference, 3D</td>
<td>Memari et al., 2020</td>
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</table>

While an in-depth analysis of the advantages and disadvantages of each code is beyond our focus, each model has distinct capabilities. Briefly, the SEAWAT model is a MODFLOW and MT3DMS-based computer program developed by the USGS (Langevin et al., 2008) to simulate variable-density groundwater flow coupled with multi-species solute and heat transport in three dimensions. A MODFLOW package called SWI2 (Bakker et al., 2013)
simulates the 'sharp interface' or 'Dupuit interface' between the saltwater and freshwater (Dausman et al., 2010). However, in an anisotropic aquifer with high hydraulic dispersivity, the SWI2 package may not adequately capture SWI processes (Dausman et al., 2010). More recently, the density coupled version of MODFLOW-USG code (Panday et al., 2017) was used for simulating saltwater intrusion, such as in Memari et al (2020) for laboratory experiment and Kresic and Panday (2021) in a Karst aquifer. One notable benefit of the MODFLOW-USG code is that it can simulate asymmetric problems using coaxial cylindrical model cells. SUTRA is another public-domain code developed by the USGS. It features variable-density groundwater flow in saturated and unsaturated media, but it is formulated in terms of reference pressure, unlike SEAWAT, which is formulated based on reference head. HydroGeoSphere (Therrien et al., 2006) is a commercial model capable of simulating water flow in a fully integrated mode (Brunner and Simmons, 2012), incorporating both variable saturation conditions and density-dependent fluid flow. These capabilities make it ideal for simulating storm surge impacts on groundwater salination (Paldor and Michael, 2021; Yu et al., 2016). COMSOL Multiphysics is a finite element model and solver (COMSOL, 2008; Q. Li et al., 2009) that has been used to simulate the storm surge effect, such as in the Philippines (Cardenas et al., 2015), but not for modeling SWI in CONUS yet. TRANSDENS simulates density dependent groundwater flow in saturated aquifer, as shown by Abarca et al. (2013) in Massachusetts. Another numerical model used in CONUS is FEMWATER, which is a 3D variable density and variably saturated model capable of simulating SWI. In an experiment in Florida, Xiao et al. (2018) reported that FEMWATER simulations corresponded to SEAWAT simulations. In summary, while sophisticated codes exist, they are generally applied to study a particular
driver of SWI. Covering the full complexity of SWI remains a challenge for current modeling methods.

3.2.2 Geochemical Investigations: The chemical composition of seawater is relatively constant worldwide, but groundwater composition in coastal aquifers is variable (Millero et al., 2008). Therefore, geochemical tracers are used to investigate SWI in coastal aquifers. Key tracers are EC, ions such as chloride, sodium, sulfate, calcium, magnesium, silicon, and isotopes of oxygen, hydrogen, boron, carbon, strontium, and helium. EC and TDS are the most widely reported salinity parameters globally (Thorslund and van Vliet, 2020). In CONUS, chloride is monitored for saltwater intrusion investigations in general (Barlow, 2003), except, in some states, where specific conductance is monitored as a surrogate for chloride concentration, such as in South Carolina. In Long Island, New York, electromagnetic (EM) surveys are conducted periodically, and the EM signature is translated to chloride concentration. Drawbacks of using tracers are that the results are point specific, i.e., tracer concentrations can differ significantly between wells as aquifers are heterogeneous. Also, sample collection from multiple observation points is time-consuming and can be cost-prohibitive.

During the SWI process, besides mixing of two immiscible fluids, mineral dissolution-precipitation and ion exchange occur, which require geochemical investigations for detailed analysis (Habtemichael and Fuentes, 2016). The geochemical model PHREEQC (Parkhurst and Appelo, 2013) is a widely-used open-access tool in CONUS for analyzing these processes, as has been done in the Biscayne aquifer in Florida (Habtemichael and Fuentes, 2016) and in semi-arid coastal areas of Texas (Murgulet et al., 2016). Another geochemical model, PHAST (Parkhurst et al., 2010), simulates multicomponent, reactive
solute transport in a three-dimensional saturated flow system and uses input from PHREEQC and HST3D (Kipp, 1997). This model can simulate a wide range of kinetic geochemical reactions. Geochemist Workbench (Bethke et al., 2021) is another modeling platform and used for geochemical investigation of SWI, particularly for ionic balance. An example of its application is from Merseyside, UK (Mohamed and Worden, 2006). Apparently, no published study based on the Geochemist Workbench modeling platform was found in CONUS.

3.2.3 Geophysical Investigations: Geophysical techniques have found increasing application after 1980s largely because these techniques are non-invasive and cost-effective, especially when mapping SWI over a large area. These techniques are convenient for evaluating the spatial and temporal variation of subsurface and groundwater, therefore; hydrogeologists use geophysical techniques in lieu of sufficient observation wells (Stumm et al., 2020). Further, the interface of saltwater and its dynamism over time can be monitored using a single geophysical technique or a combination of several ones. A widely used and particularly well-suited technique for saltwater intrusion mapping is electrical resistivity imaging (ERI) because of the method’s sensitivity to the electrical conductivity of pore fluids which typically results in a strong contrast between fresh- and saltwater (Goebel et al., 2017; Hermans and Paepen, 2020). Therefore, ERI is well suited for characterizing saltwater in coastal aquifers including identifying sub-surface lithology, and groundwater flow paths (Pidlisecky et al., 2016) and conduits. ERI produces a continuous resistivity profile of the subsurface; therefore, it detects the size of conductive media such as saltwater intrusions into groundwater. This technique has been used in coastal areas of CONUS. An example for a large scale ERI study of SWI is from Monterey Bay in San
Francisco, where surface ERI survey was conducted along 40 km of coastline with an investigation depth of 280 m (Goebel et al., 2017). However, there are some limitations to ERI: its resolution decreases exponentially with depth (Loke et al., 2013), making it less reliable in deeper parts of an aquifer, and it measures the bulk resistivity which is different from fluid resistivity. Moreover, due to the contrast in resistivity between geological materials, the bedrock surface in coastal aquifers can mimic a freshwater layer, and a clay layer can mimic a saltwater layer in a resistivity profile. Because of these uncertainties, it is necessary to use the resistivity profiles in conjunction with borehole data for a more reliable interpretation of the subsurface. ERI can be integrated with other seismic techniques, such as HVSR (Nakamura, 1989), to measure the depth of unconsolidated aquifers above the bedrock (Panthi et al., 2022). Alternatively, cross-hole resistivity imaging can be used to overcome the resolution problem associated with surface resistivity imaging. In this technique, electrodes are placed into wells at depths (Palacios et al., 2020) to obtain higher resolution resistivity, such as saltwater mixing zone, where a change in resistivity over depth is expected to occur. Coupled with hydrogeophysical inversion methods using saltwater intrusion numerical models can significantly improve the spatial resolution of geophysical data (Herckenrath et al., 2012). Similarly, combining the resistivity profiles taken onshore and offshore allows for improved imaging of SWI (Hermans and Paepen, 2020). Also, combining terrestrial resistivity data with frequency domain EM can enhance the resolution of three-dimensional images as shown by Paepen et al. (2020) in Belgium and by Goebel et al. (2019) in California. Integrating ERI with the Multichannel Analysis of Surface Waves (MASW) technique, which measures the shear-wave velocity, provides a high resolution three-dimensional resistivity model as tested in
southern Italy (Martorana et al., 2014). Since there is seasonality in SWI (Palacios et al., 2020; Werner et al., 2013), time-series measurement of resistivity, both parallel and perpendicular to the shoreline, is an excellent way of correlating hydrological controls, such as rainfall, groundwater head, and sea level with SWI (de Franco et al., 2009; Palacios et al., 2020).

The Electromagnetic (EM) method is another geophysical technique for mapping SWI. Both time-domain (Nenna et al., 2013; Stumm et al., 2020) and frequency-domain (Paepen et al., 2020) EM methods are suitable in saltwater environments and sometime considered alternatives to monitoring well networks (McFarland and Beach, 2019). Frequency domain EM allows for better resolution in near-surface aquifers than time domain (Jørgensen et al., 2012). A special case of EM is Airborne Electromagnetic (AEM) which has been used in saltwater intrusion investigation in CONUS in locations such as New York (Stumm and Como, 2017), Florida (Langevin et al., 2003), and California (Goebel et al., 2019). AEM surveys can be carried out using helicopters (Werner et al., 2013). For instance, in Monterey Bay in California, in areas where traditional monitoring networks are not available, offshore airborne EM data were combined with onshore resistivity data to extend a resistivity model to a depth up to 200 m below sea level (Goebel et al., 2019).

Geophysical techniques combined with remote sensing platforms are also applicable to submarine groundwater discharge (SGD) investigations. In head-controlled SWI systems, the amount of groundwater discharge to the ocean can be correlated to the migration of the saltwater-freshwater interface (Paepen et al., 2020). Therefore, thermal infrared data acquired through satellite images (Jou-Claus et al., 2020) or with small unmanned aerial
vehicles (Young and Pradhanang, 2021) can be used to identify SGD locations and SWI monitoring on a regional to local scale, respectively.

Some geophysical techniques used outside CONUS are worth mentioning here. For instance, Cone Penetration Testing (CPT), as demonstrated by Pauw et al. (2017), has been used for SWI investigation in the Netherlands. During CPT, a cone with an electrical resistivity sensor penetrates the ground from the surface to measure the resistivity, a proxy of salinity. An advantage of CPT is its high vertical resolution (~2cm). Another potential geophysical technique to map saltwater intrusion is the sub-bottom profiler (SBP). In the SBP shallow seismic technique, sound energy is transmitted in the form of short pulses. The difference between transmitted and reflected sound energy is a function of the subsurface and can be analyzed to create subsurface imagery. Coupling the SBP technique with EM has been implemented to investigate the surface water and groundwater interaction in a Karst system in Mexico (Bücker et al., 2021).

In general, geophysical methods in SWI studies require specialized and often expensive equipment, extensive training, and also ground observations to interpret the results (de Franco et al., 2009; Demirci et al., 2020). Moreover, the saltwater distribution estimated by the geophysical techniques is approximate because a contrast in geology in an aquifer, such as presence of clay layers, affects the resistivity or electromagnetic signals. Such signals can be misinterpreted and therefore, geophysical data sets require lithostratigraphic data from nearby boreholes for correct interpretation.
3.3 SWI Monitoring Well Networks

Periodic monitoring of salinity in groundwater via observation wells is the traditional and most accurate measurement method for tracking the saltwater – freshwater interface. The number of monitoring wells in each area and the length of time these wells have been monitored determines the degree to which SWI processes can be resolved in space and time.

Figure 2-5: Observation wells with regular geochemical tracers (EC or TDS) measurements for SWI normalized by the length of coastline. Figure S2 (Supplementary Information) provides details about the well dataset selection methodology. Wells were grouped by distance from the shoreline.
Across CONUS, there are very few well networks listed in WQP that are designated to monitor the specific conductance or TDS of coastal groundwater (Figure 2-5). How far inland SWI can be encountered is highly variable regionally and depends on many factors, such as topography and groundwater management. In this study, example distances of 2, 5, and 20 km from the shoreline were used to search for monitoring wells in WQP. Altogether, there are 109, 172 and 308 wells within 2 km, 5 km and 20 km, respectively, in CONUS that are sampled at least once in twelve months between 2000 to 2020. Florida has the most wells within the 5 km and 20 km envelope, but Washington state has the highest number of wells within 2 km from shoreline (Figure 2-5). Connecticut stands out as a state with densest monitoring network. No SWI monitoring wells for either electrical conductivity or TDS or chloride concentrations were located in Maine, New Hampshire, Massachusetts, and Oregon within 20 km of the shoreline. Notably, other than along Florida’s west coast, only six wells in Louisiana within 20 km from the shoreline were identified along the Gulf of Mexico coast. Hence, the WQP dataset has extensive gaps along the Gulf of Mexico coast, followed by the Northeastern region and parts of the Pacific Coast. Also, no Chloride monitoring wells in WQP passed our selection criteria in CONUS but is plausible that such data exists at agencies that do not report to the WQP.

SWI monitoring networks not part of WQP and updated after Barlow et al. (2003) are presented in Table S3 (Supplementary Information). In the United States, the USGS is the leading groundwater quality monitoring agency, and some states’ local government agencies, such as towns, coordinate with the USGS in operating these networks. Networks operated by local entities that do not report to the USGS also exist, but their data are
typically not readily assessable. Irrespectively, the publicly available USGS time-series data sets are critical resources for SWI model development (Jasechko et al., 2020; Michael et al., 2017). The monitoring frequency varies by location from sub-daily to every five years only. Florida has the densest monitoring network and highest frequency monitoring schedule in CONUS. Prinos (2019) developed a map showing how the interface between salt and freshwater is migrating inland overtime in Florida based on data from this network. On the pacific coast in Island County, Washington State, the chloride concentration in public wells is monitored periodically and the sampling frequency is classified by risk (Island County, 2021). The wells with a high risk of SWI are sampled twice per year and the wells with low risk are not monitored until the next risk assessment.

Besides government led SWI data and field observation networks, a few networks are maintained by research groups. For instance, salinity data was previously collected for the United States by Thorslund and van Vliet (2020) but the database’s usefulness was limited by data discontinuity and the lack of meta data information, such as sampling depth. Depth-specific salt concentration provides important information (Stoeckl and Houben, 2012) that can generally be gained when salt concentration (or more likely EC) inside a well is monitored with a multi-level sensors or similar instruments. An example where depth-specific measurements are employed is a monitoring well network is from Monterey Bay in California (Pidlisecky et al., 2016), where sentinel wells are used as early warning system of SWI. Further, we found that in many cases, SWI monitoring networks were established for specific projects and abandoned once the project ended, such as in Ventura.
County in California (Izbicki et al., 1995) and Keyport at Kitsap county in Washington (Opatz and Dinicola, 2018).

4. Discussion

4.1 Enhancement in SWI Monitoring Network

Monitoring SWI is easy to ignore and hard to initiate and to continue because groundwater is a hidden resource below the ground surface, and SWI progresses slowly in general. The USGS SWI monitoring network (Table S3, Supplementary Information) did not grow significantly since the early 2000s (Barlow, 2003). USGS has developed a portal to report the water quality data that is a part of WQP. But the network is relatively small and does not include data from other networks that are not obligated to report to the USGS. Hence, as can be deduced from Table S3 (Supplementary Information), the usefulness of current databases for SWI studies in CONUS is fairly limited. The importance of current SWI networks as a research and management tool could be enhanced by amending them with SWI data currently kept by local and state agencies. Following the recommendations by Jasechko et al. (2020), Barlow et al. (2010), and Werner et al. (2013), SWI monitoring network in CONUS could be improved by developing a standard monitoring protocol for measuring salinity parameter/s, sampling frequency, sampling depth within a well, and most importantly, continuity of monitoring.

4.2 Geophysical Data Integration and Standardization

The limited availability of directly monitored geochemical data promoted the adaptation of geophysical methods in SWI studies. However, geophysical methods largely remain under-utilized in CONUS. As outlined, the most frequently used geophysical techniques
for mapping saltwater intrusion are ERI and the time-domain and frequency-domain Electromagnetic (EM) methods. Since geophysical techniques have been employed in the study of SWI processes, a large amount of data was collected by different government agencies, academic and private institutions. However, as with the geochemical water data, the information is widely scattered and not readily available. It will be worthy to compile the geophysical data in a well-organized and easily accessible central database. However, such an undertaking would require well-defined protocols regarding data collection methods, data processing procedures, like noise filtering and inversion, and interpretation standards, as well as a long-term commitment to maintaining such a geophysical database.

4.3 Coupling Processes in Numerical Modeling

Model projections of future SWI issues are crucial for informed decision making about how to manage and possibly adopt water supply systems, develop plans for coastal ecosystem restoration, and to safeguard coastal infrastructures. To that regard, numerical models are quintessential tools for understanding SWI processes at various temporal and spatial scales. Yet, the scientific literature describes only a few multidisciplinary efforts in CONUS toward integrating geophysical and other monitoring data into modeling frameworks (Werner et al., 2013). For instance, Beaujean et al. (2014) and Herckenrath et al. (2012) utilized ERI and EM data, respectively, to estimate hydrogeological parameters needed for setting up and calibrating SWI numerical models. Similarly, although groundwater and surface water bodies, such as rivers and lakes, are often linked and may co-respond to the SWI process, only a few SWI modeling studies in (n=15) included in our systematic review accounted for both these compartments whereas the majority of modeling studies only considered one or the other. Also, most models described SWI as a
horizontal propagation of the saltwater-freshwater interface, with very few addressing its vertical propagation as resulting from saltwater inundation and infiltration after a storm surge. The importance of modeling both horizontal and lateral SWI during storm events was highlighted by Paldor and Michael (2021). Finally, Werner et al. (2013) points out that for informed decision-making it is necessary to account for aquifer heterogeneity in SWI models. This likely requires a closer collaboration of numerical modelers with sedimentologist, geologist, and geophysicist and even bringing on board well drillers and other non-academic sources with insight into the local aquifer stratigraphy.

4.4 Application of Big Data and Machine Learning

The strong interdisciplinary and integrated nature of SWI research can identify potentially complex interactions between natural and anthropogenic processes. However, when study regions are large, processed-based models are computationally expensive and time-consuming to set up. In this regard, data-driven modeling approaches, such as artificial neural networks (ANN), offer advantages particularly when adequate data, such as 3D hydrogeological properties, are not available for setting up a process-based model. Initiatives are emerging in applying data-driven approaches to SWI research. For example, ANN was used for predicting the extent of SWI and optimizing pumping regiments in coastal aquifers in Greek Island (Kourakos and Mantoglou, 2009). Other ANN-based approaches for modeling SWI management optimization were reviewed by Singh (2014). Lal and Datta (2018) used machine learning for simulating the salinity in monitoring wells in India under variable pumping scenarios. They concluded that a surrogate modeling framework could be applied for water resource management as a "computationally efficient substitute for the complex numerical model". In another example, Ranjbar et al. (2020)
combined ANN with SEAWAT to predict SWI under variable recharge and pumping scenarios, and tested the model in Iran. It was concluded that a conjunctive model is more efficient than the SEAWAT model alone. Similarly, Moazamnia et al. (2020) found that the performance of a SWI vulnerability model improved significantly applying a fuzzy membership function. Overall, machine learning-assisted modeling of SWI has progressed significantly in recent years and is a major research area. We note, however; that purely data-driven models do not necessarily require a priori process understanding, but their success depends even more on data availability and data quality than numerical models (Shen, 2018). Therefore these methods are also data-intensive (Peters-Lidard et al., 2017). For example, Sahour et al. (2020) mapped the spatial distribution of groundwater salinity from established relationships between groundwater electrical conductivity and controlling factors by statistical and machine learning methods. To develop and verify these models, it required accurate measurements of electrical conductivity from 160 piezometric wells with monthly sampling frequency over a several years as well as data on controlling factors such as groundwater head, evapotranspiration, and distance to the sea.

5. Conclusion and Broader Significance
Saltwater intrusion investigations can protect the drinking water supply of coastal population and aid in sustaining critical coastal ecosystems, agriculture, aquaculture, and infrastructure. This study presents a critical review of a selection of SWI case studies, existing investigation approaches, and data collection techniques currently employed in CONUS. Most studies concentrated on populous states, namely Florida and California. Few studies covered the Northeastern CONUS, even though the rate of sea level rise is higher in this region. In some areas, particularly along the Gulf of Mexico, except Florida,
the apparent lack of studies and data about the existing SWI monitoring network might be related to some local and state agencies not reporting SWI studies to central databases, such as the WQP. Also, research sites operated by academic or nongovernmental institutions have data that may not be publicly available in all cases.

In consequence, information about SWI monitoring networks and related data are comparably sparse in CONUS. What data are available is often difficult to correlate across sites because data collection methods, sampling frequencies and depth together with other parameters are not standardized. Specifically, it would be beneficial if salt concentration (or its proxies) in an observation well is monitored at multiple levels rather than at a single depth, as is typically the case when a single electric logger is used for monitoring. An enhanced vertical resolution of the salt concentration profile in a given well would take relatively little monitoring effort but would provide better-informed insights into the underlying SWI processes in coastal areas. Another insight gained from this investigation is that the sampling frequency in existing SWI monitoring networks is low in many areas. More frequent scientific monitoring, together with a denser monitoring well network, can serve as an early-warning system and support the decision process where to implement SWI mitigation measures. Such measures might involve building reverse osmosis treatment systems or artificial recharge infrastructures, like recharge wells or channels, or subsurface physical barriers that may serve as a hydraulic barrier to prevent SWI. Also, most research in CONUS focused on the consequences of over-pumping and sea level rise on SWI, while few considered the effects of storm surge and drought. Understanding the magnitude and impact of these episodic hydrologic events and the time it takes for the coastal aquifer to return to baseline salinity levels is of utmost importance to the coastal
population. This is not just because of the importance of coastal aquifers for drinking water production but also because the frequency, if not the intensity, of storm events and droughts, is predicted to increase in the future.

