

MONITORING VAN SALINITEIT VARIATIES TIJDENS GETIJDENCYCLI MET BEHULP VAN ELEKTRISCHE RESISTIVITEIT TOMOGRAFIE: EEN HAALBAARHEIDSSTUDIE

MONITORING SALINITY CHANGES DURING TIDAL CYCLES USING
ELECTRICAL RESISTIVITY TOMOGRAPHY: A FEASIBILITY STUDY

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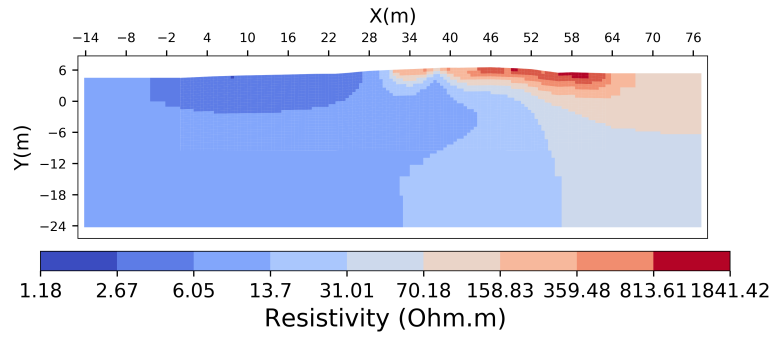
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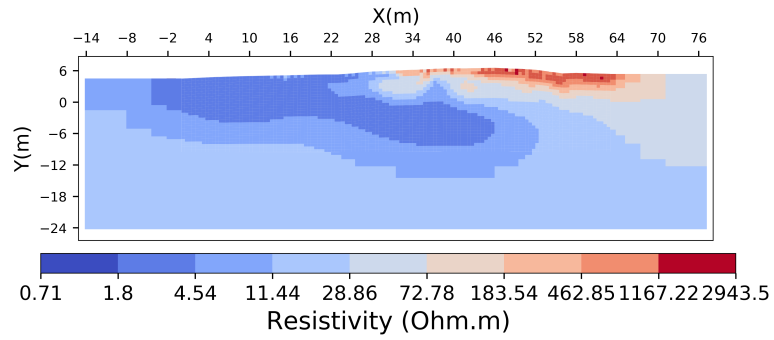
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The story of my research

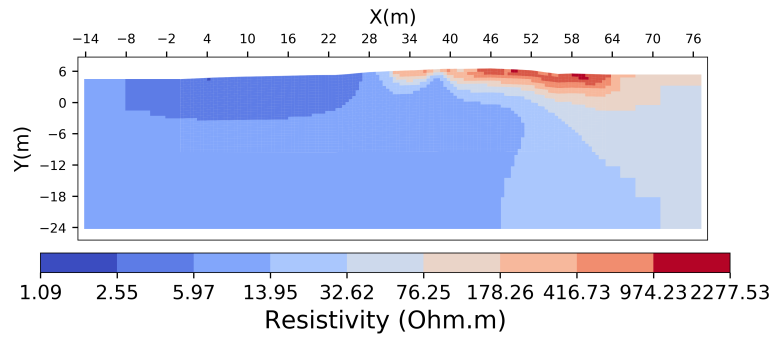
Unlike most of my classmates and predecessors I did not always know I wanted to become a scientist, *au contraire*, it was quite the roller coaster getting here. In primary school I was a very enthusiastic and confident child right up to the moment when my fifth grade teacher, Miss Ann, decided I was just a dumb and slow and she never let an opportunity slip to remind me of it. So, when it was time to decide what I wanted to do in high school I had evolved into an insecure girl with low self-esteem and so I decided to do what I was comfortable with - cooking. Against the will and efforts of my parents and teacher Miss Isabel I started high school at the culinary institute 'Ter Groene Poorte' in Bruges where six years later I graduated as a baker and pastry chef. I cannot say that I enjoyed my time there, it became clear quite quickly that I didn't fit in and I soon realised I needed more stimulus on an intellectual level. I honestly don't know how I got interested in geology but it happened and in September 2014 I decided to take the leap and enroll at Ghent University. Starting a bachelor degree in science with virtually no knowledge of maths, physics or chemistry was certainly not easy but after a lot of hard work I got my bachelor diploma against all expectations - including myself. I used regret having spent four years obtaining it instead of three but because of that I was given opportunities that were otherwise not have been possible, including my internships at VLIZ and BBRI, that have made me into the scientist I am today. Because of my lack of knowledge from high school, I really struggled with my physics courses to the point where I passionately resented the field but ironically I later pursued geophysics (I like to think that geophysics chose me and not the other way around). I spend much time during my internships at VLIZ learning to work with bathymetry and seismics but for my master thesis I wanted to learn something new, so learning to work with ERT was an added advantage. I liked the subject because of the social relevance, saltwater intrusion threatens freshwater supplies in coastal communities around the globe so a good knowledge of this system is needed in order to protect the reserves. But what I liked most and what drew me to the subject is the fact that something similar has never been done before, this made it really exciting to work on, trying out different things, seeing what worked and what didn't was really interesting but also stressful at times. I think the figure below nicely illustrates this, it shows a moment at the beginning of my field data processing in which I tested different error models to see which best suited my data. The last year was really fun, interesting and I learned a lot of useful skills which I'm sure will benefit my future as a geologist. It has all together been an amazing experience.