Since 2000, the number of publications covering numerical modeling and geochemical approaches to SWI investigations in CONUS increased, while published studies using geophysical investigation approaches stagnated. With regard to numerical modeling, significant developments have taken place over the review period (1980 to 2020); however, many of the SWI modeling studies still fail to couple surface hydrological processes with groundwater hydraulics. The most significant impact of numerical models in SWI studies is that they can optimize pumping schedules and identify optimal installation locations for future groundwater abstraction wells or aid in planning artificial and coastal resiliency infrastructure. However, the capabilities of current models remain limited as models are typically focused on resolving one or the other SWI process quite well but are not capable of capturing the totality of underlying processes that lead to SWI. Data-driven models, such as ANN and machine learning, are more recently adopted by researchers and promise to overcome the limitations of numerical models. These new tools do not necessarily require a priori process understanding and can be applied to predict the SWI extent and optimize groundwater withdrawal. However, they operate with Big Data elements only.

Since current networks are sparse and geochemical SWI data is limited, geophysical techniques could potentially meet the data needs of data-driven models. Time series geophysical techniques can, for example, capture the temporal change in hydrological
controls of the SWI, such as drought, variable recharge, or sea-level rise, on larger spatial scales compared to single monitoring wells. But it remains unclear why geophysical methods are not more frequently employed since immense progress has been made in bringing down the cost and availability of the required equipment together with easing the way how the data is processed and interpreted today. Even though geophysical techniques have been used to study SWI in the past, the full potential of these techniques appears not been fully harnessed yet. Also, this review illustrates the importance of aquifer geology, as is exemplified by the complexity of SWI research in the coastal Karst aquifers in Florida. Especially in complex geologic settings, the combination of geophysical and geochemical monitoring and modeling of interconnected surface and subsurface processes utilizing Big Data and machine learning together with process-based modeling carry significant promise towards encountering the negative impacts of SWI on coastal aquifers in CONUS. Overall, we conclude that current SWI monitoring network’s capabilities must be enhanced both in time and space. Data collection methods would benefit from standardization and employing geochemical, geophysical, and data processing methods currently not widely used in CONUS. The inflow of new data would dramatically improve the spatial and temporal resolution of SWI processes and thereby increase the resiliency of coastal areas in response to pressures stemming from sea level rise, a changing climate, and increasing population growth.

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Supplementary Information I: List of articles reviewed and their focuses

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<td>System understanding</td>
<td>Storm surge</td>
</tr>
<tr>
<td>A Density-Dependent Flow And Transport Analysis Of The Effects Of Groundwater Development In A Freshwater Lens Of Limited Areal Extent: The Geneva Area (Florida, USA) Case Study</td>
<td>Panday et al. (1993)</td>
<td>Florida</td>
<td>NM</td>
<td>Process understanding</td>
<td>NA</td>
</tr>
<tr>
<td>An Analysis Of Long-Term Salinity Patterns In The Louisiana USA Coastal Zone</td>
<td>Fuller et al. (1990)</td>
<td>Louisiana</td>
<td>GC</td>
<td>SWI trend</td>
<td>NA</td>
</tr>
<tr>
<td>Time Domain Electromagnetic Soundings For Mapping Sea-Water Intrusion In Monterey County, California</td>
<td>Mills et al. (1988)</td>
<td>California</td>
<td>GP</td>
<td>Tool application</td>
<td>Over-pumping</td>
</tr>
<tr>
<td>Origins Of Seawater Intrusion In A Coastal Aquifer - A Case Study Of The Pajaro Valley, California</td>
<td>Bond and Bredehoeft (1987)</td>
<td>California</td>
<td>NM</td>
<td>Optimal management</td>
<td>NA</td>
</tr>
<tr>
<td>Potential For Saltwater Intrusion Into The Upper Floridan Aquifer, Hernando County, Florida</td>
<td>Ryder and Mahon (1989)</td>
<td>Florida</td>
<td>GC</td>
<td>Process understanding</td>
<td>NA</td>
</tr>
<tr>
<td>Saltwater Study Of The Miami River And Its Tributaries, Dade County, Florida</td>
<td>Vernon (1966)</td>
<td>Florida</td>
<td>GC</td>
<td>Process understanding</td>
<td>Canal construction</td>
</tr>
</tbody>
</table>
Supplementary Information II: Schematic diagram of the monitoring well selection process

<table>
<thead>
<tr>
<th>Step 1: Pre-Selection from WQP *</th>
<th>Step 2: CONUS SWI wells</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coordinates (valid latitude, longitude, and coordinate reference datum)</td>
<td>Selection of coastal (&lt;= 2, 5 &amp; 20 km to shoreline) wells within CONUS only</td>
</tr>
<tr>
<td>Dates available</td>
<td>Drop duplicates</td>
</tr>
<tr>
<td>Original variables:</td>
<td>Combine wells with TDS &amp; Conductivity samples</td>
</tr>
<tr>
<td>- Total (Dissolved) Solids</td>
<td>- Keep samples from year 2000 onwards</td>
</tr>
<tr>
<td>- Specific Conductance/Conductivity</td>
<td>- Keep wells with samples available at least once every 12 months for at least 5 years</td>
</tr>
<tr>
<td>Activity media and sample fraction indicating groundwater samples</td>
<td></td>
</tr>
<tr>
<td>- Units in/convertible to mg/L (TDS) and μS/cm (Conductivity)</td>
<td></td>
</tr>
</tbody>
</table>

Supplementary Information III: Saltwater intrusion monitoring wells network in CONUS

Method of data collection: We contacted twelve USGS water Science Centers in coastal states, including state geological office, selected local agencies, such as county offices, town offices, and state water boards asking location and number of wells, species being monitored and frequency of monitoring for saltwater intrusion. All the centers responded to our questions and provided the information directly by sending us the list of wells and their details or by pointing us their scientific investigation reports. We compiled the information and listed below.

Table S3: SWI monitoring wells in CONUS (Updated after Barlow et al. 2003)
<table>
<thead>
<tr>
<th>Location</th>
<th>County/Municipality</th>
<th>Sample Size</th>
<th>Measured Constituent</th>
<th>Data Source/Agency</th>
<th>Frequency</th>
</tr>
</thead>
<tbody>
<tr>
<td>NY</td>
<td>Long Island</td>
<td>9</td>
<td>Chloride converted from Borehole EM.</td>
<td>USGS and NY Department of Environmental Conservation</td>
<td>Periodic</td>
</tr>
<tr>
<td></td>
<td>Manhattan Island</td>
<td>14</td>
<td>Chloride converted from Borehole EM.</td>
<td>USGS</td>
<td>Periodic</td>
</tr>
<tr>
<td>FL</td>
<td>Miami, Fort Myers</td>
<td>208</td>
<td>Chloride</td>
<td>Real-time to periodic</td>
<td></td>
</tr>
<tr>
<td>NJ</td>
<td>Monmouth County</td>
<td>4</td>
<td>Specific conductance</td>
<td>USGS</td>
<td>Every 15 minutes (sensor)</td>
</tr>
<tr>
<td></td>
<td>Southern coast</td>
<td>68</td>
<td>Chloride</td>
<td>USGS</td>
<td>Every three years</td>
</tr>
<tr>
<td>DE</td>
<td>New Castle County</td>
<td>12</td>
<td>Specific Conductance</td>
<td>Delaware Department of Natural Resources and Environmental Control</td>
<td>Twice a year</td>
</tr>
<tr>
<td></td>
<td>Sussex County</td>
<td>32</td>
<td>Specific Conductance</td>
<td>Delaware Department of Natural Resources and Environmental Control and Delaware Geological Survey</td>
<td>Once to twice per year</td>
</tr>
<tr>
<td>LA</td>
<td>Southern coast</td>
<td>150</td>
<td>Chloride</td>
<td>USGS</td>
<td>Annual</td>
</tr>
<tr>
<td>RI</td>
<td>Charlestown</td>
<td>9</td>
<td>Specific Conductance</td>
<td>University of Rhode Island Department of Geosciences</td>
<td>Every 6 minutes (sensor)</td>
</tr>
<tr>
<td>CA</td>
<td>Los Angeles Basin</td>
<td>78</td>
<td>Chloride</td>
<td>USGS California Water Science Center</td>
<td>Quarterly to yearly</td>
</tr>
<tr>
<td>Location</td>
<td>Wells</td>
<td>Parameter</td>
<td>Agency/Source</td>
<td>Frequency</td>
<td>Notes</td>
</tr>
<tr>
<td>---------------------</td>
<td>-------</td>
<td>-----------</td>
<td>-------------------------------------------------------------------------------</td>
<td>-------------------------</td>
<td>----------------------------------------------------------------------</td>
</tr>
<tr>
<td>Ventura County</td>
<td>232</td>
<td>Specific Conductance</td>
<td>USGS</td>
<td>No updates</td>
<td>USGS</td>
</tr>
<tr>
<td>Monterey Bay</td>
<td>63</td>
<td>Specific Conductance</td>
<td>Monterey County Water Resources Agency</td>
<td>Continuous to annual</td>
<td>[Zidar and Feeney, (2019)]</td>
</tr>
<tr>
<td>Ocean City</td>
<td>15</td>
<td>Chloride</td>
<td>Town of Ocean City</td>
<td>Monthly</td>
<td><a href="https://oceancitymd.gov/oc/">https://oceancitymd.gov/oc/</a></td>
</tr>
<tr>
<td>NC Coastline</td>
<td>10 (within 5 km from the shoreline)</td>
<td>Specific Conductance</td>
<td>North Carolina Division of Water Resources</td>
<td>Sporadically</td>
<td><a href="https://ncdenr.maps.arcgis.com/apps/webappviewer/index.html?id=6a0293762cf249ed92b657bd9b7465cf">https://ncdenr.maps.arcgis.com/apps/webappviewer/index.html?id=6a0293762cf249ed92b657bd9b7465cf</a></td>
</tr>
<tr>
<td>SC Coastal towns</td>
<td>11</td>
<td>Specific Conductance</td>
<td>South Carolina Department of Natural Resources</td>
<td>Hourly (Conductivity sensors)</td>
<td><a href="http://hydrology.dnr.sc.gov/saltwater-intrusion-monitoring.html">http://hydrology.dnr.sc.gov/saltwater-intrusion-monitoring.html</a></td>
</tr>
<tr>
<td>TX Brazoria County</td>
<td>10 (not all wells are)</td>
<td>Specific Conductance</td>
<td>USGS and Brazoria County</td>
<td>Once in every four years (until 2015)</td>
<td><a href="http://www.bcgroungwater.org/images">USGS portal, http://www.bcgroungwater.org/images</a></td>
</tr>
<tr>
<td>氯化物 (on the shoreline)</td>
<td>Chloride</td>
<td>/bcg/documents/BCGCD_MISSION.pdf</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Chapter 3: Time-lapse geophysical measurements for monitoring coastal groundwater dynamics in an unconfined aquifer (Manuscript 2)

*Under Review in Groundwater Journal*

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Abstract

The coastal zone, where land meets water, is hydrodynamically active because of the interplay of various hydrological controls and the interaction of variable-density fluids. These forces change over time, causing the system to be in a state of dynamic equilibrium. Saltwater intrusion and submarine groundwater discharge are the major hydrological processes in coastal aquifer systems and are interdependent. Therefore, monitoring these two major and complex processes in coastal aquifers is essential for sustainable coastal zone management but also is challenging. Herein, we use non-invasive geophysical techniques i.e., the time-lapse electrical resistivity survey method, in conjunction with groundwater monitoring to demonstrate the application of the geophysical techniques in monitoring the coastal hydrodynamics in varying time scales and in different hydrogeological settings. The southern coast of Rhode Island serves as the test site. Our approach provides insight into the saltwater–freshwater interaction in heterogeneous aquifers. From our data, we developed baseline saltwater intrusion maps that show seasonality in the saltwater intrusion. The time-lapse electrical resistivity survey method is applicable for rapidly estimating fresh groundwater discharge. Though the results are site-specific, our techniques in monitoring these two hydrological processes using geophysical techniques have broader implications.
Keywords: Coastal Aquifer, Hydrodynamics, Saltwater Intrusion, Submarine Groundwater Discharge, Geophysical Techniques, Time-lapse electrical resistivity survey

1. Introduction

The dynamics of coastal aquifers are complex and the aquifer-ocean interaction is an emerging topic in hydrology with implications for water resources and ecosystem management (Day-Lewis et al., 2006). There are several environmental and anthropogenic factors affecting the availability of freshwater in coastal aquifers. The dynamic hydrological forces that control freshwater availability in near-shore aquifers act on vastly different spatial and temporal scales. Those major environmental forces include waves and tides, seasonal and interannual changes in sea level, seasonal change in terrestrial groundwater level caused by evapotranspiration and precipitation effects, and episodic storm surge events, among others (Heiss and Michael, 2014; Robinson et al., 2018; Santos et al., 2021). Over-pumping aquifers to feed the ever-increasing coastal population is a significant anthropogenic force. Climate change, the rise in sea level, and the increase in frequency and intensity of coastal storms, together with urbanization, stress the hydrodynamically active coastal aquifers with regard to both quantity and quality of groundwater. Variable density effects (saline water is denser than freshwater) further complicate the coastal hydrodynamic processes (Robinson et al., 2018). Therefore, it is crucial to understand how the groundwater and biogeochemical changes in coastal areas at spatial and temporal scales may impact water security and ecosystem management (Michael et al., 2017).
Figure 3-1: Schematic diagram showing groundwater dynamics in an unconfined coastal aquifer (Modified after Robinson et al. 2018 and Michael et al. 2005). SWI refers to Saltwater intrusion, fSGD refers to the fresh component of submarine groundwater discharge and UPS refers to upper saline plume driven by tide and wave actions.

Saltwater intrusion (SWI) is the landward movement of saline seawater into fresh groundwater in coastal aquifers. It is a significant problem for supplying drinking water to the ever-increasing coastal population and sustaining critical ecosystem functions that depend on freshwater (Michael et al., 2017; Werner et al., 2013). Virtually all coastal aquifers are susceptible to saltwater intrusion. Still, the degree of susceptibility depends on
the dynamics of the coastal processes, particularly the rate of sea level rise, groundwater flux from terrestrial aquifers, which is a function of groundwater recharge and abstraction, and the land surface topography (Michael et al., 2013). Due to the difference in density between saline water and freshwater, the freshwater buoys on the saline water, but when the groundwater head and flux decline, the saltwater starts encroaching into the aquifers making the fresh groundwater saline. There are several incidents of groundwater salinization in different parts of the world, such as Monterey Bay in California (Goebel et al., 2019; Pidlisecky et al., 2016), Biscayne aquifer in Florida (Costall et al., 2018; Prinos, 2019), places in Northern Africa (Agoubi, 2021), in Asia (Satish Kumar et al., 2016), in Australia (Werner, 2010), and in Europe (Cai et al., 2014). Steps must be taken to slow down the rate of intrusion, which prerequisites the current understanding of freshwater and saltwater distribution and how the various drivers control it.

Submarine groundwater discharge (SGD) is any subsurface flow of water from the porous seabed or conduits to coastal waters (Burnett et al. 2003; Moore 2010). SGD, which connects terrestrial and marine systems, is largely invisible but an active hydrological process that transports nutrients, chemicals and metals, and contaminants of emerging concerns to the coasts and estuaries. While the recirculation of saline water through the seabed is driven by the contrast in density, fresh groundwater discharge from the terrestrial environment toward the ocean is controlled by the head gradient (Darcian flow). Collectively, these two processes are termed total SGD (Taniguchi et al. 2002). Fresh SGD (fSGD) is a source of new nutrients, whereas saline SGD often releases recycled nutrients from seabed sediments (Santos et al., 2021). The recirculating brackish water constitutes
the greatest proportion of the total SGD globally. In total, the estimated global SGD flux is 3 to 4 times greater than the global freshwater fluxes from rivers into the oceans (Kwon et. al. 2014) and SGD influx results in a larger amount of dissolved chemicals transported to the ocean than rivers (Rodellas et al., 2015). The spatial scale of the interface through which SGD is occurring ranges from meters to several kilometers.

The SGD flux, to some extent, is essential as fisheries depend on the nutrients. Typically, the coastal areas' high population density can enhance the influx of anthropogenic pollutants into coastal waters. An excessive nutrient and anthropogenic pollutant load can result in algal blooms and then hypoxia, which is a significant driver of biogeochemical processes (Wang et al. 2020). For instance, the infamous ‘dead zone’ in the Gulf of Mexico is caused by a large amount of agricultural nutrients transported through surface water and groundwater into coastal waters (NOAA, 2021; Santos et al., 2009). While the role of SGD is recognized, its quantification is challenging because of the difficulty in measurement (Befus et al., 2017) due to heterogeneity and diffused and temporally variable seepage; therefore, it is largely overlooked in the coastal nutrient budget.

SWI and SGD are complementary processes (Robinson et al., 2018; Werner et al., 2013), and there is a delicate balance between hydraulic and density gradients in groundwater and seawater (Taniguchi et al., 2002) that drives both processes (Figure 3-1). When the fresh groundwater discharge into the ocean or estuaries increases, the saltwater interface moves towards the ocean (Kuan et al., 2019), but when the flux is less, the reverse happens. The major drivers controlling SWI and SGD can be classified into climatic and non-climatic.
Climatic drivers include the variability in precipitation and change in groundwater recharge, rising sea level, and increased intensity and frequency of storm surge over-wash. Over-extraction of groundwater from coastal aquifers and the development of navigation and irrigation canals are anthropogenic drivers of SWI. Climate change and variability significantly alter the hydrological cycle from a regional to a global scale (Wu et al. 2013). This alteration affects SGD via a series of events such as variation in precipitation rate, melting glaciers, and enhanced evapotranspiration due to warmer climates, and sea level rise. Because of the complex physical, geochemical, and mechanical processes involved, the SWI and SGD studies remain open research topics despite being studied for several years. Moreover, studying the interplay of SWI and SGD at both spatial and temporal scale provides insights in this unique coastal hydrological system.

Various methods exist to track SWI and SGW. Installing observation well network is a traditional and more accurate method for SWI research (Prinos, 2019). Still, it is cost prohibitive and is confined at a point scale lacking to capture the subsurface complexities. Also, the density of the monitoring well network is sparse (Barlow, 2003), and the sampling frequency and depth are not consistent (Panthi et al., 2022). Geochemical tracers are often used for SWI (Habtemichael and Fuentes, 2016; Xu et al., 2016) and SGD monitoring (Adyasari et al., 2021; Dimova et al., 2011). Previous studies relied on seepage meters to monitor the SGD (Michael et al., 2003; Taniguchi et al., 2003), which can measure both saline and fresh SGD at a point scale. Recently, airborne thermal infrared imaging technique has been used to identify SGD plumes, such as in a study on Long Island, New York (Tamborski et al., 2015). Researchers have also used several modeling frameworks to simulate SWI (Langevin and Zygnerski, 2013; Paldor and Michael, 2021) and SGD
fluxes (Befus et al., 2017; Sawyer et al., 2016) at various spatial scales. But those models need high-resolution geological and groundwater head and salinity observation datasets for calibration.

These limitations open opportunities for geophysical techniques which are commonly used to complement the sparse monitoring well data and to study subsurface heterogeneity (Pidlisecky et al., 2016). Non-invasive geophysical techniques are widely employed for monitoring and mapping SWI at larger spatial and temporal scales. Recently, these techniques also have been used for quantifying fresh SGD. For example, Paepen et al. (2020) combined electrical resistivity imaging (ERI) and electromagnetic techniques to detect the zone of fresh SGD. Hermans and Paepen (2020) proposed a new method which combined land based and offshore resistivity images for characterizing SWI and SGD. These methods are qualitative and can identify the zone of fresh SGD, but do not calculate the fresh SGD flux. Overcoming this issue, Dimova et al. (2012) used time-lapse ERI to calculate the fresh SGD flux in a volcanic coastal aquifer in Hawaii Island. They measured the electrical resistivity using dipole-dipole geometric array that limits the measurement accuracy at deeper aquifer. Together, these studies show that the research domain of geophysical techniques is expanding from SWI monitoring to SGD calculation; however, they lack testing in more diverse hydrogeological settings. Also, there is a general lack of characterizing the vertical distribution of salt and freshwater and how it responds to terrestrial hydrological controls at a regional scale.
We employed the 2-D ERI technique in conjunction with ground observation borehole data to investigate the distribution of fresh and saltwater in unconfined coastal aquifers and the factors controlling the distribution of freshwater and saltwater. Our objectives were two folds. First, identify the locations and depths of fresh groundwater and saline seawater in coastal aquifers and their responses to various drivers, including tidal forces and seasonal changes. Secondly, to calculate the fresh component of SGD flux using geophysical measurements. Tracking the status of saltwater intrusion and the SGD flux helps to implement SWI control measures. Though the results are somewhat site-specific, our approach and methods are replicable to other unconfined and glacial deposit coastal aquifers, and the results have broader implications, including for ecosystem management and water security.

2. Hydrogeology of the test site

We chose the southern coast of Rhode Island to investigate the hydrodynamics of an unconfined aquifer. The surficial deposits in the study area are mainly glacial outwash deposited by meltwater and glacial till deposited by the movement of glaciers. Stratified outwash deposits consist of well-sorted, layered sediments ranging in size from clay to gravel deposited by glacial meltwater and overlies the till in the valleys and lowlands in coastal areas. A thin, discontinuous layer of till deposited directly on the bedrock by glacial ice is composed of a poorly sorted mixture of sediments ranging in size from clay to boulders. The bedrock topography, that ranges from the surface outcrop to 30 m deep from the land surface, is irregular and formed by seaward sloping bedrock valleys (Masterson et al., 2006). Further, the rate of sea level rise, i.e. 2.5 mm per year (Goddard et al., 2015), in the northeastern
U.S. is higher than the global average because of the isostatic rebound (Karegar et al., 2016; Talke et al., 2018).

In the study area, groundwater is the only source of drinking water for the coastal communities (Cox et al., 2019). There is a higher demand for groundwater in summer for two reasons: higher evapotranspiration and a seasonal influx of tourists (Wild and Nimiroski, 2005). Groundwater in the study area is stored and transmitted through surficial sediments and the underlying bedrock. The groundwater recharge rate in this part of Rhode Island ranges from 335 to 710 mm/year, which is approximately 50 to 55% of the annual precipitation (Masterson et al., 2006). Besides direct groundwater discharge to the ocean, there are a few coastal salt ponds where terrestrial groundwater discharges into. However, fresh groundwater discharge's pattern and flux to these coastal ponds are mainly unknown (Masterson et al., 2006). There is currently no evaluation of how the drinking water in the study area is or will be affected by the SWI. Importantly, the study site is representative for other formerly glaciated regions in the New England region and around the world. That is, the coastal aquifers in Rhode Island are characterized by glacial deposits overlying bedrock, resulting in large spatial heterogeneity in hydrogeological parameters. Besides the New England region of the US, similar geologic settings exist in parts of northern Europe and Asia.

We chose three distinct hydrogeological test sites (Figure 3-2) to develop a better understanding of the heterogeneity in coastal hydrodynamics. These are (1) a barrier island, where saltwater bodies sandwich freshwater, (2) an inland site, where SWI is not expected;
and (3) a beach-inland site, where continuous discharge of groundwater from an inland aquifer directly to the ocean dominates.

![Map of the field sites along the southern coast of Rhode Island.](image)

Figure 3-2: Map of the field sites along the southern coast of Rhode Island. The upper left inset map shows the test site on a regional map. The aerial images of the test sites correlate with red stars on the map. The yellow-filled circles show the location of observation wells.

3. Methods

3.1 ERI measurements and data processing

Electrical resistivity imaging (ERI) surveys are particularly well suited for characterizing saltwater and freshwater distribution in aquifers, as they are sensitive to the changes in subsurface electrical conductivity caused by the change in fluid salinity in the pore space (Goebel et al., 2017; Singha et al., 2022). The capabilities of the electrical method have
been demonstrated in a number of recent studies (Butler, 2005; Martínez et al., 2009; Ronczka et al., 2015; Goebel et al., 2017). In the ERI survey, a current is introduced in the ground using pairs of buried electrodes (~ 30 cm below the ground surface). Another pair of electrodes measure the electrical potential drop between the electrode pair (voltage). Applying Ohm's law calculates the resistance once the current and voltage are known. The resistivity of the subsurface is estimated by inverting the apparent resistivity data from the electrode pairs. This method provides 2D and 3D images of the electricity resistivity distribution and, thereby, a visualization of the water table depth, the aquifer architecture, and the saltwater/freshwater interfaces, if present.

We used a SuperSting-R8 - 56 electrodes multichannel ERI instrument to produce 2-D resistivity profiles. The details on electrode spacing, length, and depth of the survey transect are listed in Table 1. For our sandy beach sites, we buried the electrodes and wetted the ground around the electrodes with a slurry of clay (cat litter) and saltwater to minimize the contact resistance to a recommended threshold (5000 Ω). The Wenner-Schlumberger geometrical configuration was used for all the measurements to obtain a high horizontal resolution. The elevation of the land surface was measured using a Real Time Kinematic (RTK) Global Positioning System (GPS) with the reference of the vertical datum NAVD-88 for the topography corrections. We employed the SimPEG geophysical forward modeling inversion code (Cockett et al., 2015) to recover the subsurface distribution of electrical resistivity. Before the forward modeling, we filtered and removed any anomalous measurements from the raw data.
We collected the ERI measurements in three different seasons: spring (high groundwater level), summer (declining groundwater level), and fall (low groundwater level) to record the potential factors controlling saline water distribution. Moreover, to capture the impact of tidal pumping, we conducted the time-lapse continuous resistivity survey in a tidal cycle.

Table 3-1: Electrical resistivity imaging survey locations details. Refer to Map 1 for the spatial distribution of the transects. *Nearest well on D-D’ transect.

<table>
<thead>
<tr>
<th>Transect (Reference Figure 3-2)</th>
<th>Location</th>
<th>Length of transect</th>
<th>Electrode spacing</th>
<th>Depth of survey</th>
<th>Distance to the ocean/lagoon</th>
<th>Distance to the nearest monitoring well</th>
</tr>
</thead>
<tbody>
<tr>
<td>A-A’</td>
<td>Barrier Island site</td>
<td>178.5 m</td>
<td>3.25 m</td>
<td>35 m</td>
<td>20 m</td>
<td>1 m</td>
</tr>
<tr>
<td></td>
<td>(East Beach, Charlestown, RI)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B-B’</td>
<td>Inland site</td>
<td>220 m</td>
<td>4 m</td>
<td>44 m</td>
<td>500 m</td>
<td>50 m</td>
</tr>
<tr>
<td></td>
<td>(Charlestown, RI)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C-C’</td>
<td>Beach Inland site, shore-perpendicular</td>
<td>275 m</td>
<td>5 m</td>
<td>55 m</td>
<td>22 m</td>
<td>1 m*</td>
</tr>
</tbody>
</table>
3.2 Precipitation data and well observations

Precipitation data was retrieved from CHIRPS, which blends ground-level observation with satellite products (Funk et al., 2015). The CHIRPS gridded data has 0.05° spatial resolution and daily scale temporal resolution. We included the data from 1992 to 2021 for long average monthly precipitation, and based on these values, computed the percentage deviation (anomaly) in monthly precipitation for our study period (2019-2021).

We monitored an observation well near the inland test site (Figure 3-2) to analyze the groundwater fluctuation. The well was fully screened from the water table throughout to the bottom. We instrumented the well with Solinst® Levelogger with a pressure transducer, temperature, and electrical conductivity (E.C.) probes that autocorrect the E.C. for temperature. The acquisition time interval for all parameters was sub-daily. A barologger at the surface near the well location was used to compensate for the effects of atmospheric
pressure changes. Also, a temporary observation well was installed along the transect D-D' (Figure 3-2) to monitor the groundwater level and salinity due to tidal pumping. The sampling frequency of the logger (Solinst® Levelogger) was 6 minutes to match the NOAA's tide gauge at Newport, Rhode Island (Station ID: 8452660).

3.3 SGD calculation technique

The fSGD was calculated on the D-D' shoreline parallel transect (Figure 3-2) by taking two resistivity profiles from contrasting tidal stages: one from the high tide when the saltwater-freshwater interface migrates landward and one during low tide when the interface migrates seaward i.e., in a time when the hydraulic gradient between groundwater and sea level enables fSGD. Fluid conductivity is the function of salinity. The inverted resistivity profiles were first transformed to the pore-fluid conductivity using Archie's law (Archie, 1942) and a site-specific formation factor (Eqn.1). The formation factor is the ratio of bulk resistivity to fluid resistivity.

Archie's law applies only to clay-free porous media, such as in the current study, and relates intrinsic parameters that represent the sediment in an aquifer to the bulk and fluid resistivity:

\[ R_b = R_f \alpha n^{-q} S^{-k} \]  \hspace{1cm} (1)

where \( R_b \) is the bulk resistivity data from the field survey, \( R_f \) is the porewater (fluid) resistivity, \( S \) is the degree of fluid saturation, \( k \) is the saturation factor, \( n \) is the porosity of the total matrix, \( q \) is the cementation factor of the sediment and \( \alpha \) is the coefficient of
effective electrical tortuosity. All these parameters are dimensionless. For the porosity, we collected aquifer samples from three locations below the water table and analyzed them in the laboratory. By analyzing the grab sediment samples, the average porosity \( n \) was obtained to be 0.35 for the transect D-D'. Moreover, both \( \alpha \) and \( q \) were assumed to be 1, which are typical parameters for comparably high hydraulic conductivity, unconsolidated sand and gravel deposits (Comte and Banton, 2007; Herckenrath et al., 2012; Cardenas et al., 2015). We considered the water-saturated system only, so \( S = 1 \). Furthermore, we assumed water is the only conducting media occupying the pore space and we neglected small, localized pockets of clay and peat deposit at the test site. Because of the high degree of petrophysical homogeneity of the test site, the bulk resistivity is directly correlated to the fluid resistivity and inversely to fluid conductivity.

The fSGD is estimated using a salinity mass balance box model (Eqn. 2) that was developed and implemented by Dimova et al. (2012) on a Hawaii island.

\[
V_{sal} * S_h = \left[ V_{sal} + V_{fSGD} \right] * S_l \quad (2)
\]

where \( V_{sal} \) is the area of preferential flow paths observed during the low tide, \( V_{fSGD} \) is the total volume of fresh groundwater discharge in a tidal cycle from the flow path, \( S_h \) and \( S_l \) are the average salt concentrations in the flow path locations during high and low tide, respectively.
The model is based on several assumptions. First, the principle of conservation of salt mass is applied. The entire volume of the fSGD is exported to the surface so that the volume of the preferential flow path box remains constant. Another assumption is that saltwater dispersion during the tidal water mass exchange is not considered. Also, we observed that there was 20 minutes of time to propagate tidal signals from the ocean to the near-shore groundwater observation well.

4. Results

4.1 Groundwater response to precipitation: Inland aquifer

![Graph showing groundwater response to precipitation variability.](image)

Figure 3-3: Groundwater response to precipitation variability observed in an inland monitoring well. The blue columns on the precipitation indicate higher than normal while the red indicates lower than normal precipitation. The tan-colored shaded three vertical columns represent the ERI measurement periods.

The groundwater level responds to the precipitation, with some time lag, and seasonal change in evaporation effects (data not shown). Figure 3-3 shows the temporal variability
i.e., deviation from long term average in groundwater level and its response to precipitation in the inland aquifer (site reference in Figure 3-2) where there is no signature of daily and seasonal change in sea level (data not shown). Also, the observation well was outside the radius of influence of productions in the study area. Three ERI measurements were collected between 2019 and 2021 (see Sec. 4.2). It is noted that there was a significant flash meteorological drought in the study area in the summer and fall of 2020, which resulted in a drop in groundwater levels at a regional scale (Lombard et al., 2021). In response to the drought, the precipitation declined from the long-term mean in May 2020 and continued declining until the end of October 2020. The peak deficit from the long-term monthly mean (>60%) was reached in September (Fig. 2). With about a month of lag time, groundwater levels followed the downward precipitation trend. The groundwater level declined by 0.78 m relative to the level just before the drought (April 2020). The groundwater level recovered to some degree by winter. Together, the data set demonstrates the persistent impact of meteorological drought on hydrological drought.
4.2 Temporal variability in electrical resistivity - Inland aquifer

Figure 3-4: Seasonal electrical resistivity distribution in the inland aquifer (transect B-B', Fig 1). (a) summer of 2019, (b) fall of 2020 and (c) spring of 2021. The blue color indicates water, and the dark red color is the unsaturated zone. The black dashed lines are the water tables measured in an observation well.

The bulk electrical resistivity distribution in the inland aquifer (site reference in Figure 3-2) is depicted in Figure 3-4. The minimum resistivity value was above 100 Ω.m (dark blue), and the maximum resistivity was below 10000 Ω.m (dark red). Electrical conductivity measurements at a monitoring well (15 m deep) at this location confirms that the water is fresh. The static water table depth is at ~4 m below the surface. Depth to water table becomes shallower towards inland (left side of the transect), indicating the direction of groundwater flow is towards the lagoon/salt pond (right). Because of the presence of a localized silty clay layer on the section (0 to 50 m horizontal axis), which was confirmed by coring during drilling, the top surface on the profile is relatively low resistive than the rest of the profile.
The resistivity profiles clearly show the impact of drought on declining groundwater levels. The maximum water level was observed in pre-drought summer of 2019 (Figure 3-4a), and the minimum in fall of 2020 (figure 3-4b) i.e., at the height of the drought period. Figure 3-4c shows that the groundwater level recovered in the following spring, although not quite to the pre-drought level (Figure 3-4a). These results correspond with the observed well data shown in Figure 3-3.

4.3 Temporal variability in electrical resistivity - Barrier Island

![Electrical resistivity distribution at the barrier island](image)

Figure 3-5: Electrical resistivity distribution at the barrier island (transect A-A', Figure 2). (a) summer of 2019, (b) fall of 2020 and (c) spring of 2021. A salt pond is on the left side, and the Atlantic Ocean is on the right. The orange to dark red colors indicates brackish water; the light green color indicates freshwater, and the black color is the unsaturated zone.

Figure 3-5 shows the spatial and temporal distribution of electrical resistivity on a barrier island. The minimum resistivity value was below 1 Ω.m indicating seawater while the
maximum value was above 10000 Ω.m indicating dry sand at the top layer. The freshwater layer on the island is very narrow and shallow i.e., it extends as deep as 3 m from the land surface near the 140 m mark which coincides with the location of a berm. Away from the berm, the depth to and the thickness of the freshwater lens decreases rapidly on either side. In the pre-drought summer of 2019 (Fig. 3-5a), the layer of freshwater across the island was continuous. During the peak of the drought in Fall of 2022, the freshwater lens was squeezed horizontally and vertically. The saltwater invasion rate is higher towards the salt pond side because of the lower elevation. The freshwater lens recovered and returned to a normal condition in response to the high rainfall in the following spring (2021). In total, the maximum amount of freshwater was available during the highest recharge time (Spring of 2021, Figure 3-5c), while the minimum amount of freshwater was available during the drought time (Fall of 2021, 3-5b). Overall, the time-lapse ERI demonstrates the freshwater lens response to variable recharge conditions and the applicability of geophysical measurements in capturing the extent of the freshwater lens on this barrier island.
4.4 Saltwater distribution variations at the Beach-Inland Site

Figure 3-6: Electrical resistivity distribution at the Beach-Inland site. A perpendicular (C-C’) and parallel (D-D’) transect to the shoreline was evaluated. Refer to Figure 3-2 and Table 3-1 for the survey transect details. Transect C-C’ shows the inland propagation of the saltwater, while the parallel transect shows the fresh groundwater flow paths from the aquifer to the ocean. The two insets show the geological lithologies identified during the installation of temporary boreholes on the survey transect.

Figure 3-6 shows the electrical resistivity distribution in two unique coastal settings. The transect (C-C’) extends from the ocean towards inland, tracing the flow of freshwater. The transect D-D’ is in the near-inter-tidal zone (18 m inland from the high tide line) and parallel to the shoreline. It coincides with the saltwater-freshwater mixing zone. Subsurface heterogeneity was observed in form of a layer of clay and fine silt embedded in the predominantly sand and cobble/gravel sediments that made up most of the deposits along transect D-D’. This area of subsurface heterogeneity was located between 50 and 75 m
from the transect’s origin. In it, the clay has high conductance and thus lower resistance properties, as expressed by the deep blue colors in the resistivity profile D-D'. Besides that, the shoreline parallel transect clearly depicts what could be interpreted as preferential freshwater flow paths directed to the ocean. The quasi steady-state interface of salt and freshwater in the shoreline parallel transect is 10-15 m below the land surface (Figure 3-6b).

Following the Ghyben-Herzberg principle (Ghyben, 1889; Herzberg, 1901), the saltwater interface depth increases with the increase in the freshwater head in the inland direction. This is clearly reflected by the increasing reddish colors, reflecting the higher resistivity of the freshwater layer in the landward direction in transect C-C' (Figure 3-6). The interpreted depth to water table (freshwater) was ~5 meters at the center of the shoreline-perpendicular transect (C-C'), while the water table at the intersection of shoreline parallel and perpendicular transect was at 2.13 m in high tide and 2.28 m at low tide. Note here that there was ~4 m of elevation difference between these two locations. This confirms well with the depth estimate according to Ghyben-Herzberg principle.
4.5 Tidal pumping at the Beach-Inland Site

Figure 3-7: (a) Fluid salinity distribution, reported as total dissolved solids (TDS), along the shoreline of the Beach-Inland site. (b) Depth to groundwater in a temporary monitoring well installed along the survey transect. TDS data show the response of the coastal aquifer to tidal pumping. T1, T2, T3, and T4 correspond to end times of the ERI measurements during the tidal stages. Boxes A, B, C, and D are considered the preferential freshwater flow paths during low tide cycle T4.

Figure 3-7 shows the salinity distribution, reported as total dissolved solids (TDS), along the shoreline in a near-intertidal zone of the beach-inland site (site reference in Figure 3-2). The observed changes in freshwater to saltwater flows were attributed to the tidal pumping. Saltwater intrusion happens at high tide, and fresh groundwater discharge occurs at low tide because of the change in hydraulic head between groundwater and sea level. During the falling tide, the porewater conductivity along the resistivity profile decreased,
which started increasing with the rising tide. The saltwater-freshwater interface is diffuse but about 10 to 15 m deep from the land surface. Also, we identified four major locations, identified by rectangular boxes A to D in Fig. 6a. Within these boxes, there is a sharp change in seawater proportion from high tide to low tide. These boxed areas are considered the preferential flow paths for fresh groundwater discharge toward the ocean. On the interpreted salinity profiles, the maximum salt concentration was found to be less than 10 mg/L, signifying the discharge of fresh groundwater in the test site.

4.6 SGD estimation

Table 3-2: Fresh SGD calculation at the beach-inland test site. The symbols correspond to the salinity mass balance box model equation 2. The locations of boxes A, B, C, and D correspond to Figure 3-7.