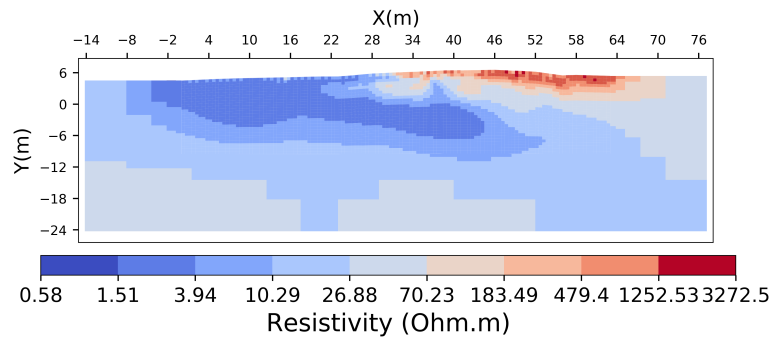
(a) Threshold 10%, envelope error model.



(b) Threshold 10%, trend line error model.



(c) Threshold 2,5%, envelope error model.



(d) Threshold 2,5%, trend line error model.

Samenvatting

In deze master thesis, het effect van getijden op de zoutwater intrusie in de westelijk Belgische kustvlakte is bestudeerd gebruikmakende van elektrische resistiviteit tomografie (ERT). Zoutwater intrusie is de landwaartse infiltratie van zout grondwater, afkomstig van de zee, in de zoetwater aquifer. Zoet grondwater stroomt richting de oceaan waar er zich een transitie zone ontwikkeld door de menging met het meer dense zeewater. Dit leidt in veel gevallen tot een karakteristieke morfologie waar een zoutwater wig zich bevindt onder zoet water. Een zoutwater intrusie ontwikkelt zich in alle kustgebieden en bedreigt de voorraad drinkwater voor veel kustgemeenschappen. Aangezien een grote fractie van de bevolking zich situeert in kustgebieden, geschat op 70 % van de totale wereldbevolking, is de studie van de zoutwater intrusie van groot belang om de drinkwater voorraden te beschermen. De positie van de transitie zone hangt af van het zeewater niveau, er is al veel onderzoek uitgevoerd naar het effect van klimaatverandering en zeeniveau schommelingen op de positie van de intrusie. Het algemene effect is dat een verhoging van het zeeniveau zorgt voor een stijging van de zoutwater stijghoogte waardoor dat het zoute grondwater verder landinwaarts stroomt. Naar de morfologie van de intrusie zijn er twee eindleden, enerzijds leidt dit tot een verplaatsing van de transitie zone landinwaarts. Anderzijds blijft de zone op dezelfde locatie staan maar verandert de breedte van de meng zone waarbij deze scherp is bij laag water maar breed wordt ten gevolge van zeeniveau stijging. Zowel seizoengebonden variaties als fluctuaties op de schaal van een getijdencyclus zijn waargenomen op het zoet- en zoutwater distributie in kustgebieden.

In deze thesis wordt de mogelijkheid van ERT om de zoutwater intrusie te monitoren beoordeeld met het doel om te voordelen en limitaties in te schatten naar toekomstige lange termijn monitoring. ERT is de methode bij uitstek om zoutwater intrusie te karakteriseren en monitoren en wordt dan ook wijd gebruikt voor dit doel. Echter, een monitoring op de schaal van één enkele getijden cyclus waarbij de elektroden gedeeltelijk en variabel bedekt zijn met zeewater is naar de kennis van de auteurs nog niet eerder uitgevoerd.

In de context van dit onderzoek wordt de zoutwater intrusie in de Westhoek bestudeerd, dit is een natuurgebied dat zich bevindt in de Belgische westkust nabij de Franse grens. Dit duingebied is deel van de noordwestelijke Europese duinen en heeft zich ontwikkeld in twee stadia, waarvan de laatste nog steeds actief is. Om de biodiversiteit in de duinen te promoten zijn in het verleden twee artificiële slufter gemaakt in de duinen om bij hoogtij zeewater te laten infiltreren in het duingebied. Het in standhouden van de slufter vergt veel onderhoud wat niet verdergezet is dus de slufter is momenteel terug gesloten door de aanhoudende duinvorming. Het studiegebied bevindt zich ter hoogte van de Oostelijke slufter. Het strand hier is met zijn afmeting van 450 meter bij de breedste van België, verder is het een zeer vlak strand met een helling van amper 1.1 % waardoor het effect van de getijden zeer groot is, bij hoogtij komt het zeewater tot 10 à 20 meter. Het studiegebied is opgesteld uit 30 meter zandige afzettingen die samen het quartair aquifer systeem vormen, de zanden zijn bovenaan fijner en onderaan grover

en hebben een goede doorlatendheid, nabij de Franse grens is de aquifer vrij homogeen verder richting het Oosten wordt ze echter meer heterogeen door de aanwezigheid van kleiige lagen. De aquifer wordt onderaan begrensd door de formatie van Kortrijk, een kleilaag van 100 meter dik die doorgaans als impermeabel beschouwd wordt, wat ook hier het geval is. De hydrogeologische situatie in dit gebied met name de zoet- en zoutwater distributie is uniek, naast het klassieke effect van de zoutwater intrusie zoals beschreven hierboven is er in dit gebied ook een zoutwater lens aanwezig onder het strand, bovenop een zoetwater tong die uitvloeit ter hoogte van de laagwater lijn als onderzeese grondwaterlozing (*submarine groundwater discharge*; SDG). De aanwezigheid van een inverse dichtheid distributie wordt verwacht om onstabiel te zijn maar op deze site is het omgekeerde waar, uitgebreid onderzoek is uitgevoerd naar de origine van deze situatie en hieruit blijkt dat ze in een dynamisch evenwicht bevindt die ontwikkeld met een breed strand met lage helling, gedomineerd door een grote intertidale range, wat het geval is in de Westhoek. De lens ontwikkeld zich doordat zout water infiltreert op het achterstrand, nabij de duinen, tijdens hoog tij. Tijdens laag tij stroomt het geïnfilteerde zeewater zeewaarts en vloeit uit ter hoogte van de laagwater lijn. De onderliggende zoetwater tong is afkomstig van duinen en vloeit uit tot achter de laagwater lijn.