<table>
<thead>
<tr>
<th>Box</th>
<th>Box Area (m²)</th>
<th>Average salinity, high tide (S_h) (mg/L)</th>
<th>Average salinity, low tide (S_l) (mg/L)</th>
<th>(S_h - S_l) (mg/L)</th>
<th>fSGD per tidal cycle and box (m³)</th>
<th>fSGD per m² per tidal cycle (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>18</td>
<td>0.70</td>
<td>0.50</td>
<td>0.20</td>
<td>7.2</td>
<td>0.40</td>
</tr>
<tr>
<td>B</td>
<td>20</td>
<td>0.55</td>
<td>0.35</td>
<td>0.20</td>
<td>11.4</td>
<td>0.57</td>
</tr>
<tr>
<td>C</td>
<td>98</td>
<td>0.35</td>
<td>0.25</td>
<td>0.10</td>
<td>39.2</td>
<td>0.68</td>
</tr>
<tr>
<td>D</td>
<td>165</td>
<td>0.22</td>
<td>0.14</td>
<td>0.08</td>
<td>94.3</td>
<td>0.57</td>
</tr>
<tr>
<td></td>
<td><strong>Total fSGD per tidal cycle</strong></td>
<td></td>
<td></td>
<td></td>
<td><strong>152.1</strong></td>
<td></td>
</tr>
<tr>
<td></td>
<td><strong>Total fSGD per day</strong></td>
<td></td>
<td></td>
<td></td>
<td><strong>304.2</strong></td>
<td></td>
</tr>
</tbody>
</table>
Using the fluid salinity contrast between the high and low tide measurements, we identified four major preferential flow paths shown as rectangular boxes A to D (Figure 3-7d). The average salinity in high and low tides in each box was used to calculate the fresh groundwater discharge rate during the falling tide period. We assessed a total fresh SGD of 152.1 cubic meters in a tidal cycle, equivalent to 2.3 cubic meters per day per meter length of the shoreline (Table 3-2). While calculating the fSGD, we truncated a subsection of the transect (0 to 90 m) because of the presence of a clay layer (Figure 3-6) where little to no flow of freshwater flow was indicated by the ERI data.

5. Discussion

5.1 Spatial and temporal distribution of saltwater and freshwater

Saltwater intrusion at the test sites studied herein showed distinct behavior due to their unique hydrogeological characteristics. We detected the presence of saltwater in the beach-inland aquifer but not in an inland aquifer which is ~500 m far from the saltwater body (salt pond, at least not up to the depth of 40 m from the land surface). Also, the inland aquifer does not response to the diurnal tide cycle with changes in both head and salinity, while changes in head were measured in the barrier island aquifer in responds to the diurnal cycle. During the 2020 summer and fall drought, the water level in the inland aquifer dropped by ~1 m according to the interpreted resistivity profile. This observation was corroborated by measurements in an observation well. However, there is no signal of
The seasonality in the inland aquifer groundwater level is attributed to evapotranspiration effects. On the barrier island, there is a seasonal variation in subsurface resistivity. We conjecture that the decrease in fresh water, as measured by a decrease in resistivity, is linked to the drought that started in May 2020 and extended until October 2020. The drought squeezed the freshwater lens vertically and horizontally. Also, the absence of a sharp boundary between the fresh groundwater and brackish water suggest that a more expansive mixing zone formed in the barrier island aquifer.

5.2 Predictive performance of the fSGD model

The hydraulic head difference between the terrestrial groundwater and seawater is the primary driving force of tidally driven groundwater flow in coastal sediments. The shoreline parallel (D-D’) resistivity profile at low tide exposed the freshwater preferential flow paths. During high tide, flow through these areas is unlikely because of gradient reversals. The ERI measurement along D-D’ did not give indication if and how far saltwater might have intruded inland along those preferential flow path. It would have taken tide cycle measurements along C-C’ to shed light on this potential interaction.

There are several factors contributing to uncertainty in the ERI-based fSGD calculations. Dimova et al. (2012) state that these uncertainties arise from the resistivity measurements, data inversion, and estimation of the formation factor when transforming bulk resistivity to fluid conductivity. To reduce these uncertainties, we removed the measurement's inconsistent data (5%) and controlled the RMS and L2 misfit to 10%. We further reduced the uncertainty from the formation factor by installing a data logger (Solinst TLC) into a
temporary well to measure the salinity at the top of the water column, which has less than a 5% instrumental error.

Previously reported fSGD fluxes in the Northeastern region are consistent with our estimates. Michael et al. (2005) in a seepage meter study in Waquoit Bay in eastern Massachusetts found that the average fresh groundwater discharge is in the range of 3.5 m$^3$/day per meter length of shoreline, with seasonal fluctuations, the highest being in summer and the lowest in the winter. This seasonality is out of phase with the annual recharge cycle, possibly due to the lag between peak recharge and the arrival of the groundwater at the coast. In Waquoit Bay, with sandy unconfined aquifer as in our test side and with ~ same amplitude of tides, Mulligan and Charette (2006a) estimated a discharge of 0.6 to 5.6 m$^3$/day per meter length of shoreline using Darcy’s law. However, in the Pettaquamscutt estuary in Rhode Island, Kelly and Moral (2002a), using Radium data and a box model, estimated the seasonal range of groundwater flux between 42 to 420 m$^3$/day per meter length of the shoreline. This higher degree of fSGD flux is likely due to the larger size of the aquifer compared to the one investigated herein. There are very few studies relying on geophysical methods to quantify the groundwater discharge. Dimova et al. (2012) developed and utilized an ERI method on the Hawaii island and estimated fSGD to be 65 m$^3$/day per meter length of shoreline in an aquifer consisting of volcanic rock. Overall, the estimation of fSGD in our study area is within the range previously calculated in Waquoit Bay but tends to be lower compared to other studies. Differences could be related to the aquifer type, its size and prevailing hydraulic gradients, besides other site-specific factors.
5.3 Future research questions

Coastal aquifer-ocean interactions are emerging hydrologic research topics and could benefit from breakthroughs in geophysical method developments. This research generated a benchmark data set on saltwater and freshwater dynamics at the land-ocean interface. But additional studies are required to better conceptualize hydrogeologic processes in this dynamic environment. For instance, consistent geophysical surveys over consecutive years could help to establish trends and eventually a reliable prediction of the SWI rate. In general, SWI should be studied with comparatively high temporal resolution (~ monthly) because changes in precipitation and, consequently recharge, as demonstrated by our barrier island aquifer test site during a drought period, have a measurable impact on coastal aquifers.

Similarly, there is evidence that fSGD changes with the seasons as these change the hydraulic head in terrestrial groundwater (Michael et al., 2005). However, there are several lines of evidence that seasonal changes lag several months before being expressed as fSGD at the land-water interface, such as in Waquoit bay in eastern Massachusetts (Michael et al., 2005; Mulligan and Charette, 2006b), and in Rhode Island (Kelly and Moran, 2002b). This is particularly true for regions with a higher evapotranspiration rate.

Furthermore, extending the resistivity profiles along the shoreline will provide a regional perspective for fSGD, capturing spatial heterogeneity in the subsurface, particularly in identifying the preferential flow paths. Utilizing the geophysical data, such as water table depth, saltwater concentration, and lithological heterogeneity, is important to develop high-resolution groundwater flow models capable of simulating fSGD under variable boundary conditions - specially to address variable recharge scenarios under future climates.
Depending on data density and spatial resolution, such groundwater flow models, ideally in combination with field measurements, might be helpful to identify potential locations of fSGD. Finally, the fSGD estimation by geophysical techniques needs further verification, especially in heterogeneous hydrogeological settings. Geochemical tools, such as Radium isotopic tracers and seepage meters, could further strengthen geophysical methods.

6. Conclusion

In this study, we demonstrated the application of time-lapse surface-based electrical resistivity imaging in monitoring coastal hydrodynamics. Our data assists in understanding in delineating the water table, mapping the extend of SWI, and calculating the freshwater component of the submarine groundwater discharge from an unconfined coastal aquifer. In this study area, we verified that the coastal (= barrier inland site) aquifer is hydrodynamically active and responds to the seasonal changes in recharge. The inland aquifer, only 500 m from coastal water, does not show any sign of salinization, but the groundwater level dropped by 0.8 m in response to a regional drought in 2020. A more pronounced drop in the water table elevation could lead to a reversal of the prevailing hydraulic gradient, possibly resulting in SWI under future, prolonged drought. In contrast, the freshwater lens on the barrier island showed a clear response and shrank during the drought event.

Geophysical techniques have been used for saltwater intrusion mapping in several locations in the past, but here, we presented a case study that demonstrates the use of time-lapse ERI for assessing fSGD. The results of this study provide a rapid assessment of fSGD over
comparably large areas, larger than what can be assessed using a conventional seepage meter. Hence, time-lapse ERI can assist in coastal zone management studies.

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Author’s contribution statement

Jeeban Panthi: Research design, literature review and data collection, formal analysis, and manuscript writing. Thomas Boving: Research design, manuscript writing and editing, funding acquisition. Soni M. Pradhanang: Manuscript review and writing, supervision, field coordination and funding acquisition. Mamoon Ismail: Manuscript writing and editing, data collection.

Declaration of competing interest

The authors declare no competing interests. All the authors have read and agreed on the content of the manuscript.
Data availability

The monitoring well data and raw electrical resistivity profiles have been uploaded to CUAHSI's Hydroshare data-sharing portal. Access will be provided after the manuscript is accepted.

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Supplementary Information I: Resistivity inversion forward model

Figure SI 1: Examples of ERT inversion model performance for the (a) inland test site, and (b) barrier island test site (Location reference in Figure 1)
Chapter 4: Delineating Bedrock Topography with Geophysical Techniques: An Implication for Groundwater Mapping (Manuscript 3)

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Abstract

Bedrock topography delineation is essential for shallow groundwater mapping because the bedrock surface is the lower boundary of the unconsolidated aquifer system and is difficult to map if covered by thick surficial deposits. Non-invasive geophysical techniques are suitable tools for quantifying the depth to bedrock at a single point location or the bedrock topography using an interpolation between multiple measurements, boings, outcrops, and knowledge of the local bedrock’s brittle and ductile structure. In this study, first, we developed a high-resolution bedrock topography map for the southern coast of Rhode Island, USA, with a wide range of available lithological data. Second, we employed the Horizontal to Vertical Spectral Ratio (HVSR) seismic method to develop a power-law regression between the resonance frequency and the depth to bedrock, and we demonstrated statistical techniques to refine the relationship. Like in most other formerly glaciated regions worldwide, the surficial deposits are glacial outwash and till. It was found that the predictive performance of HVSR was better for glacial outwash than till mixed outwash. In addition, we highlighted the importance of the HVSR technique in interpreting the electrical resistivity profiles for groundwater mapping in both inland and coastal aquifers. Though the quantitative results are site-specific, the approach and insights are generalizable to any unconfined aquifers.

Keywords: Geophysical technique, passive seismic, bedrock, groundwater, HVSR, glacial deposits, coastal aquifers

1. Introduction

Mapping the depth to bedrock is important for hydrogeological studies of complex aquifer systems. Bedrock is considered impermeable, except where fractured, and typically is the lowermost boundary of the unconsolidated aquifer (Bohidar et al., 2001; Liu and Tokunaga, 2019). Therefore, the thickness of the unconsolidated sediment is a major factor in determining the water-bearing capacity of an unconfined aquifer (Bohidar et al., 2001) and designing remediation for contamination sites (Brown et al., 2013). In fractured bedrock, groundwater infiltrates beneath the bedrock surface, and the fractures serve as the preferential flow paths. In addition, the bedrock surface's topography can determine local and regional groundwater flow patterns (Fairchild et al., 2013). Ultimately, the bedrock topography determines the thickness of overlying unconsolidated aquifers, especially in areas covered by glacio-fluvial sediments. Therefore, delineating the bedrock topography, the lower boundary of the unconsolidated aquifer, is of utmost importance for hydrogeological studies. The unconsolidated aquifers are the primary water source for many towns and farms, and they are in hydraulic contact with the underlying bedrock.
aquifers. Furthermore, the unconsolidated aquifers are vulnerable to contamination, such as nitrate pollution from wastewater discharge to the water table through septic systems leach fields, which are very common worldwide.

There are various methods to determine bedrock topography. The most common and traditional way is to drill boreholes, analyze the lithological stratigraphy from the borehole data, identify depth to bedrock, and extrapolate between boreholes. Though this is the most accurate technique, it is also the most expensive and limited in spatial resolution. For that reason, drilling alone generally is insufficient to generate high-resolution bedrock topography at regional scales. Statistical tools, such as spatial interpolation or kriging, can aid in producing regional bedrock maps. However, interpolation between data points from far apart drilling locations can lead to significant inaccuracies, especially in the heavily sculptured bedrock terrain of formerly glaciated landscapes. Many hydrogeologists utilize geophysical techniques to map the bedrock in lieu of an insufficient number of boring logs.

Surface geophysical methods are non-invasive, and of relatively low cost. Combined with ground controls, such as well-completion reports where lithology was reported, these techniques aid in developing hydrogeological models of the subsurface for groundwater flow and solute transport (Lane et al., 2008). The most commonly used geophysical techniques to map the bedrock topography include passive or active seismic surveys (e.g., Adly et al., 2017), electrical resistivity surveys (e.g., Cardarelli and De Donno 2017), gravity modeling (e.g., Ibrahim and Hinze 1972), and electromagnetic surveys (e.g., Anschütz et al. 2017).

An example of a passive seismic technique is the Horizontal-to-Vertical Spectral Ratio (HVSR), also known as Nakamura's method. HVSR relies on ambient passive seismic waves that resonate in the sedimentary layers with distinct resonance frequencies that may be used to model geologic structures, such as the bedrock surface (Nakamura, 1989; Sauret et al., 2015). HVSR has been widely used to map the depth of unconsolidated material in the subsurface, such as sediments above bedrock (e.g. Seiderman 2018, Tun et al. 2016), or to map the thickness of glacial ice (e.g. Picotti et al., 2017). The method relies on three-component seismometers to measure the time series of vertical and horizontal components of ambient seismic noises at the earth's surface. Ambient seismic noises, also termed microtremors, are primarily caused by oceanic waves, wind, and other meteorological sources (Bonnefoy-Claudet et al., 2006). They could also originate from human activities such as road traffic.

Electrical resistivity imaging (ERI) is another geophysical method that has been used extensively in geophysical and environmental investigations (Cardarelli and De Donno,
2017; Kumar et al., 2021), including for bedrock topography mapping in terrestrial (Cheng et al., 2019; Xu et al., 2016; Zhou et al., 2000) as well as in marine environment (Rinaldi et al., 2006). ERI has been used for mapping the overburden deposit also in coastal areas (Adepelumi et al., 2006), however the applicability of ERI alone for mapping bedrock topography in coastal areas is challenging (Cong-Thi et al., 2021) because of the exponential decrease of electrical signal with depth. Even if the boundary between fresh and saltwater is sharp, it appears as transitional on the processed image, mimicking bedrock topography. Because of this uncertainty, ERI in coastal zones requires ground-truthing (Zhou et al., 2000), or needs to be combined with another geophysical technique to improve interpretations of resistivity profiles.

HVSR and ERI techniques have been used in geological exploration for years; however, a significant challenge remains in integrating them for aquifer characterization. This paper demonstrates an approach of combining HVSR and ERI surveys for groundwater mapping in coastal aquifers. The specific research questions addressed in this study are: (a) What is the impact of limited borehole data points for bedrock topography mapping? (b) What are the regression relationships between the resonance frequencies and depth to bedrock? and (c) How comparable are ERI and HVSR techniques and how do they complement each other when mapping bedrock topography in a coastal aquifer? Based on the available well-completion reports, we developed a high-resolution map of the coastal zone bedrock topography of the study area and demonstrated the impact of limited data points in bedrock mapping. Our study's results were then compared with geophysical studies in the region (Johnson and Lane, 2016; Fairchild et al., 2013). Though the HVSR and ERI results are site-specific, the proposed approach of characterizing subsurface for groundwater mapping by integrating both the techniques is potentially transferable to other coastal and unconfined aquifers.

2. Study area and Hydrogeological setting

This study focused on the southern coast of Rhode Island (Figure 4-1). The region is dominated by thick glacial deposits on top of an irregular and fractured bedrock surface (Frohlich and Kelly, 1988; Masterson et al., 2006). The bedrock in the area is comprised mainly of two granitic units: the Narragansett Pier and Westerly Granite of the Pennsylvanian or post-Pennsylvanian age (Quinn, 1971). The granite is dark pink to pale gray and medium-grained (Moore, 1964) and is massively or locally inconsistently foliated. The surficial deposits are mainly stratified and well-sorted glacial outwash and sandy ablation deposits, a mix of unstratified and poorly sorted material known as till. A thin, discontinuous layer of subglacial till, deposited directly on the bedrock by glacial ice, is composed of a poorly sorted mixture of sediments ranging in size from clay to boulders. Stratified deposits consisting of well-sorted, layered sediments ranging in size from clay to gravel were deposited by glacial meltwater (also called the glacial outwash) and overlie the till in the valleys and coastal areas. Glacial outwash deposits form the primary aquifer.
Mixed sediments of till and stratified deposits are locally known as the Charlestown moraine that formed at the ice margin when the retreating glacier paused for a period of time. The moraine is the northern boundary of this study area (Figure 4-1d). Postglacial deposits of peat and alluvium locally overlie glacial deposits.

Groundwater pumping by private wells is the primary source of water in the study area (Baker and Panciera, 1990). This domestic water usage is mainly non-consumptive, and water is returned to the subsurface via onsite wastewater treatment systems. Typically, drinking water wells are screened in coarse-grained stratified deposits composed of sand and gravel (Masterson et al., 2006), which have higher storage and transmissivity properties than other geologic units in the study area (Friesz, 2010). The water table in the area ranges from near the surface (<1 meter (m)) to 30 m below, with the median being 6 m below the surface (RIDEM, 2015) as indicated by a USGS observation well (USGS 412214071394001 RI-CHW 18) (Cox et al., 2019). However, there is seasonality in groundwater level elevation. The water table is at the highest level in April and the lowest in October, mainly driven by the evapotranspiration effect and higher pumping to accommodate summer tourism in Rhode Island's coastal towns (Wild and Nimiroski, 2005).

Figure 4-1: Study area map. (a) the location of the study area (red polygon) in the southern Rhode Island (black outline), (b) surficial deposits and passive seismic measurement locations (Data: Rhode Island Geographic Information System, RIGIS), (c) bedrock composition (Data: RIGIS) and resistivity and seismic cross-validating sites, and (d) geologic cross-section A-A’ (reference figure 1c) showing surficial deposits and bedrock topography (Modified from Masterson et al. 2007).
3. Method

3.1 Generation of bedrock surface

Private well-completion reports (n = 1250) from 2000 to 2018 were obtained from the Rhode Island Department of Health (RIDOH). The depth to bedrock was available only from 501 well reports. The geospatial coordinates of all well locations were extracted from the Geocoder tool (https://geocoding.geo.census.gov/geocoder/) by the U.S. Census Bureau. In addition to the well reports, published bedrock topography point dataset (n = 1226) from 1948-1968 were obtained from the USGS (DiGiacomo-Cohen and Campbell, 2021) for Rhode Island. Thus, a total of 1727 borehole data sets were evaluated. All well-log elevations were transformed to the vertical datum of NGVD29 by using 10-feet LIDAR topography data (RIGIS, 2011). A regional bedrock surface map was prepared using the kriging geostatistical technique in ArcGIS10.7 (ESRI, 2019).

3.2 HVSR measurement

When there is a high impedance contrast between the sediment of an unconsolidated aquifer and the underlying compact bedrock, a sharp peak can be observed at the resonance frequency (Sorensen and Asten 2007; Gosar 2017). In general, the resonance frequency defines oscillations in geologic materials due to an external driving force, i.e., passive waves in the case of HVSR. In the HVSR method, the resonance frequency is determined by combining the two horizontal spectra and dividing by the vertical (up-down) spectrum (thus the H/V ratio). The thickness of surficial materials is empirically related to the resonance frequency and top layer’s shear wave velocity (Nakamura, 1989). This makes it possible to estimate the thickness of the sediment or the bedrock topography with a margin of error of less than 15% (Delgado et al., 2000a). However, estimating depth to bedrock from the resonance frequency in areas with heterogeneous lithology can be error-prone (Van Noten et al., 2018). For instance, the resonance frequency conversion to depth by applying a mean shear-wave velocity will sometimes over- or underestimate the bedrock topography. Therefore, an empirical power-law relationship can be used between the resonance frequency obtained from the HVSR analysis and the depth to bedrock obtained from borehole data. Note that it is important to sample the HVSR resonance frequency over a range of depth to bedrock to obtain a meaningful regression. The empirical relationships between the resonance frequency and the depth to bedrock, however, are mostly site-specific (Thabet, 2019) because shear wave velocity varies across sites, and the resonance frequency varies with depth at any site. This requires that a sufficiently large number of paired field measurements and HVSR dataset is available to establish a robust relationship of resonance frequency and thickness at a regional scale (Gosar and Lenart, 2010). Also, independent physical or geophysical information must be available to validate the resonance frequency to depth to bedrock relationship.

Passive seismic HVSR measurements were taken at 45 locations (Figure 4-1b) where the depth to bedrock was known from private well logs. Sample locations were selected based on the range of bedrock depth, the spatial distribution of the wells, and type of surficial
deposits. In addition, 11 samples were collected from the cross-validation site (Figure 4-1c) where the depth to bedrock at the measurement point was unknown but was known in the vicinity (red box in Figure 4-1c). All measurements were taken in the summer of 2019, 2020 and 2022, and early in the morning to minimize the traffic noise. The HVSR measurement duration was 20 minutes for each survey. Environmental factors such as weather, traffic, ground type, building density, structures, and monochromatic sources of vibration were recorded for each measurement location. Extreme environmental conditions, e.g., storm events, were avoided while sampling to reduce the low-frequency perturbation noise in the sample. A Micromed Tromino model TRZ Tromographs (Moho Science and Technology) was used to measure the ambient seismic waves. The instrument was coupled to the ground with a standard gravimeter plate. This self-contained, broadband (0.10–64 Hz) seismometer is explicitly designed for engineering and geologic applications (Chandler and Lively 2014). The seismometer records vibrations in the three orthogonal components (East-West, North-South, and vertical).

The HVSR data were analyzed with Grilla software (MOHO, 2020). A noise window of 20 seconds was selected automatically using an “anti-trigger” criterion based on the short time average/long time averages. In this way, the non-stationary transient signals present in the ambient noise recordings, such as those produced by local traffic, are excluded from the calculation of HVSR relationship (Haghshenas et al., 2008). After removing non-stationary transient signals, each component was transformed to the frequency domain. Then, the ratio of the horizontal to vertical amplitudes was calculated and averaged.

Figure 4-2: (a) An H/V spectra of the ambient noise at the site, (b) single component spectra for three components, and (c) temporal distribution of frequencies.

The resonance frequency was computed for all the sample windows. The seismic data with acceptable quality show a clear discrepancy between the three-dimensional components (opening between red and blue lines in Figure 4-2b around 3 Hz) and a clear peak at the corresponding frequency in spectral ratio (Figure 4-2a). Figure 4-2c displays a spectrogram showing the spectral ratio is consistent over the full 20-minute measurement. To isolate the good quality data from the samples, we applied three approaches: data visualization such
as clear and single peak in H/V spectra, consistency of resonance frequency in the entire duration of measurement, as shown in Figure 4-2, SESAME (2004) criteria (Annex I), and data-quality coding (Annex II).

A power-law (Ibs-von Seht and Wohlenberg, 1999) represents the relationship between the resonance frequency and the depth to bedrock as shown in equation (1):

\[ y = af^{-b} \]  

(1)

where \( y \) is the depth to bedrock from the surface, \( f \) is the resonance frequency, and coefficient \( a \) and exponent \( b \) are the intercept and slope, respectively.

For the simplest case of a uniform sediment layer, the depth to bedrock and the resonance frequency can be related as shown in equation (2):

\[ y = \frac{V_s}{4f} \]  

(2)

where \( V_s \) is the shear wave velocity, and \( f \) is the HVSR resonance frequency.

To further strengthen the regression model, a weighted least square regression was employed using the SESAME evaluation criteria and quality coding as shown in Annex II. This technique is better suited when there is heteroscedasticity in data so that equal weightage cannot be applied to all the data points.

### 3.3 Electrical resistivity measurement

ERI relies on injecting current in the ground using two electrodes. Additional pairs of electrodes measure the electrical potential, which varies depending on saturation, porosity, water chemistry, tortuosity, cementation, and the matrix of the lithology. The variation in potential difference can be related to the contrast between the resistivity of the saturated porous aquifer, and compact bedrock surface (Olhoeft, 1981). The resistivity of the subsurface is then estimated by inverting the apparent resistivity data from the electrode pairs. The bedrock surface's exact depth is challenging to be determined from the ERI in coastal environments where freshwater overlies saltwater.

Two ERI survey were carried out on the test site. One near the shoreline along a 275-m transect in an open area and the other in inland along 220 m transect (Figure 4-1c) using a SuperSting R8 instrument with 56 electrodes during the spring of 2021. To reduce the contact resistance with the ground to an acceptable level (below 5000 Ohm (Ω)) the ground
around the electrodes was wetted with saltwater. A Wenner-Schlumberger array was used for the data collection, which is better suited than a dipole-dipole array in resolving horizontal layers (Ward, 1990). The resistivity inversion was performed with the RES2DINV software by Loke (2013). There was a large variation in apparent resistivity values near the surface (minimum 650 and maximum 10500 Ω m in inland site) along the resistivity profile. Therefore, during the inversion we refined a model where the cell width was half of the electrode spacing. The robust inversion method was used to produce a sharp boundary between two surfaces i.e., the overburden material and the bedrock. An existing observation well near the inland resistivity transect was used to validate the resistivity data for the depth to water table using the HVSR relationship (well data not shown here).

4. Results

4.1 Bedrock mapping based on existing well log report

Figure 4-3: Depth to bedrock from land surface in southern Rhode Island: (a) USGS data based on well log and topographic surface data from 1948 to 1968 (DiGiacomo-Cohen and Campbell, 2021), and (b) merging USGS with data from more recent well-completion
reports. Rectangular box shows an area covered by a moraine and where major improvements of the bedrock map resulted from the increased data coverage.

Figure 4-3 shows the depth to bedrock in the study area. The bedrock topography ranges from surficial outcrops to 80 m below the surface. Figure 4-3a shows data from 1226 well-log reports used by USGS bedrock dataset for the period of 1948 to 1968 (DiGiacomo-Cohen and Campbell, 2021). That dataset was merged with additional 501 well-completion reports from 2000 to 2018 (Figure 4-3b). The additional data resulted in a marked improvement in the kriged bedrock surface, and particularly along an area covered by a moraine.
4.2 HVSR site-specific relationship

Out of 45 samples, 38 were qualified for use in the further investigations. The remaining 7 samples failed on both the SESAME (2004) criteria. The resonance frequency was correlated to the depth to bedrock for all 38 samples (Figure 4-4a). In the weighted

Figure 4-4: The relationship between the HVSR resonance frequency and the measured depth to bedrock for the study area (a) all samples, (b) all samples with weighted regression based on their data quality (see annex II for the quality weightage criteria), (c) glacial outwash only, and (d) till and mixed till and outwash locations.
regression, the samples with highest quality code were weighted 4 times higher than the ones with the samples participated with the lowest quality. As expected, the weighted regression improved the model from $R^2=0.91$ to $R^2=0.94$ (Figure 4-4b). The model’s performance was also evaluated based on surficial geology data. The well-sorted glacial outwash deposits exhibited a much higher coefficient of determination ($R^2=0.96$), and narrower uncertainty envelope (Figure 4-4c) relative to poorly sorted till and mixed outwash ($R^2=0.94$; Figure 4d), or the combination of the two (Figures 4-4a and 4-4b). This result suggests that resonance frequency measurements in glacial outwash areas are more uniform and deliver more reliable depth to bedrock estimates. Also, this demonstrates the process of improving the model performance, particularly with the weighted regression and by segregating the data based on surficial deposits (glacial outwash vs. till mix in our case).

4.3 Comparison of regional and international regression equations

Figure 4-5: Comparison of the relationship between the HVSR resonance frequency and the depth to bedrock among published regression equations across the globe. The dashed lines (black: glacial outwash; red: glacial till) are the relationships from this study. The detailed description of data used in this figure is provided in Annex III.

Figure 4-5 shows the relationship between the resonance frequency and the depth to bedrock in different parts of the world as compared to this study (dashed). Because of data limitations, the HVSR regression model was calibrated with data from sites with deposits up to 33 m thick. The variation in the regression models underlines that HVSR requires a site-specific regression model for better predictability. However, this study’s regression model for glacial outwash (Figure 4-4c) is comparable to other locales with similar geology, such as Johnson and Lane (2016) in Connecticut and Fairchild et al. (2013) in coastal Massachusetts.
4.4 Cross-validation

Figure 4-6: (a) Contours representing the depth to bedrock surface (black lines) derived from HVSR seismic data on top of a previously published representation of the bedrock surface shown with a color scale (DiGiacomo-Cohen and Campbell, 2021). The extent of this map is shown on figure 1c (b) Cross-plot between measured (well-completion reports) and HVSR calculated depth to bedrock for all the 29 samples. The shaded area represents the 95% confidence interval. The separate regression models (figure 4-4b, 4-4c) were employed for glacial outwash and till mix deposits to calculate the depth to bedrock.

Figure 4-6a depicts the cross-validation of the calculated depth to bedrock using the regression equation between the HVSR resonance frequency and the depth to bedrock. The background color map is interpolated from the well-log data (Figure 4-3a). The regression model for the glacial outwash material (Figure 4-4c) shows a good agreement in capturing the same patterns for shallow as well as deep bedrock topography as interfered from well-completion reports. There are discrepancies between the calculated depth using the HVSR regression equation for glacial outwash and the depth from the well reports, as seen on the southeast corner of the validation site (Figure 4-6b), which could be because of the limited number of HVSR samples for cross-validation. Note here that none of the HVSR measurements on the validation sites were taken from wells where depth to bedrock was known. Figure 4-6b shows the cross-plot between the measured and calculated depth to bedrock from all the seismic samples taken from known sites for developing the empirical relationship (Figures 4-4c and 4-4d). As the shaded envelopes (95% confidence intervals) indicate and as expected from the higher coefficient of determination (Figures 4-4c and 4-4d), there is a higher level of certainty in depth to bedrock data in the glacial outwash area than the till. In both the deposits, the uncertainty increases with the depth. Overall, the relationship for both geologic materials combined reasonably predicts the depth to bedrock with an overall 17% margin of error for all samples. The average residual declines to 13% in areas covered by outwash and increases to 23% in till deposits.
Figure 4-7: Depth to bedrock cross-validation with electrical resistivity imaging. (a) Resistivity profile from inland site and (b) resistivity profile from coastal site (for location refer to Figure 4-1c). The yellow plus symbols are the HVSR seismic measurement points along the resistivity profile. The dashed black line between the plus symbols is the calculated depth to bedrock using the HVSR relationship for glacial outwash (Figure 4c).

Figure 4-7 shows the inverted ERI profile and the estimated depth to bedrock using HVSR and regression equation for outwash (Figure 4-4c) at four locations along the profiles. The red color indicates a resistive layer close to the surface consistent with dry sediment deposits above the water table. The blue color in the ERI profile (Figure 4-7a) signifies a conductive layer, which is interpreted as freshwater and was confirmed with data from a water table observation well in a nearby location (~50 m from the profile line). The dark blue color (Figure 4-7b) is brackish water since this transect lies in near-shore environment. The calculated average depth to bedrock based on the HVSR regression (Figure 4-4c) is 25 m below land surface for an inland transect (Figure 4-7a) but was between 13 to 20 m for the coastal transect. The land surface topography was flat for both the transects. The bedrock topography is dipping towards the ocean in the near-shore transect. On the inland transect, the resistivity of the bedrock (below the conductive saturated glacial deposits) was about 500 to 1000 Ω.m. According to Ronning (2014) in a study in Norway in an inland environment, a crystalline bedrock’s resistivity is classified into three categories: resistivity >3000 Ω.m is intact rock, 3000-500 Ω.m indicates fractured rock with a presence of fluid,
and $<500 \ \Omega \cdot m$ indicates fractured rocks mixed with clay and water, in the absence of other contaminants such as saltwater from road salts and sea, and leachate from landfill sites. Here, the bedrock resistivity is above $500 \ \Omega \cdot m$, which indicates the presence of water-filled fractures. Similarly, Morse et al. (2012) defined a bedrock resistivity value of 1000 to 2000 $\Omega \cdot m$. This range is similar to crystalline bedrock reported by Adepuilimi et al. (2006). However, on the near-surface transect, the resistivity of the subsurface below the calculated depth to bedrock is lower than above. This is because of the presence of the more conductive media i.e., brackish water. The data demonstrate that the HVSR bedrock mapping technique can play an essential role in interpreting the resistivity profile and eventually aide in groundwater mapping. Moreover, as shown in Figure 4-7b, in a coastal environment, it is challenging to delineate the bedrock topography only by using ERI, whereas by combining it with the HVSR technique aids in mapping the bedrock topography.

5. Discussion

5.1 HVSR relation
A statistically significant number of acceptable measurements ($n = 38$) resulting from high-quality resonance frequency peaks were collected. The HVSR regression model for glacial outwash developed with 24 measurements provided reasonably good ($<13\%$ bias) depth to bedrock estimates ($R^2 = 0.96$). At till sites, the depth estimated by the regression model was $23\%$ off in average compared to the well-completion reports. This could be due to the unsorted and loose materials embedded in the glacial till deposits, overriding monochromatic noise sources, poor instrument coupling, or inaccurate well-completion reports (Seiderman, 2018). Moreover, the quality of well-log records (handwritten in many cases) was low in some instances and made it difficult to extract the depth to bedrock from borehole data. This could weaken the HVSR relationship. Together, these results suggest a need to establish a more reliable regression equation to relate HVSR to depth to bedrock in till. Also, the calculated shear wave velocity (Equation 2) for combined samples was $321 \ \text{m/sec}$, but the value for glacial outwash was found to be $287\pm58 \ \text{m/sec}$ and that for till mix was $380\pm110 \ \text{m/sec}$. The higher velocity in till mix deposit can be because of the presence of clay mixed with till in some locations. Overall, the regression equations are closely related to the one developed for the western Cape Cod (Fairchild et al., 2013) in similar glacial deposits.

5.2 Comparison between the geophysical techniques
The inland ERI profile is in good agreement with the HVSR calculated depth to bedrock. However, it can be difficult to estimate the depth to bedrock with the ERI technique alone, because the measurement error increases exponentially with depth (Loke et al., 2013), making the interpretation of deeper sections less reliable. Also, the resistivity value decreases if the bedrock is fractured, thereby mimicking relatively electro-conductive
layers, such as clay, and porous materials filled with conductive fluids. Also, due to the presence of saltwater in deeper layers in the coastal environment, delineating bedrock topography only with ERI is challenging. This is especially true where well lithology is not available. These complexities make the bedrock depth estimation extremely difficult with only the ERI method in deeper layers. However, if collocated measurements are made, HVSR techniques can aide in interpreting the resistivity profiles, as shown in Figures 4-7a and 4-7b.

5.3 Bedrock mapping implications for groundwater studies

Figure 4-8: Electrical conductivity (EC) of the water column in monitoring wells in (a) in a near-shore well (<100 m from the ocean) compared to (b) a well further inland (~1 km from the ocean and 500 m from a salt pond) at Charlestown, Rhode Island. The location of these monitoring wells is shown in figure 1 (c). Fresh and seawater EC values are marked with a dashed red line for comparison. Also shown is the depth to the bedrock as estimated from the HVSR based regression model (figure 4c) and obtained from well-completion reports. The EC was monitored lowering an EC tape from the wellhead all the way to the bottom and with no or minimal stirring to prevent the water column from mixing.

Delineating bedrock topography in a coastal aquifer is important for water supply system because saltwater from the ocean can potentially contaminate the groundwater inland via intrusion along bedrock fractures (Giese and Barthel, 2021). Figure 4-8a shows an abrupt change in the electrical conductivity (EC) of groundwater, a proxy measurement of saltwater, in a monitoring well just below the bedrock surface (depth to the bedrock surface is 8.8 m) in a near-shore well (<100 m from the ocean). In a well ~1 km inland and 500 m from a coastal salt pond (Figure 4-8b), the EC readings change abruptly but at a greater depth (51 m below the ground surface) and within the bedrock. These findings confirm the expected thinning of the freshwater column closer to the ocean, which increases the vulnerability of near-shore aquifers to saltwater intrusion due to over-pumping and
saltwater-upconing (Zhou et al., 2005). Also, the abrupt EC increase in the inland well coincided with a major fracture noted in the drill log (Figure 4-8b). In the absence of other salt sources, such as nearby deicing salt storage facilities, the presence of saltwater at depth at this location suggest that seawater can intrude significant distances (at least 1 km in the study area) inland along major fractures. This finding has implications for well installations in areas farther away from the coastline because drilling wells deep into the bedrock can produce saltwater contamination even when the well is not pumped, as it is the case for the domestic use well shown in Figure 4-8b.

Bedrock topography mapping in areas covered by comparatively thin veneers of glacial deposits is essential for groundwater studies for several reasons. For instance, the bedrock topography controls groundwater flow and consequently the advective flow of contaminants in the surficial aquifer. As shallow overburden deposits have limited storage capacity for infiltrating water, knowledge about their thickness can benefit stormwater management and contaminated site remediation design and operation (Shahri et al., 2020). For reference, in the study areas at least 3 feet (~1 m) separation is required between the bottom of a stormwater management facility (such as septic tank) and the bedrock (RIDEM, 2015).

6. Conclusion

A high-resolution bedrock topography map was produced using well-log information from private and public wells. The depth to bedrock in the study area ranges from the surface (=exposed outcrops) to 80 m below land surface. A new map was produced from a larger number of well-completion reports with depth to rock information and, compared to prior maps, improved the accuracy and resolution of the bedrock topography, particularly in areas farther inland that are covered by a moraine. A set of HVSR relationships for the depth to bedrock was developed using ambient seismic measurements in two types of surficial deposits (glacial outwash and till) as well as their aggregate. The HVSR relation estimated the bedrock depth with high confidence (13% bias) in areas covered by comparatively homogenous glacial outwash deposit. Where heterogeneous till deposits were present, the error increased to 23% on average, suggesting that the make-up of the unconsolidated sediments above the bedrock directly impacted the predictive quality of the HVSR model. The relationship between the surficial aquifer thickness and the HVSR measurements for glacial outwash in this study area is comparable with the studies carried out in regions with similar hydrogeology. The HVSR regression model was validated with seismic samples from within the study area. The HVSR results showed good agreement with the depth to bedrock observations. In addition, cross-validation with two ERI profiles, one in inland and the other in coastal environment, showed good agreement with the layer contrast in resistivity, indicating that the HVSR technique supplements the resistivity data interpretation in subsurface mapping. Therefore, when HVSR is integrated with electrical resistivity imaging technique, sub-surface characterization and particularly groundwater mapping is more robust than each technique on its own.
Acknowledgment

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Data availability: All the data used in this study, including the seismic samples, well lithology and resistivity profiles, will be uploaded to CUAHSI’s Hydroshare webportal. The portal link will be provided here after the manuscript is accepted.

Conflict of Interest: The authors declare no conflict of interest.

The use of firm, trade, and brand names is for descriptive purposes only and does not constitute an endorsement by the U.S. Geological Survey or the University of Rhode Island.