Zoals vermeld is ERT de methode bij uitstek om de zoutwater intrusie te bestuderen, elektrische potentiaal methodes zijn namelijk zeer gevoelig naar variaties in saliniteit, zeewater, dat een hoog zoutgehalte heeft geleid namelijk veel beter stroom dan zoet water. Bij ERT wordt een stroom geïnjecteerd in de ondergrond tussen twee elektroden, vervolgens wordt het potentiaalverschil gemeten tussen twee andere elektroden. Door systematisch de afstand tussen de elektroden te vergroten en de opstelling te verplaatsen langs een transect, kan een 2D doorsnede gereconstrueerd worden van de resistiviteit distributie in de ondergrond. Op het veld worden potentiaal verschillen gemeten maar waar we in geïnteresseerd zijn is de resistiviteit distributie in de ondergrond dus het model die het gevolg is van de gemeten data. Dit is een invers probleem en kan niet analytisch opgelost worden maar een numerieke oplossing is nodig. Een oplossing wordt gevonden door het verschil tussen het model en de data te beperken op een iteratieve manier, waarbij de iteraties stoppen wanneer de gewogen *root mean square error* kleiner wordt dan 1, wanneer er geen verbetering is of als een vooropgesteld aantal iteraties uitgevoerd is. Bij numerieke problemen van dit type zijn er oneindig veel modellen die de data kunnen verklaren daarom wordt een karakteristiek van het model meegenomen in de inversie, om ervoor te zorgen dat het model geologisch plausibel is wordt er vaak gebruik gemaakt van een gladmakende beperking (*smoothing constraint*) die ervoor zorgt dat er geen abrupte veranderingen voorkomen. Dit soort karakteristiek kan echter te streng zijn en resulteert vaak in een model dat te vloeiend is. Om dit effect te beperken kan er extra informatie betrokken worden in het algoritme in de vorm van een referentie model, dit kan elke soort data omslaan die relevante informatie geeft over het studiegebied en kan eerder gemeten ERT data bevatten. In de inversie wordt dan rekening gehouden met deze data, aan dit model kunnen gewichten toegekend worden wat de wetenschapper in staat stelt te bepalen welke delen belangrijk zijn. Hier wordt dit gebruikt om de

zeewater laag expliciet te definiëren. In monitoring studies word er doorgaans gebruikgemaakt van *time-lapse* inversie, hier word de inversie berekend doormiddel van de verschillen van data en model met een achtergrond model, een *time-lapse* inversie focust dus op de verschillen tussen twee datasets in plaats van de absolute waarden. De data in deze studie wordt geïnverteerd op de drie bovenstaand manieren, dus een onafhankelijke inversie, inversie met een referentie model en een *time-lapse* inversie. Bij een *time-lapse* inversie word er verwacht dat de datasets gemeten zijn in dezelfde omstandigheden in deze studie moet de data gemeten bij laag water dus eerst gecorrigeerd worden om te reflecteren hoe de data er zou uitzien bij hoogwater. Een correcte reconstructie van het error-model is zeer belangrijk en word gedaan door het meten van een reciprook profiel, hier worden de posities van de potentiaal en stroom elektroden gewisseld, in theorie moet de meting hetzelfde resultaat geven, een schatting van de error kan dus gemaakt worden op basis van het verschil tussen een normaal en reciprook profiel. Om grondig te zijn zou een reciprook profiel gemeten moeten worden voor elk profiel, aangezien er hier gemeten word in een vlug veranderende dynamische omgeving, was een geen tijd om voor elk profiel een reciprook profile te meten dus er word veronderstelt dat de fout min of meer constant blijft.

Om te probleemstelling aan te pakken werd in twee fasen gewerkt, als eerste word het effect van de water laag gereconstrueerd op synthetische modellen, hier word de focus gelegd op verschillende inversie types en hoe de data erdoor beïnvloed wordt verder word er ook gekeken naar het effect van drogen tijdens laag tij, wat ervoor zorgt dat de bovenste laag een hogere resistiviteit heeft. Vervolgens wordt er getoetst als monitoring op het veld, tijdens een getijden-cyclus, mogelijk is en als de bevindingen en overeenkomst zijn met de synthetische studie. Deze stap is belangrijk om de beperkingen te leren naar lange termijn monitoring en om te bepalen aan welke aspecten nog meer belang gehecht moet worden naar de toekomst toe.