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Table Annex I: SESAME criteria for analyzing the data quality (SESAME WP04, 2004)

### Criteria for a reliable H/V curve

[All 3 should be fulfilled]

<table>
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<th>Condition</th>
<th>Value</th>
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<tr>
<td>$f_0 &gt; 10 / L_w$</td>
<td>3.59 &gt; 0.50</td>
</tr>
<tr>
<td>$n_c(f_0) &gt; 200$</td>
<td>4312.5 &gt; 200</td>
</tr>
<tr>
<td>$s_A(f) &lt; 2$ for $0.5f_0 &lt; f &lt; 2f_0$ if $f_0 &gt; 0.5$ Hz</td>
<td>Exceeded 0 out of 174 times</td>
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<tr>
<td>Exists $f^-$ in $[f_0/4, f_0]$</td>
<td>$A_{H/V}(f^-) &lt; A_\theta / 2$</td>
</tr>
<tr>
<td>Exists $f^+$ in $[f_0, 4f_0]$</td>
<td>$A_{H/V}(f^+) &lt; A_\theta / 2$</td>
</tr>
<tr>
<td>$A_\theta &gt; 2$</td>
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<tr>
<td>$f_{peak}[A_{H/V}(f) \pm s_A(f)] = f_0 \pm 5%$</td>
<td>[0.04208] &lt; 0.05</td>
</tr>
<tr>
<td>$s_t &lt; e(f_0)$</td>
<td>0.15122 &lt; 0.17969</td>
</tr>
<tr>
<td>$s_A(f_0) &lt; q(f_0)$</td>
<td>0.3037 &lt; 1.58</td>
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### Criteria for a clear H/V peak

[At least 5 out of 6 should be fulfilled]

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<tr>
<td>Exists $f^+$ in $[f_0, 4f_0]$</td>
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<tr>
<td>$s_t &lt; e(f_0)$</td>
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</tr>
<tr>
<td>$s_A(f_0) &lt; q(f_0)$</td>
<td>0.3037 &lt; 1.58</td>
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<td>window length</td>
</tr>
<tr>
<td>$n_w$</td>
<td>number of windows used in the analysis</td>
</tr>
<tr>
<td>$n_c = L_w \cdot n_w \cdot f_0$</td>
<td>number of significant cycles</td>
</tr>
<tr>
<td>Term</td>
<td>Description</td>
</tr>
<tr>
<td>------------</td>
<td>-----------------------------------------------------------------------------</td>
</tr>
<tr>
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</tr>
<tr>
<td>$f_{\text{peak}}$</td>
<td>peak frequency</td>
</tr>
<tr>
<td>$f_0$</td>
<td>H/V peak frequency</td>
</tr>
<tr>
<td>$s_f$</td>
<td>standard deviation of H/V peak frequency</td>
</tr>
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<td>$e(f_0)$</td>
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<tr>
<td>$A_{\text{H/V}}(f)$</td>
<td>H/V curve amplitude at frequency $f$</td>
</tr>
<tr>
<td>$f^-$</td>
<td>frequency between $f_0/4$ and $f_0$ for which $A_{\text{H/V}}(f^-) &lt; A_0/2$</td>
</tr>
<tr>
<td>$f^+$</td>
<td>frequency between $f_0$ and $4f_0$ for which $A_{\text{H/V}}(f^+) &lt; A_0/2$</td>
</tr>
<tr>
<td>$s_A(f)$</td>
<td>standard deviation of $A_{\text{H/V}}(f)$, $s_A(f)$ is the factor by which the mean $A_{\text{H/V}}(f)$ curve should be multiplied or divided</td>
</tr>
<tr>
<td>$s_{\text{logH/V}}(f)$</td>
<td>standard deviation of log $A_{\text{H/V}}(f)$ curve</td>
</tr>
<tr>
<td>$q(f_0)$</td>
<td>threshold value for the stability condition $s_A(f) &lt; q(f_0)$</td>
</tr>
</tbody>
</table>

**Annex II: Data quality coding key for this research paper**

- **Weightage 4:** If the data pass all the Sesame criteria and no other issues in coupling
- **Weightage 3:** If the data pass all the Sesame criteria and there are issues in coupling
- **Weightage 2:** If the data fails to pass only 1 criterion and minor issues in coupling
- **Weightage 1:** If the data fails to pass more than 1 criteria and/or has other significant issues in coupling,
  - failed in only 1 criterion and has other major problem in coupling
- **Weightage 0:** If the data fails to pass both the criteria
Table Annex III: List of HVSR regressions from across the globe (supplement to Figure 4-5), where \(a\) is the intercept, \(b\) is the slope of the regression equation and \(R^2\) is the coefficient of determination.

<table>
<thead>
<tr>
<th>Location</th>
<th>Study</th>
<th>Geologic materials</th>
<th>(a)</th>
<th>(b)</th>
<th>(R^2)</th>
<th>Sampling points</th>
<th>HVSR Frequency range (Hz)</th>
<th>Depth range (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower Rhine, Embayment, Germany</td>
<td>Ibs-von Seth and Wohlenberg (1999)</td>
<td>Sedimentary deposit</td>
<td>96.0</td>
<td>1.39</td>
<td>0.98</td>
<td>102</td>
<td>0.10,5.0</td>
<td>15–1600</td>
</tr>
<tr>
<td>Segura River Valley</td>
<td>Delgado et al. (2000a)</td>
<td>Sedimentary deposit, carbonate rock</td>
<td>55.1</td>
<td>1.26</td>
<td>0.97</td>
<td>180</td>
<td>1.16,8.3</td>
<td>4.1–43.5</td>
</tr>
<tr>
<td>Southeastern Spain</td>
<td>Delgado et al. (2000b)</td>
<td>Sedimentary deposit, carbonate rock</td>
<td>55.6</td>
<td>1.27</td>
<td>0.98</td>
<td>33</td>
<td>1.10,8.3</td>
<td>4.1–44.7</td>
</tr>
<tr>
<td>Cologne area, Germany</td>
<td>Parolai et al. (2002)</td>
<td>Sedimentary deposits</td>
<td>108.00</td>
<td>1.55</td>
<td>N/A</td>
<td>337</td>
<td>0.25,20.00</td>
<td>2–500</td>
</tr>
<tr>
<td>Lower Rhine, Embayment, Germany</td>
<td>Hinzen et al. (2004)</td>
<td>Sedimentary deposit, Paleozoic rock</td>
<td>137.00</td>
<td>1.19</td>
<td>0.96</td>
<td>152</td>
<td>0.20,2.5</td>
<td>10–1200</td>
</tr>
<tr>
<td>Southern coast of Istanbul, Turkey</td>
<td>Birgoren et al. (2009)</td>
<td>Paleozoic rock</td>
<td>150.99</td>
<td>1.15</td>
<td>0.99</td>
<td>15</td>
<td>0.40,1.5</td>
<td>20–449</td>
</tr>
<tr>
<td>Izmit Bay area, Turkey</td>
<td>Ozalaybay et al. (2011)</td>
<td>Sedimentary deposits</td>
<td>141.00</td>
<td>1.27</td>
<td>0.91</td>
<td>239</td>
<td>0.25,3.8</td>
<td>20–1100</td>
</tr>
<tr>
<td>Western Sydney, Australia</td>
<td>Harutoonianian et al. (2013)</td>
<td>Shale and sandstone beds</td>
<td>73.0</td>
<td>1.17</td>
<td>0.94</td>
<td>15</td>
<td>4.20,27.0</td>
<td>1.2–13.3</td>
</tr>
<tr>
<td>Cape Cod, Massachusetts, USA</td>
<td>Fairchild et al. (2013)</td>
<td>Glacial outwash</td>
<td>90.5</td>
<td>1.00</td>
<td>1.00</td>
<td>164</td>
<td>0.73,2.4</td>
<td>118–460</td>
</tr>
<tr>
<td>L’Aquila, central Italy</td>
<td>Del Monaco et al. (2013)</td>
<td>Lacustrine/fluvial sediment and carbonate rock</td>
<td>53.4</td>
<td>1.01</td>
<td>0.41</td>
<td>25</td>
<td>4.00,10.0</td>
<td>3–20</td>
</tr>
<tr>
<td>Eskisehir Quaternary</td>
<td>Tu’n et al. (2016)</td>
<td>Unknown</td>
<td>136.00</td>
<td>1.36</td>
<td>0.99</td>
<td>30</td>
<td>0.30,15.00</td>
<td>0–500</td>
</tr>
<tr>
<td>Location</td>
<td>Author(s)</td>
<td>Deposits Description</td>
<td>Vwater</td>
<td>Vmin</td>
<td>Vmax</td>
<td>Vmin</td>
<td>Mv</td>
<td>Width (m)</td>
</tr>
<tr>
<td>--------------------------------</td>
<td>------------------------------------</td>
<td>------------------------------------------------------------</td>
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<td>------------</td>
</tr>
<tr>
<td>Japan</td>
<td>Thabet (2019)</td>
<td>Sedimentary bedrock</td>
<td>1.19</td>
<td>0.94</td>
<td>224</td>
<td>0.01</td>
<td>2.5</td>
<td>2–1500</td>
</tr>
<tr>
<td>Tylerville, CT (USA)</td>
<td>Johnson and Lane (2016)</td>
<td>Unknown</td>
<td>1.18</td>
<td>0.93</td>
<td>176</td>
<td>1.9</td>
<td>18</td>
<td>3.8–48</td>
</tr>
<tr>
<td>Ljubljana Moor, Slovenia</td>
<td>Gosar and Lenart (2010)</td>
<td>Lacustrine and fluvial deposits</td>
<td>1.25</td>
<td>0.57</td>
<td>53</td>
<td>0.7</td>
<td>10</td>
<td>4–150</td>
</tr>
<tr>
<td>Bam city, Iran</td>
<td>Motamed et al. (2007)</td>
<td>Alluvial deposits</td>
<td>1.97</td>
<td>49</td>
<td>1.00</td>
<td>7.0</td>
<td>2–150</td>
<td></td>
</tr>
<tr>
<td>South coast, R.I. (USA)</td>
<td>This Study (Glacial outwash)</td>
<td>Glacial outwash and crystalline bedrock</td>
<td>0.95</td>
<td>0.96</td>
<td>24</td>
<td>2.1</td>
<td>16</td>
<td>3.7–33.5</td>
</tr>
<tr>
<td>South coast, R.I. (USA)</td>
<td>This Study (Glacial till mix)</td>
<td>Glacial till and crystalline bedrock</td>
<td>1.01</td>
<td>0.94</td>
<td>14</td>
<td>2.25</td>
<td>18</td>
<td>3.4–39.6</td>
</tr>
</tbody>
</table>

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Chapter 5: The squeezing of freshwater lenses in barrier island: A combined geophysical and numerical analysis (Manuscript 4)

_In Prep. for Journal of Hydrology_

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Abstract

Coastal aquifers, which supplies freshwater to over one billion people globally, are frequently impacted by saltwater intrusion, particularly in the case of island freshwater lenses. Despite the proliferation of studies exploring the impacts of sea-level rise, storm surges, and over-pumping, the impact of droughts on coastal aquifers remains under-investigated. This lack of attention is particularly concerning given the heightened vulnerability of barrier island aquifers, which rely solely on aerial recharge for freshwater, compared to their inland counterparts. An in-depth understanding of recharge and salinization processes is imperative for sustainable water resource management, particularly in the face of potential climate change impacts. This study presents a new evaluation approach for the fate of a freshwater lens during drought conditions, incorporating a combination of in-situ observations, geophysical measurements, and numerical modeling, which has not been previously reported in the literature. The study examines the response of a shallow unconfined aquifer located on a barrier island to the 2020 drought in the Northeastern United States by developing a density-driven flow model utilizing time-lapse electrical resistivity imaging and in-situ groundwater head and salinity data collected from monitoring wells. The results indicate that the freshwater lens is squeezed by 11% due to reduced recharge during the drought in the summer and fall seasons and returns to its normal position over the spring season. These findings highlight the
vulnerability of shallow unconfined coastal aquifers to droughts, emphasizing the importance of in-depth studies in this area.

**Keywords:** Freshwater Lens; Barrier Island; Drought; Electrical Resistivity; Numerical Model

**Highlights**

1. Integrating observations, geophysics, and modeling to assess freshwater lens (FWL).
2. FWL devastated by drought, recovered in next season with higher recharge rate.
3. Barrier islands' freshwater is sensitive to climate-induced rainfall variability.

**1. Introduction**

Barrier islands, narrow strips of sandy unconsolidated sediment parallel to coastlines, can provide water to coastal populations and sustain crucial groundwater-dependent habitats for migratory birds and other critical species. Spanning about 7% of the coastlines globally, barrier islands protect lagoons and bays further inland. Globally, coastline consists of more than 2000 barrier islands, with ~75% located in the Northern Hemisphere (Stutz and Pilkey, 2011). The United States stands for the highest fraction, with 24% of its shoreline being barrier islands (Stutz and Pilkey, 2011). On the U.S. Atlantic and Gulf of Mexico coasts, the combined length of barrier islands is 3700 km, with a total area of 6800 km². The average width of these islands is 1.8 km (Zhang and Leatherman, 2011). Globally, many
Barrier islands are densely populated and the population density is projected to increase in the future (Neumann et al., 2015). According to 2000 U.S. Census, 1.4 million people live on barrier islands in the U.S. and half of it in Florida alone. The population densities of barrier islands are three times those of the US coastal states on average, and the population increased by 14% from 1990 to 2000 (Zhang and Leatherman, 2011). Fresh groundwater on barrier islands is arguably the most important resource as the island communities disproportionality rely on groundwater for water supply given the general absence of surface water resources and limited hydrologic adaptive capacity (Werner et al., 2017). However, the freshwater availability in such fragile areas is stressed by natural as well as anthropogenic factors (Holding et al., 2016; Schneider and Kruse, 2006).

Barrier Island hydrology is unique and is driven by complex interactions between the ocean, typically unconsolidated geologic deposits that make up most barrier islands and precipitation. The islands are surrounded by sea on one side and ponds, lagoons, bays, or wetlands on the other side, except where the islands are connected to the mainland (Figure 5-1). In the absence of rivers or streams, aerial recharge by precipitation is the only freshwater source (Anderson et al., 2000). Rain quickly infiltrates the predominantly sandy deposits and reaches the water table. From there, the groundwater typically drains into both the ocean and inland water bodies. This process is referred to as submarine groundwater discharge (SGD). Because of the contrast in density, the freshwater buoys on top of the saltwater, depresses the underlying saltwater downwards and forms an elongated convex lens-shaped freshwater lens (FWL) (Briggs et al., 2020; Post et al., 2019). Given the low elevation of barrier islands—typically not exceeding 5 m—and high infiltration capability of unconsolidated sand deposits (Holt et al., 2019), these deposits allow for high infiltration
The size of the freshwater lens on the island is determined by the width of the island and the hydraulic gradient between the water table and seawater, which is mainly influenced by the island's elevation and vegetation composition. (Ault, 2016a; Schneider and Kruse, 2006).

The hydrostatic equilibrium between saltwater and freshwater on barrier islands is delicate and very sensitive to changes in sea level, storm surge, recharge, and pumping (Werner et al., 2017). These islands are prone to storm surge over-wash and related saltwater infiltration; however, infiltrating precipitation water displaces the saltwater over time. Therefore, the recovery time of the freshwater lens after a storm event is mainly a function of recharge, which is governed by aquifer hydraulic properties, such as hydraulic conductivity and thickness of the vadose zone, and the heterogeneities therein (Holding and Allen, 2015; Holding et al., 2016). Water loss by transpiration is minimal because the vegetation on the barrier island is often limited to dune grasses and shrubs (Masterson et al., 2014). Groundwater residence time on barrier islands varies with the internal structure, island geometry, climate, and geological factors. It can be range from 37 to 150 years (Hofmann et al., 2020).
Figure 5-1: Conceptual model of freshwater lens in a barrier island, and its response to variable recharge.

Freshwater lenses on barrier islands are dynamic and highly susceptible to both natural and anthropogenic stresses; therefore, barrier island aquifers are continuously threatened by salinization (Ault, 2016b; Bailey et al., 2016). Rising sea levels, increasing frequency and intensity of storm surges and drought events, and over-pumping are the primary threats (Anderson Jr. and Lauer, 2008; Holding et al., 2016; Ketabchi et al., 2016; Werner et al., 2017). Sea level rise will exacerbate surge flooding, coastal erosion, and saltwater intrusion (Zhang and Leatherman, 2011). The freshwater lens changes substantially in response to sea-level rise and storm surge events, as the elevation of many barrier islands is within a few meters of the current sea level (Masterson et al., 2014). Also, the effects of climate change will be the most pronounced on barrier islands (IPCC, 2021; White and Falkland, 2010). This is because climate change alters the precipitation and evapotranspiration regime and thereby the groundwater recharge (Taylor et al., 2013; Thomas and Famiglietti,
which is the only source of freshwater on the islands. Due to hydrological droughts, the groundwater recharge on small islands is expected to decline by as high as 58% by 2060s (Holding et al., 2016). For example, the freshwater lens in Dauphin Island, a barrier island in Alabama, might not be able to sustain any further population growth due to the impact of climate change (Chang et al., 2016). Similarly, the freshwater lens on Bonriki Island at Kiribati in the central Pacific Ocean is projected to decline as droughts become more frequent and severe (Post et al., 2019). Similar conditions threaten some islands in the Caribbean, which face a severe reduction in precipitation in the future (Ault, 2016b).

During drought events, reduced recharge may cause shrinkage of the freshwater lens on barrier islands. At the same time, the lack of aerial recharge might be exacerbated by over-pumping from the freshwater lens to meet increased demand. The degree of over-pumping and anthropogenic stresses on the aquifer system might increase further because these islands are increasingly attractive for recreational and residential developments (Zhang and Leatherman, 2011). Further, it is well known that water demand for the tourism sector exceeds local demands in most coastal and tourist islands, especially in the summer season (Mansour et al., 2017; Nutbrown, 1976). In sum, freshwater lenses in barrier islands are under compounding pressure from climate change and human stressors. Therefore, it is essential to understand the dynamics of the freshwater lens under varying stress and scenarios to ensure sustainable management of the aquifer systems.
A great number of studies characterise and simulate the freshwater lens on island aquifers, including the impacts of sea-level rise (Masterson et al., 2014; Masterson and Garabedian, 2007) and storm surge (Adyasari et al., 2021; Housego et al., 2021). For instance, Fetter (1972) developed an analytical model using simple geometric boundaries and homogenous hydraulic conductivity to simulate the maximum depth of freshwater. Similarly, Tang et al. (2021) derived an analytical solution to simulate variable recharge scenarios for both steady-state and transient conditions. Combining variations in sea level, beach erosion, and variable recharge caused by climate change, Chesnaux et al. (2021) developed an analytical solution for estimating the combined effect of these stresses on the freshwater lens on an island. Similarly, Bedekar et al. (2019) investigated the response of a freshwater lens on oceanic island aquifer in a laboratory scale sand tank model and simulated by using a numerical code. These analytical solutions are based on the sharp-interface approximation without considering the mixing zone; therefore, they are less suitable than numerical models under future management and climate scenarios (Llopis-Albert and Pulido-Velazquez, 2014), and should not be used to calculate the volume of freshwater availability (Tang et al., 2021).

Significant progress has been made in numerical modeling of the freshwater lens. For instance, Underwood et al. (1992) developed a numerical model to investigate the hydrological controls of a freshwater lens, including the effects of drought, on an island and concluded that recharge is the primary control of the thickness of the freshwater lens. Similarly, Bailey et al. (2013) studied the freshwater lens response to drought on islands in Micronesia, using a numerical model, and reported that only six out of 105 islands will
sustain freshwater during an intense drought. There is a general agreement that the freshwater lens responds to reduced recharge. Urish et al. (1989) used field observations from a Rhode Island barrier island to propose asymmetrical freshwater lens formation due to variable salt concentrations from ocean and estuary, influenced by terrestrial groundwater discharge.

The occurrence of saltwater intrusion is a widely recognized phenomenon, yet the number of case studies remains limited, particularly in barrier islands environments. The determination of the underlying causes of saltwater intrusion is crucial for the responsible utilization and management of at-risk groundwater resources. As explicitly noted by Werner et al. (2013), the integration of diverse observational data in saltwater intrusion modeling is currently limited. A comprehensive understanding of the impact of droughts on freshwater lenses in barrier islands requires a multi-disciplinary approach that integrates both geophysical and numerical modeling.

We hypothesize that combining time-lapse resistivity imaging, numerical simulation, and in-situ measurements of groundwater head and salt concentration can enhance the understanding of variable recharge on the freshwater lens dynamics on a barrier island. Previous studies have employed resistivity imaging (Baharuddin et al., 2013; Oberle et al., 2017) or numerical modeling (Alsumaiei and Bailey, 2018; Post et al., 2019) in isolation to investigate drought impacts. Pavlovski et a. (2022) demonstrated calibration of saltwater intrusion model with geophysical data, but ignored the strength of in-situ observation wells
in constraining the geophysical inversion. The objectives of this study are to: (1) quantify the rate of FWL depletion under reduced recharge on barrier islands, (2) explore recovery time after depletion, and (3) demonstrate the effectiveness of in-situ monitoring, electrical resistivity imaging, and numerical modeling in investigating hydrodynamics in barrier islands. To the authors best knowledge, this is the first study to integrate field observation, time-lapse geophysical measurements and numerical model for investigating the freshwater lens response to variable recharge.

This study employs a barrier island in Rhode Island as a test site that serves a representative model for sandy and highly permeable barrier islands globally, particularly in regard to the Holocene unconsolidated deposit-above-bedrock geology found along the Atlantic coast of the USA (Masterson et al., 2006). The elevation of the barrier island, being less than five meters above sea level, mirrors the low elevation of many similar islands worldwide. The Northeastern region, where the test site is located, experiences a higher rate of sea level rise compared to the global average, intensified by glacial isostatic rebound (Goddard et al., 2015). Future projections from climate models concur that the region is expected to experience an overall increase in precipitation, but with a projected decrease in summer precipitation (Fan et al., 2015; Rawlins et al., 2012), potentially leading to seasonal drought episodes. The occurrence of a significant drought event in 2020 in the Northeastern region provided the opportunity to test the hypothesis using ground-truth data, providing enhanced understanding and insights into the impact of droughts on barrier island aquifers. Furthermore, geological lithology and groundwater monitoring data from the test site are
available from previous research, which enables to compare the freshwater lens geometry and size.

2. Study site description

The study site, East Beach, is located on the southern coast of Rhode Island, Northeast United States (Fig. 5-2a and 5-2b). The Atlantic Ocean bounds the site to the south and a shallow salt pond (Ninigret pond) on the north side. The island extends 4.5 km and a 30 m wide breachway further east connects the salt pond with the ocean. The site's topography is characterized by sandy dunes that slope gently towards the pond and the ocean (Figure 5-2c). The highest elevation of the dune is 4.5 m from the NAVD'88 datum. Unconsolidated glacial outwash deposits and approximately 2.5 meters thick lagoon sediments (Masterson et al., 2006; Urish, 1982) sit on top of granite bedrock (confirmed by drilling). The glacial outwash deposits consist of fine to medium-grained sand. The depth to bedrock (crystalline granite) is ca. 30 m from the land surface. The annual precipitation is ca. 1200 mm in the study area, with yearly evapotranspiration being ca. 500 mm between April till October. Recharge is comparatively high, ca. 700 mm per year (Satora et al., 1987; Urish, 1982), because of increased permeability of the sandy deposits and minimal vegetation cover. The site is affected by semi-diurnal tides with an average amplitude of 0.88 m in the ocean and 0.12 m in the salt pond. The depth to the water table from the land surface in the central part of the barrier island is ca.1 m (Figure5-2c), with slightly elevated water level at the dune that divides the groundwater flow.
Figure 5-2: (a) Test site is located on the southern coast of Rhode Island, USA. (b) Close-up view of the East Beach barrier island separates a salt pond and the Atlantic Ocean. The red arrow (A-A’) indicates the electrical resistivity transect. (c) The cross-section view of the barrier island (transect A’A’ in figure b) shows the location of the observation wells (indicated by A, B, and C), island topography, lithology, and the annual average water table on the island. The dark-gray surface at the bottom is the calculated bedrock topography (Panthi et al., 2022). Vertical exaggeration is 2.5 times. Distance of well A from the ocean (high tide mark) is 50 m while well C is 10 m far from the pond. Well B is 115 m from the ocean and 85 m from the pond.
3. Data collection

3.1 Instrumentation and sampling
The test site is located on protected state land with no wells closer than 200 m. We established three piezometers on a line perpendicular to the long axis of the barrier island (Figure 5-2c). The two-inch galvanized iron wells were fully screened. Each well was outfitted with an always-submerged water level, conductivity, and temperature (LCT) sensor (*Solinst Canada*). The elevation of the observation wells and the land surface at multiple points were measured using an RTK GPS and converted to NAVD-88 vertical datum. Because of topographical variation, not all the sensors were at the same depth from the surface, but all were at the same absolute elevation i.e., 0.5 m below NAVD-88. In Well A (nearest the ocean side), the sensor was at 3.75 m below the surface, in Well B and C, it was at 2.39 m and 1.44 m below the land surface, respectively. None of the loggers ever felt dry. The sampling frequency was 6 minutes, matching the temporal resolution of tide gauge nearest the research site (Newport, RI; NOAA ID: 8452660). The logger data were corrected for the atmospheric pressure fluctuations using a barometric pressure transducer installed near the site (<5 km of aerial distance). Every three months, the logger data was downloaded, and after re-calibration for conductivity, the loggers were reinstalled at the same depth. To verify the data collected through the loggers, the water level and the vertical salinity profiles were taken at each of the data download cycles. To obtain the freshwater-equivalent groundwater head from the water level data measured in the wells, we back-calculated the water density using the electrical conductivity measured in the middle of the well screen. The process of converting water level to freshwater equivalent head is described in Post et al. (2007), Langevin et al. (2008) and Housego et al. (2021).
In addition to this transect, we drilled a deeper well (16 m below ground surface) ca. 500 m west of the test site but approximately at the same distance from the shoreline as the center of the test transect. The vertical electrical conductivity profile and water temperature at every half meter of depth were sampled once every two weeks for two full years (n=52) since December 2020. Note, not clay or peat deposits were encountered during drilling.

3.2 Resistivity imaging

The surface-based electrical resistivity imaging (ERI) technique relies on an electrical current injected into the earth's surface using a pair of electrodes. Another set of electrodes measure the electrical potential between the electrodes. The electrical potential drop is the function of geological materials. This method is well suited for saltwater mapping as the technique is sensitive to the change in electrical conductivity in the subsurface fluid. ERI is a widely used method for saltwater mapping of coastal aquifers (Goebel et al., 2017; Hermans and Paepen, 2020; Palacios et al., 2020) and island aquifers (Kiflai et al., 2020; Nakada et al., 2012; Oberle et al., 2017).

We generated ERI profiles along the A-A’ transect in 2019 Summer (medium recharge), 2020 Fall (low recharge), 2021 springs (high recharge). Note here that a severe drought occurred during the 2020 summer and fall seasons with the monthly precipitation reduced by as much as 60% relative to the long-term average for the study area. The ERT profiles were taken using an AGI Superstring R8 (Advanced Geosciences Inc.) 8-channel imaging system with 56 stainless steel electrodes with a continuous electrode spacing of 3.25 m. We marked the electrode location on the surface during our first survey in 2019 and
replicated the same path during subsequent surveys. The Wenner-Schlumberger's geometrical configuration was chosen because it better resolves the horizontal scale (Ward, 1990). To maintain the contact resistance between the electrode and surficial deposits below the recommended threshold (5000 Ω), we buried the electrodes in clay slurry (mix of cat litter and saltwater). The observed resistivity profiles were inverted using a forward modeling python package SimPEG (Cockett et al., 2015; Heagy et al., 2017) and the Gauss-Newton optimization approach with 5% of standard deviation. Prior to the forward modeling, we filtered and excluded data artifact with an apparent resistivity of more than 5 Ω.m below 10 m from the land surface, as there was no evidence of resistive layer below that depth. We applied the same data filtering and optimization approach to all time-series resistivity profiles.

A petrophysical model is required to segregate the contribution of sediment aquifers and pore water to the bulk resistivity signals in ERI profiles. Archie’s law (Archie, 1942), which applies only to the clay-free porous media, such as in the current study, explains the relationship between intrinsic parameters that represent the sediment in an aquifer, and the bulk and fluid resistivity as given in Eqn. 1:

\[ R_b = R_f \alpha n^{-q} S^{-k} \] (1)

where, \( R_b \) is the bulk resistivity from the field survey, \( R_f \) is the porewater (fluid) resistivity, \( S \) is the degree of fluid saturation, \( k \) is the saturation factor, \( n \) is porosity of the total matrix, \( q \) is the coefficient of cementation factor of the sediment and \( \alpha \) is the coefficient of
effective electrical tortuosity. All these parameters are dimensionless. For the porosity, we collected aquifer samples from the test site from three different depths below the water table and analyzed them in the laboratory. The average porosity was 0.35. Moreover, both $\alpha$ and $q$ were assumed to be 1, which are accepted parameters for unconsolidated sand and gravel deposits (Cardenas et al., 2015; Comte and Banton, 2007; Herckenrath et al., 2012). These parameters represent the high hydraulic conductivity in an unconsolidated aquifer. We considered the water-saturated deposits only i.e., $S = 1$. Furthermore, we assumed water is the only conducting media occupying the pore space. The petrophysical homogeneity of the test site permitted to directly correlate the bulk resistivity to the fluid resistivity and then inversely correlate it to the fluid conductivity. For the comparison of resistivity profiles with the numerical model and the field water samples from the observation wells, the fluid resistivity values were transformed to fluid conductivity [S/m]. The electrical conductivity (at 25°C) was then converted to total dissolved solids (TDS) [g/L] using a linear regression equation developed in our laboratory assuming NaCl is the only solute in the media.

3.4. Numerical modeling

We developed a transient, vertical density-dependent two-dimensional groundwater flow and dispersive solute transport model for the ERI transect shown in Figure 5-2, using the numerical code SEAWAT (Langevin et al., 2008). SEAWAT combines MODFLOW (Langevin et al., 2022) and MT3DMS (Bedekar et al., 2016) to calculate the density-dependent flow of solute and water. The numerical code has been used previously for barrier island hydrological investigations, such as for geomorphological and storm tide
analysis in Germany (Holt et al., 2019), for climate change and urbanization impact assessment in Alabama (Chang et al., 2016) and in Florida for assessing the impacts of sea-level rise and precipitation change (Xiao et al., 2016).

The model domains consisted of a two-dimensional rectangular model grid with 200 columns and 40 layers and an irregular grid spacing along the z-axis. The dimension of each cell was 1 m laterally and between 0.5 and 1 m vertically. To capture the saltwater interface and mixing zone, the vertical extent of each of the grids was 0.5 m for the first 10 meters of depth from the water table, while it was 1 m for the remaining layers. The model was developed only for the saturated zone, with the elevation of the top layer of the model at 1 m above NAVD'88. Hydrologic parameters used in this model are listed in Table 5-1. We considered the bedrock as a no-flow and no solute flux boundary, as in studies by Liu and Tokunaga (2019) and Eissa et al. (2013). The bedrock was at 30 m below NAVD’88 based on a passive seismic sampling data by Panthi et al. (2022) and observations by Urish (1982). Constant head and constant concentration boundaries were applied to both lateral ends of the model domain. The head towards the pond boundary was 0.5 m above NAVD-88 while the ‘effective head’ was 1 m towards the ocean side to account for the wave and tide actions. The TDS concentration of seawater and pond water averaged 31.5 g/L and 26.6 g/L, respectively, with insignificant seasonal variations. The initial TDS concentration of 31.5 g/L was applied to the entire model domain. Specified recharge was applied on the top layer and it was the only time-varying boundary condition. The TDS of freshwater, i.e., precipitation, was assumed to be 0 g/L. The average TDS value between the seawater and recharge water (15.75 g/L) was chosen to delineate the saltwater-freshwater interface, the
bottom of the freshwater lens, in the model as well as in the inverted conductivity profiles (Masterson et al., 2014; Pavlovskii et al., 2022). The constant head values were tide-averaged based on tide gauges in Ninigret pond and Newport. The hydraulic conductivity of the overburden glacial outwash was 40 m/day and 2.5 m/day for the buried lagoon sediment (Figure 5-2), which is a continuous layer at ~5.5 to 8 m from the top layer (Urish, 1982). The recharge rate was calculated using Master Recession Curve method. That method accounts for water table fluctuation but at our test site, water table fluctuation is not only function of recharge but also of tides. To eliminate the tidal influence, we used data from a surrogate well located farther inland (~ 5 km), having a similar geology (glacial outwash). To simulate the negative recharge during drought, we used a modified Thornthwaite method (Pereira and Pruitt, 2004; Thornthwaite, 1948) and weather station (Kingston station, ~20 km from the test site) data to implement a reference evapotranspiration. The resulting evapotranspiration was 3 m from the top layer of the model.

We simulated the transient model for 200 years (spin up period) from the initial condition until the model reached a quasi-steady state with respect to the simulated change in salt concentration with time. Actual steady-state condition is almost impossible to achieve in hydrodynamically active coastal aquifers (Holt et al., 2019). Next, we calibrated the model as described below and implemented various recharge scenarios to analyze the response of the freshwater lens. We did not implement groundwater abstraction from the model because there are no wells near the test site.
Table 5-1: Hydrogeologic, dispersion and diffusion parameters used in numerical modeling

<table>
<thead>
<tr>
<th>Hydrogeologic parameters</th>
<th>Unit</th>
<th>Parameter value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Porosity, n</td>
<td>(-)</td>
<td>0.35</td>
</tr>
<tr>
<td>Effective porosity, nₑ</td>
<td>(-)</td>
<td>0.29 (field calculated, Figure 5-7)</td>
</tr>
<tr>
<td>Specific yield, Sᵧ</td>
<td>(-)</td>
<td>0.25 (measured in the lab)</td>
</tr>
<tr>
<td>Specific storage, Sₛ</td>
<td>(-)</td>
<td>1E-05</td>
</tr>
<tr>
<td>Hydraulic conductivity, Kₓ and Kᵧ</td>
<td>m/d</td>
<td>Initial 1 to 40 (Urish, 1982)</td>
</tr>
<tr>
<td>Anisotropy, K_z:Kₓ</td>
<td></td>
<td>Calibrated 1 to 15</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1:10</td>
</tr>
<tr>
<td>Density of sea water (Ocean side)</td>
<td>kg/m³</td>
<td>1025</td>
</tr>
<tr>
<td>Density of pond water (Pond side)</td>
<td>kg/m³</td>
<td>1020</td>
</tr>
<tr>
<td>Salt concentration (sea water)</td>
<td>g/L</td>
<td>31.5</td>
</tr>
<tr>
<td>Salt concentration (pond water)</td>
<td>g/L</td>
<td>26.6</td>
</tr>
<tr>
<td>Molecular diffusion coefficient for NaCl, MDCOEFF</td>
<td>m²/sec</td>
<td>0.00308</td>
</tr>
<tr>
<td>Change in density over salinity</td>
<td>(-)</td>
<td>0.71</td>
</tr>
<tr>
<td>Longitudinal dispersivity, $\alpha_L$</td>
<td>m</td>
<td>5 m (Schulze-Makuch, 2005)</td>
</tr>
<tr>
<td>-------------------------------------</td>
<td>---</td>
<td>-----------------------------</td>
</tr>
</tbody>
</table>

We used the explicit third order, total variation diminishing (TVD) scheme in SEAWAT to simulate the solute transport. TVD conserves the mass and minimizes the numerical dispersion (Langevin and Zygnerski, 2013).

The depth to the freshwater interface interpreted in resistivity profile was used to constrain the saltwater intrusion model. We considered the resistivity measurements from before the drought (data from 2019 summer) in calibrating the steady-state model. While calibrating, we manually optimized the parameter ($K_x$) until it matches the depth to the calculated interface at the center of the model transect. Note here that the highest certainty in the resistivity data is at the center because of the largest number of measurements (Loke, 2013). The optimized value of $K_x$ was 15 m/day, which is close to the value reported previously, i.e. 17 m/day (Urishe, 1982). With the Ghyben-Herzberg principle (Vacher, 1988), the interface depth is highly sensitive to the change in head (roughly 40 times); therefore, calibrating saltwater intrusion model with the interface minimizes the error more effectively than calibrating the model with the head (Pavlovskii et al., 2022). The detailed modeling and data integration framework is given in Figure 5-3.
Figure 5-3: Integrated modeling framework to simulate freshwater lens and its response to variable recharge in a barrier island.
5. Results

5.1 Groundwater monitoring

Figure 5-4: Precipitation data and field observations on the deep observation well near the shoreline. (a) Monthly precipitation versus groundwater level (vertical datum: NAVD'88), (b) total dissolved solids (TDS) in groundwater. For reference, the average TDS in the seawater was 31.5 g/L and the US EPA standard level for drinking water is 0.25 g/L.

Figure 5-4 shows the relation of total dissolved solids, a proxy measure for seawater concentration in groundwater, in a near-shore observation well. There is a general seasonality pattern in TDS in the groundwater, particularly in the unconsolidated aquifer above the bedrock aquifer. The groundwater is more saline in the summer and fall seasons and less saline in winter and spring. However, individual storm events can have a
measurable, but temporary impact, such as in September 2021, when a high precipitation (168 mm in September, equivalent to 20% of the annual precipitation) event freshened the entire water column relative to the previous measurement, including the bedrock aquifer. Local precipitation flushes out the salt from the aquifer towards the ocean due to the high hydraulic gradient between the water table and the sea level. Furthermore, there is a general salinization pattern above the bedrock surface in response to the regional drought in the summer of 2022. Also, a vertical gradient in salt concentration was observed during the entire observation period. At the water table surface, the lowest salt concentration (0.124 g/L) was recorded in April 2021, and the highest (5.4 g/L) was in November 2021. Whereas, at the bottom of the well, the highest concentration (19.9 g/L) was recorded in December 2021, and the lowest concentration (7.35 g/L) was in September 2021 following the heavy precipitation in the month.
Figure 5-5: Groundwater monitoring along barrier island transect A-A’. (a) daily mean groundwater level in the three wells, (b) corresponding daily mean total dissolved solids of groundwater measured at -0.5 m from NAVD-88, and (C) daily mean sea level from NAVD-88 and daily precipitation from a meteorological station at Westerly (~ 10 km from the test site). The shaded columns represent specific hydrological events. Blue: highest precipitation event (September 2021), Green: highest mean sea level recorded, Red: highest high sea level during a storm event (January 2022 Nor’easter), Gray: drought event in 2022, Yellow: storm surge event in September 2022. The data gaps in (a) and (b) are periods of sensor malfunction.

Figure 5-5 correlates groundwater level and the water salinity fluctuations to hydrological controls, such as storm events, long-term tide cycles and droughts. The groundwater level clearly responds to the temporary rise in sea level. In Well A, for instance, the groundwater level rose 1 meter during the highest mean sea level in October 2021. During storm event, vertical saltwater intrusion increases the water table coupled with higher recharge by heavy
precipitation. The groundwater level appears responding to major precipitation events. For example, a rain event on September 1, 2021, yielded 63 mm precipitation - the highest daily precipitation ever recorded during the experiment period - elevated the groundwater level by 40 cm (Well A) relative to the previous day. The increase, however, lasted less than one day. Further, groundwater salinity measured close to the water table responded to recharge events as illustrated by well A, which experienced a drop in TDS from 24.5 g/L to 16.5 g/L in response to the high precipitation event in September 2021. Related, TDS is continuously increasing in all three piezometers in response to the 2022 drought event. Interestingly, during the 2022 Nor’easter event in January 2022, when the daily high sea level was at the highest level (1.31 m from NAVD-88) during the experimental period, Well B and C which are close to the salt pond clearly responded by an increase in electrical conductivity, however; the well A did not. This is probably because Well A was already saline.
Figure 5-6: Groundwater response to diurnal tide cycle. The water level amplitude for (a) well A [ocean side], (b) Well B [middle], and (c) Well C [pond side]. The orange vertical line represents the diurnal tide cycle at the site.

Figure 5-6 shows water table response to diurnal tide cycle. Tide induced fluctuations are clearly visible on the ocean and pond side wells with an average amplitude of 1.25 cm. However, the well at the middle of the barrier island (well B) did not show a diurnal tidal signature, indicating that the distance to either the pond or the ocean was sufficiently large (> 85 m) to dampen the tidal signal. The tidal amplitude at the pond measured at the northern side is 0.2 m while it is ~1 m in the Atlantic Ocean.
Rainfall events with cumulative amounts exceeding 10 mm (0.01 m) following two consecutive dry days were recorded along with the resulting increase in groundwater level in well B. The data from other wells was not considered as they displayed a strong correlation with tidal patterns (Figure 5-7), and evaporation loss was assumed to be negligible during high rain events. Data from storm surges and snow events, which can affect the groundwater level, were excluded. Over the experimental period from May 2021 to August 2022, twenty eligible events were recorded. The highest 2-days cumulative

Figure 5-7: Relationship between the rainfall and the groundwater level increase in the test site at well B with the precipitation events >0.01 m. The solid line is a least square linear fit forced through zero due to prior knowledge of the relationship.
precipitation, 98 mm, occurred on September 1 and 2, 2021, causing a 0.31 m rise in well water level. A linear least square fit (R²=0.89) was forced through origin because of prior knowledge on the relationship of water table and precipitation, which yielded a slope of 3.4, which translates to an effective porosity of 0.29, consistent with a fine to medium grained sand and in agreement with the average porosity of aquifer grab samples collected at the time of well installation (0.31).

5.2 Electrical resistivity distribution

Figure 5-8: Electrical resistivity profile along transect A-A'. Measurements during (a) Summer 2019, (b) Fall 2020, and (c) Spring 2021. Red to orange colors represents brackish water (ER 0.1 to 10 Ω.m), light green color is the mixing zone (10 to 100 Ω.m), the blue color (100 to 500 Ω.m) is the freshwater and black color (>500 Ω.m) represents the unsaturated zone above the water table.