Voor de synthetische studie werd een model opgesteld met een zoutwater lens boven een zoetwater tong, wat de situatie in de Westhoek repliceert. Op dit basis model word een halve getijdencyclus gesimuleerd door om de halve meter de waterlaag toe te voegen. Het potentiaalverschil dat het gevolg is van deze resistiviteit distributie kan gemakkelijk berekend worden doormiddel van het voorwaarts model. Op deze zes datasets worden vervolgens verschillende inversie methoden. Al eerste word de data onafhankelijk geïnverteerd zonder rekening te houden met de waterlaag, de resultaten uit deze inversie hebben een te lage resistiviteit in de bovenste laag van het model omdat het model compenseert voor de afwezigheid van de waterlaag. Hieruit blijkt dus dat het belangrijk is de waterlaag mee te nemen in de inversie. Vervolgens word de data geïnverteerd met behulp van een referentie model, hier verkennen we twee mogelijkheden, in beide word de waterlaag expliciet toegevoegd en een hoog gewicht toegewezen. In de eerste optie inverteren we met een homogene waarde voor de ondergrond, in de tweede optie word een realistische achtergrond gebruikt. Voor de homogene waarde werd de resistiviteit van zoet water gekozen om te garanderen dat de ondergrens van de lens gereconstrueerd kan worden, de zoutwater lens kon beter gereconstrueerd worden met deze methode in vergelijking met de bovenstaande, het expliciet definiëren van de waterlaag is dus aangewezen naar toekomstige studies

toe. Het realistisch referentie model dat hier gebruikt word is het originele model, de resultaten van deze inversie liggen zeer dicht bij het gebruikte model, in een aantal profielen word zelfs een over-fitting geobserveerd, wat vermeden dient te worden. Er werd verwacht dat het gebruiken van een realistisch referentie model de resultaten nog verder zou verbeteren waardoor kleine variaties geobserveerd kunnen worden. Dit is hier niet het geval, naar de toekomst toe zou in plaats van het echte model het inversie resultaat bij laagwater gebruikt kunnen worden als referentie model. Een belangrijke observatie is dat, bij alle inversie types het model bij hoog water veel artefacten en vervormingen bevat, de geometrie van de zoutwater lens is volledig verloren gegaan bij de inversie. Dit kan toegeschreven worden aan de zeewater laag, wanneer een dikke waterlaag aanwezig is gaat een grote fractie van de stroom verloren in de geleidende waterlaag waardoor in plaats van de ondergrond te penetreren, dit zorgt voor de sterke vervormingen geobserveerd in alle modellen bij hoog water. Wanneer de absolute verschillen berekend worden tussen profielen verschijnen er dan ook complexe patronen die het gevolg hiervan zijn. Naast het effect van de waterlaag wordt er ook gekeken naar het effect van een drogen tijdens laag tij. Een laag met lage verzadiging bovenaan het model heeft een significant hogere resistiviteit, dit is ook gereflecteerd door de inversie waarin de droge laag duidelijk teruggevonden kon worden. De *time-lapse* inversie op deze dataset was van ondermaatse kwaliteit zowel bij het verzadigd als droge model. Bij het droge model kon de correctie niet goed berekend worden, er is een zeer groot verschil tussen de geobserveerde en berekende data, terwijl dit verschil zo klein mogelijk moet zijn. Aangezien de correctie geen goed resultaat leverde is logisch dat de *time-lapse* inversie ook ondermaats is, de oplossing stopt bij een rms van 106.2 % wat veel te hoog is. Bij het verzadigde model zijn de resultaten van de correctie veel beter wat ook gereflecteerd word in de *time-lapse* inversie. Alhoewel de resultaten hier beter zijn, zijn ze nog altijd niet goed genoeg, hier stopt de inversie bij een rms van 14.6 % wat nog steeds te hoog is. De inversie focust zich waarschijnlijk op de water laag, een extra beperking waarin de resistiviteit van de water laag constant blijft zou kunnen toegevoegd worden aan de inversie om dit probleem te overkomen.