Figure 5-8 shows the time-series inverted electrical resistivity profile across the barrier island. For orientation, the dune (berm) is as high as 4.5 m and clearly visible at about 140 m on the cross-distance profiles. The vadose zone is thickest below the berm and gradually
thins out towards the lagoon. On the ocean side, it vanishes over a few meters away from the berm. During summer 2019 (Figure 5-8a), a continuous lens of freshwater up to 2.5 m thick stretched above the brackish water throughout the profile length. As expected, the freshwater lens was thickest below the berm (120th to 160th m distance marker). During Fall 2020 (Figure 5-8b), the freshwater lens was devastated horizontally and vertically in response to the drought. During that time, higher salinization prevailed towards the pond than the ocean site and the mixing zone between freshwater and saltwater widened, particularly below the berm). After the drought, in the Spring of 2021 (Figure 5-8c), the freshwater lens recovered, exceeding the thickness of the 2019 measurements. In general, thinner freshwater lenses (< 5m) are expected to recover quickly (Urish, 2010) which is supported by our data. As measured by the observation wells and seen on the resistivity profile (figure 5-4), the water level is slightly elevated below the dune. In all the measurements, due to negligible capillary fringe zone, the contrast between the saturated and unsaturated zone is fairly high.

Moreover, the water level and TDS on the island respond to the variable recharge; however, the vertical resolution of the ERI profile is too small to correlate the observed depth to the water table to the ERI profile. On the inverted profiles, the water table seems to be at the land surface because the electrodes were buried on the ground to lower the dry sand’s contact resistance. However, the observation wells confirm that the steady-state water table was at 0.5 m from NAVD-88, with a few centimeters elevated at the dune. Overall, the freshwater availability on the island very minimal and is constrained by the reduced recharge during the drought and the comparatively dry summer/fall.
5.3 Numerical simulations

Figure 5-9: Numerical simulation of the transient development of freshwater lens geometry and size on the barrier island. The black dashed curve is the interface between the saltwater and freshwater, defined as the mid concentration of freshwater and seawater (15.75 g/l).

Figure 5-9 illustrates the formation and development of freshwater lens in a barrier island. The initial salt concentration was equal to that of the seawater (31.5 g/L) and the salt concentration in the recharge was 0 g/l. The freshwater lens took 200 years to reach steady state.
Figure 5-10: The freshwater lens interpreted from the (a) ERI data (Figure 5-8a) and (b) calibrated numerical model. The black dashed curve is the interface between the saltwater and freshwater, defined as the mid concentration of freshwater and seawater (15.75 g/l). The white blank spaces on the sides of the ERI profile represents no measurements.

Figure 5-10 compares the saltwater and freshwater interface obtained from the calibrated numerical model and the interpreted ERI image. The numerically simulated geometry of the freshwater lens is consistent with the previous findings at the test site by Urish (1989) [SI II]. The ratio of the thickness of the FWL to the width of the island is 1:50, which is within the range (1:30 to 1:100) described by Vacher (1988). Further, the maximum thickness of the freshwater lens is 2.5 m in both the profiles and both models agree on the depth of the brackish water (below 3 m NAVD-88). However, in the simulation and unlike in the ERT profile, the freshwater lens is skewed towards the salt pond because of the lower
salinity concentration and effective head in the model’s boundary towards the pond than the ocean (Holt et al., 2019; Urish, 1980). In addition, the numerical simulation did not recognize spatial heterogeneity in recharge and K, which can be expected to change the shape of the lens. The freshwater lens thins out because of higher hydraulic conductivity ($K_x$) value at the top layers of the model relative to the recharge ($R$) (the ratio $R/K$ was $1.13 \times 10^{-4}$). The thinner zone of dispersion in the inverted conductivity profile was because of the higher smoothness constraints to mimic the interface between saltwater and freshwater.

![Figure 5-11](image.png)

**Figure 5-11:** (a) Percentage change in TDS during the 2020 Northeastern drought event with reference to steady-state model in salt concentration simulation in groundwater. The black dashed-curve is an isoline of 50% change and the white dashed curve is an isoline of 10% change. (b) Simulated change in TDS at three different locations along the test site by the 202 Northeastern drought.

Figure 5-11 shows the simulated impact of 2020 Northeastern drought on the freshwater lens in the test site. The percentage change is higher at the top layer compared to the lower because the lower layer is already highly saline so the freshwater on the top layer becomes
more saline during drought. For the top 1 m from the water table, the change in salt concentration is more than 50% from the baseline, while the concentration increased at least 10% for the entire freshwater lens. With this the freshwater lens was squeezed by 11% from the baseline. The higher change in the salt concentration is towards the pond followed by ocean, with least change at the center of the landmass. The maximum change towards the pond was 2 g/L while the maximum change towards the pond was 2.5 g/L.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Recharge (mm/yr)</th>
<th>Evapotranspiration (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>800</td>
<td>400</td>
</tr>
<tr>
<td>2</td>
<td>400</td>
<td>800</td>
</tr>
<tr>
<td>3</td>
<td>200</td>
<td>1000</td>
</tr>
<tr>
<td>4</td>
<td>0</td>
<td>1200</td>
</tr>
<tr>
<td>5</td>
<td>0</td>
<td>1400</td>
</tr>
<tr>
<td>6</td>
<td>0</td>
<td>1600</td>
</tr>
</tbody>
</table>

![Figure 5-12](image)

**Figure 5-12:** (a) Future scenarios of recharge and evapotranspiration. During the higher evapotranspiration scenarios, the effective recharge is negative. (b) Simulated freshwater lens contraction in response to variable recharge and evapotranspiration scenarios.

The contraction of freshwater lens in variable recharge and evapotranspiration scenarios are shown in figure 5-12. The FWL is simulated to squeeze by 8% if the recharge rate is reduced in half of the normal recharge for the six month and by 11% if the recharge rate is reduced by 75%. In extreme scenarios, if the recharge is negative because of higher
evapotranspiration (400 mm/year), the freshwater lens shrinks by as more as 16% from the baseline.

Figure 5-13: Freshwater lens response to variable recharge and recovery. (a) before the drought of 2020 summer and fall, (b) during the peak of the drought and (c) the spring of 2021 after six months of drought.

Figure 5-13 shows the devastation and recovery of freshwater lens by the drought and sufficient recharge after the drought respectively. During the 2020 Northeastern drought,
the freshwater lens was shrunk by 11% and largely towards from the ocean side than the pond side (Figure 5-13b). However, with the sufficient (normal) recharge following the winter season, the freshwater lens was recovered into its normal condition as shown in Figure 5-13c. This demonstrates that the drought impacts are seasonal, and the recovers quickly provided normal recharge occurs after drought event.

5.4 Mean residence times (MRT)

<table>
<thead>
<tr>
<th>(a) Volume of the freshwater lens</th>
<th>410 m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Effective porosity</td>
<td>0.29</td>
</tr>
<tr>
<td>Volume of freshwater</td>
<td>119 m³</td>
</tr>
<tr>
<td>Recharge rate</td>
<td>114 m³/year</td>
</tr>
<tr>
<td>Mean residence time</td>
<td>1.05 year</td>
</tr>
</tbody>
</table>

Figure 5-14: (a) Groundwater mean residence times (MRT) calculation with the annual recharge rate, (b) Monthly precipitation and potential evapotranspiration rates in the southern Rhode Island.

There is no clear seasonality in normal precipitation in the southern coast of Rhode Island, however there is a clear seasonality in the evapotranspiration (Figure 5-14b). In some of the months in summer season, such as June, the potential evapotranspiration exceeds
precipitation resulting no or negative recharge of groundwater. Considering 800 mm per year of recharge in a normal year in the test site, the groundwater mean residence times (MRT) is ~1 year, but the time varies seasonally with shorter in the spring season (high recharge) and longer in the summer season (low recharge). The MRT (Figure 5-14a) indicates that in a normal year (without considering drought impacts), the freshwater flushes the island once in every year (MRT ~1 year). However, within the island, the shorter residence time is towards the edge and the longer time is towards the center of the island.

6. Discussions

6.1 Freshwater lens response to reduced recharge

Based on our findings, it was determined through observation well data, the ERI method, and numerical simulations, that there exists a marginal freshwater lens on the test site. The cause of this thin freshwater lens can be attributed to several factors, including the narrow island, higher hydraulic conductivity which permits seawater mixing more rapidly, and the rate of recharge. Our observational evidence and simulations suggest that the freshwater lens is highly sensitive to reduced recharge during drought events, and this is without considering the potential effects of episodic storm surge over-wash and vertical saltwater intrusion. During the summer season, we observed an increase in salt concentration due to higher rates of evapotranspiration. Similar seasonal patterns were also observed in a Mediterranean aquifer in Spain (Palacios et al., 2020).
The exclusive source of freshwater to barrier islands is through aerial recharge, making an adequate amount of precipitation critical to maintaining the freshwater lens (FWL) morphology. However, climate change has resulted in changes in precipitation regimes, higher temperatures, and rising sea levels that have the potential to damage the freshwater lens, thereby reducing the freshwater availability in the future. This situation is more pronounced in low-lying barrier islands, where the topography is limited in its ability to raise the freshwater lens to match sea level. For instance, the northeastern region of the United States is expected to experience a decline in summer precipitation (Rawlins et al., 2012) coupled with increased atmospheric temperatures (Karmalkar and Horton, 2021), resulting in less freshwater recharge and higher evapotranspiration that could potentially shrink the freshwater lens in the area. Conversely, other seasons are expected to experience an increase in precipitation, which would help to flush contaminated seawater back to the ocean. Without considering the effects of vertical saltwater intrusion by storm surge over-wash, the freshwater lens in barrier islands is expected to expand and contrast seasonally.

6.2 Simulation assumption and limitation

The model results are inherently uncertain due to uncertainty in input data, model conceptualization and scenarios. On a hydrodynamically active barrier islands, the interplay between the fresh groundwater and saline seawater at various temporal scales cannot be ignored. Because of limitation of the modeling code, we did not model the impact of vertical saltwater intrusion during the storm surge over-wash. Therefore, to get a complete picture of salinization and freshening, it is important to include all the hydrological controls into account while simulating. However, since the objective of this study was to investigate the drought and reduced recharge impacts on the size, geometry,
and the evolution of the freshwater lens, which we did. In small barrier islands, the impact of short-term rain is nullified because the proportion of the freshwater being added is negligible to flush the contaminated saltwater. However, continuous recharge, in the absence of evapotranspiration, can freshen the saltwater contaminated aquifer and then the volume of freshwater lens increases. Because the test site selected was outside the impact of pumping zone of influence which made it appropriate to tease out the impacts of reduced recharge. However, many barrier islands across the world are being pumped actively, that results further contraction of freshwater lens during the drought events. Another uncertainty arises due to the use of geophysical data alone in calibrating the saltwater intrusion model as the interface calculated from the geophysical techniques are non-unique (Carrera et al., 2010). Solute transport models are difficult to develop and calibrate (Konikow and Reilly, 1999) and one should not expect the as one might expect from the groundwater flow model. Future research would benefit with deeper monitoring wells and with the time-series salinity measurements at different depth. Moreover, the simulation would further benefit with the joint inversion of geophysical data and groundwater model by better constraining the estimation of the hydrologic properties (Singha et al., 2015).

On the test site Urish (1980) estimated the geometry and the size of the freshwater lens. Our results, both from geophysical measurements and numerical simulations, show that the lens is thinner than the author reported (annex II). The discrepancy of ~ 1 m at the center of the island can potentially be attributed to the continuous action of storm surge over-wash and the slow rise in sea level in the past 40 years. Similarly, the analytical model (Fetter, 1972) for the test site using the same hydraulic conductivity value and the recharge rate
estimates a wider freshwater lens than this research (annex I). This shows that the analytical model fails to capture the geometry of the freshwater lens in narrow islands as the model is inherently weak because it neglects the subsurface heterogeneity.

6.3 Broader Implications

Freshwater management in small islands is challenging amid climate change because of the alteration of precipitation regime and increasing evaporation rate due to higher temperature across the globe. Particularly, the Northeastern region of the United States is predicted to get more precipitation in an annual scale, however the summer precipitation is predicted to decline in the future. The freshwater availability during the drought is compromised but with the sufficient recharge after the drought, the freshwater lens recovers to the normal condition. However, since the frequency and the intensity of the storm events are increasing in recent years, it possesses extra threats to water security because once contaminated by seawater vertically, it takes months to several years (Cardenas et al., 2015) to flush the salt back to the ocean.

7. Conclusion

This study has advanced our understanding of a freshwater lens on a barrier island using geophysical data and a numerical model. We demonstrated the applicability of in-situ observation, surface geophysics and numerical simulation in characterizing the freshwater lens and its responses to variable recharge. We developed a groundwater flow and solute transport model using the data observed in the field and the model was calibrated by field observations.
Results demonstrate that the freshwater lens is extremely thin and skewed towards the lagoon because of the relatively low salinity and head in the lagoon than in the ocean. The freshwater lens consequently the availability of freshwater is highly fragile in the barrier-spit island. The recovery of freshwater lens devastation by drought events is short, i.e., within six months after the drought with sufficient recharge to flush out the seawater. Future climate in the region is subject to squeeze the freshwater lens.

Key findings:

- A thin freshwater lens was found because of narrow width of barrier island and higher hydraulic conductivity of the aquifer materials.
- Asymmetric freshwater lens skewed towards salt lagoon was simulated because of the low salt concentration in the lagoon than in the ocean.
- The freshwater lens was devastated by the 2020 Northeastern drought vertically and laterally and the lens shrunk by 11% from the baseline.

Data and code availability

The electrical resistivity inversion code is available at SIMPEG GitHub and can be accessed through https://github.com/simpeg-research/ert-saltwater-intrusion. Postprocessing of the SEAWAT simulation was performed with an open access python package Matplotlib. Raw resistivity profiles are posted to Zenodo for potential reproduction of the research and the data link will be provided here after the manuscript is accepted.
CRediT authorship contribution statement

**Jeeban Panthi**: Conceptualization, Methodology, Software, Validation, Formal analysis, Writing- original draft, Writing – review and editing, Visualization. **Thomas Boving**: Methodology, Resources, Funding, Writing – review and editing. **Soni M. Pradhanang**: Methodology, Resources, Funding, Writing – review and editing. **Seogi Kang**: Software, formal analysis, Writing-review and editing.

Declaration of competing interest

The authors declare that they do not have any known competing personal and financial relationship that could have appeared to influence the research work reported in this paper.

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Holding, S., Allen, D.M., 2015. From days to decades: numerical modelling of freshwater lens


Llopis-Albert, C., Pulido-Velazquez, D., 2014. Discussion about the validity of sharp-interface


Annex I: Analytical model of the freshwater lens

According to Fetter (1972), the maximum thickness of the freshwater lens \( h \) in a single-layered infinite strip oceanic island is given by:

192
\[ h = \frac{R(L^2 - x^2)}{K(1 + \alpha)} \]

where,

\[ \alpha = \frac{\rho_f}{\rho_s - \rho_f} \]

\( R \) is the aerial recharge rate, \( L \) is the half width of the island, \( x \) is the horizontal distance from the center of the island, \( K \) is the lateral hydraulic conductivity, and \( \rho_f \) and \( \rho_s \) are the freshwater and saltwater densities, respectively.

Figure S: Steady-state analytical model of freshwater lens in the East Beach test site, where \( K \) was supposed to be 17m/day. The recharge rates (700 mm/year and 100 mm/year) were chosen to reproduce the freshwater lens during and after the 2020 Northeastern drought in the test site.
Figure S2: Freshwater lens configuration on the test site (Adapted from Urish 1989)
Annex III: Resistivity models

Figure S3: Observed (a) vs modelled apparent resistivity profiles for the test site. The measurement was taken in Fall 2020 (land surface topography not super-imposed)
Chapter 6: Concluding remarks and future research questions

Despite considerable research efforts to understand the coastal groundwater and its response to hydrological and geological controls, knowledge gap exists in characterizing and predicting saltwater intrusion in heterogeneous aquifers (Werner et al. 2013). There are many studies contributing to the understanding of coastal groundwater using geophysical techniques and modeling, however, there is a distinct gap in data-model fusion. A major challenge is to constrain the models with observations to answer pressing science questions (Michael et al. 2017). Incorporating climate change impacts, such as change in temperature and precipitation regime and consequent change in evapotranspiration, is an important but largely understudied area of research in coastal groundwater management and planning (Ketabchi et al. 2016). Therefore, this dissertation attempted to contribute filling these gaps by using various geophysical datasets and in-situ observations and employing these data in simulations to understand water table fluctuations, saltwater intrusion, and submarine groundwater discharge processes, besides other. Though this dissertation cannot answer all questions, it certainly contributes to the field. Some of the major conclusion I have come to, include:

1. Saltwater intrusion observation networks are relatively sparse, and the data sets available are often difficult to correlate across sites because data collection method, sampling frequencies and depth are not standardized (Chapter 2). I recommend maximizing the use of geophysical methods that could potentially fill the gap of point scale monitoring and helps in developing a baseline data for saltwater intrusion. Further, there is significant progress in understanding the process of saltwater intrusion; however, the impacts of vertical saltwater intrusion and
recovery time is largely understudied. More insight into these aspects is important because of changes in climate, and related increase in the intensity and frequency of storm surge over-wash.

2. I verified that the studied aquifer is hydrodynamically active and responds to the seasonal changes in recharge (Chapter 3). The inland aquifer, only 500 m from coastal water, does not show any sign of salinization, but the groundwater level dropped by 0.8 m in response to a regional drought in 2020. A more pronounced drop in the water table elevation could lead to a reversal of the prevailing hydraulic gradient, possibly resulting in SWI under future, prolonged drought conditions. In contrast, the freshwater lens on the studied barrier island shrank during the same drought event. I demonstrated the application of time-lapse electrical resistivity imaging in monitoring saltwater intrusion and terrestrial groundwater discharge in the ocean as a rapid assessment.

3. Bedrock is a geological control of groundwater flow because it is impermeable, except where fractured, and typically is the lowermost boundary of the unconsolidated aquifer (Bohidar et al. 2001; Liu and Tokunaga 2019). It is challenging to develop bedrock topography map with limited borehole lithology data points. To circumvent this issue, I developed a regression model (Chapter 4) for depth to bedrock using ambient seismic measurements, a surface based non-invasive geophysical technique. The model estimated the bedrock depth with high confidence (13% bias) in areas covered by comparatively homogenous glacial outwash deposit. However, the error increased to 23% on average, in glacial till, suggesting that the make-up of the unconsolidated sediments above the bedrock
directly impacted the predictive quality of the model. The model was validated with ERI-interpreted bedrock topography in both near-shore and inland sites and showed a good (<5% depth discrepancy) agreement.

4. Freshwater availability on the barrier island aquifers is extremely susceptible to the change in hydrological controls. By analyzing the time-lapse electrical resistivity data and ground-based observations (Chapter 5), I showed that the freshwater lens at the test site is shallow (<3 m) and was devastated by 2020 Northeastern drought. By constraining groundwater flow and solute transport model for the same site, I showed that the freshwater lens potentially shrinks by as much as 16% from the normal condition in extreme drought condition. The recovery of freshwater lens depletion by drought events is comparably short-lived, i.e., it was reversed within six months after the drought with sufficient recharge to flush out the accumulated seawater. This suggest that under future, drier climate in the region the freshwater lens may be squeezed significantly.

Based on the field investigations and analysis of my manuscripts, there are several recommendations that could serve as science questions for follow-up studies.

1. Firstly, the asymmetrical shape of the freshwater lens in barrier islands is an area that warrants further studies. On our model boundary conditions, both constant concentration and head were higher towards the ocean than the pond, which could be a reason for the observed asymmetry. Conducting further studies using a sand-tank groundwater model could potentially help to explain the asymmetry.
2. Secondly, joint inversion of electrical resistivity profiles and saltwater intrusion models would enable better parameterization of the simulations. Furthermore, having salinity observations from deep wells would be instrumental in validating the geophysical profile.

3. Finally, the validation of fresh SGD calculated by time-lapse geophysical techniques with field observations would be a groundbreaking experiment. Such a study could help to provide further insights into the dynamics of groundwater discharge and its impact on the environment.

4. Overall, I showed that there is a seasonal signature in saltwater intrusion and coastal groundwater dynamics. It is important to validate it in different parts of the world, particularly in reasons where precipitation is highly seasonal such as in the Himalayas.

In conclusion, these recommendations provide a starting point for future studies that could help to advance our understanding of the complex interactions between groundwater, surface water, and the environment.

References


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Supplementary Information

Supplementary I: Field guide video on electrical resistivity imaging

Weblink: https://www.youtube.com/watch?v=59uLmJBgiyA&t=98s

Supplementary II: ERI field set up and site selection manual

Weblink: https://drive.google.com/file/d/1up-56KEJaae9PhzbktN7hftsDMWcXfdr/view