Voor de veld studie werden twee profielen gemeten, langs hetzelfde transect. Een lang profiel met grote afstand tussen de elektroden, om diepe variaties te detecteren en een kort profiel met een kleine afstand tussen de elektroden, om subtiele variaties ter hoogte van de transitie zone te detecteren. Tijdens de meetcampagne werden vier volle getijdencycli gemeten maar uit analyse achteraf blijkt de kwaliteit van de laatste twee ondermaats was dus deze profielen zijn niet meegenomen in de inversie. Dit kwaliteitsverschil kan mogelijks te wijten zijn aan een slechte contact resistentie dus in de toekomst kan dit vermeden worden door dagelijks zeewater over de elektroden te gieten tijdens een meetcampagne. De inversie werd gedaan doormiddel van een referentiemodel, zowel een homogeen als realistisch zijn gebruikt. Aangezien de time lapse inversie een slecht resultaat leverde voor het synthetisch model – wat een best-case scenario is – dient dit eerst verder onderzocht te worden vooraleer een toepassing op de velddata mogelijk is. Voor de inversie met een realistisch referentie model werd de inversie van het eerste profiel met een homogeen referentiemodel gebruikt als achtergrond voor de volgende profielen. Gelijkaardige

resultaten zijn bekomen voor beide methoden, de invloed van het stijgende zeewater is duidelijk zichtbaar, zelfs voor het water de opstelling bereikt, dan word het sediment al gesatureerd met zout water, wat zeer duidelijk is in de profielen. Als de waterlaag aanwezig is veroorzaakt die terug sterke verstoringen die vooral duidelijk worden wanneer de verschillen gevisualiseerd worden. Tijdens de tweede cyclus van het korte profiel zijn er significant meer artefacten aanwezig in de data waardoor het moeilijk is om mogelijke variaties correct te interpreteren. Dit is hoogstwaarschijnlijk te wijten aan een ongepast error model, dat gebaseerd was op het eerste reciprook profiel. In toekomstige campagnes moet hier meer aandacht aan gegeven worden, wanneer het meten van een reciprook profiel voor elk normaal profiel niet mogelijk is kan er ook geopteerd worden om een partiele reciproke te vergaren, dit zal al een betere benadering geven voor het foutenmodel. Een realistisch referentiemodel kon kleinere variaties visualiseren in de data, inversie met een *time-lapse* schema is dus veelbelovend om mogelijke fluctuaties in de overgangszone te visualiseren maar de methode dient eerst op punt gesteld te worden.

Chapter 1

Introduction

1.1 Seawater intrusion in coastal aquifers

1.1.1 General introduction

Seawater intrusion (SI) is the landward infiltration of saline groundwater in the freshwater coastal aquifer, it develops because fresh water is flowing towards the ocean where a transition zone develops between saline and fresh water due to mixing with the dense seawater (Vandenbohede and Lebbe, 2006). Badon-Ghyben (1888) and Herzberg (1901) independently observed this phenomenon in static water bodies and attributed it to the density difference between fresh and salt water. They found that the interface is present at about forty times the freshwater head above sea level, this relation, known as the Ghyben-Herzberg equation has been widely used to describe SI (Bear et al., 1999). The most common case of SI is where the interface is inclined landwards creating a characteristic salt water wedge with fresh water on top (Fig. 1.1). The SI system is affected by both natural and anthropogenic processes including changes in land use, pumping of groundwater, climate change and sea level variations. Because it is a dynamic system and the development and geometry depend on a multitude of factors, the nature of a SI is highly site specific (Werner et al., 2013).

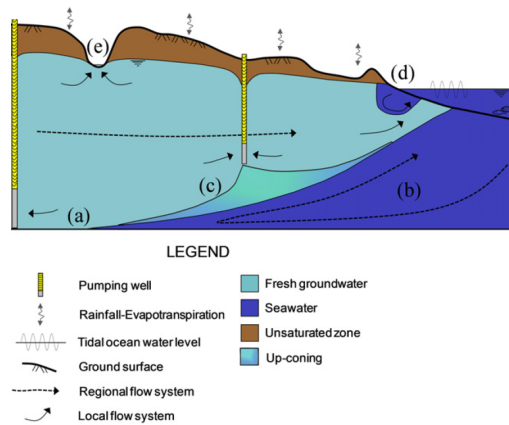


Figure 1.1: Simplified hydrogeological representation of seawater intrusion and its most common characteristics and processes: (a) wedge toe, (b) density driven circulation, (c) upconing, (d) coastal fringe processes and (e) head-controlled surface expression of groundwater (Werner et al., 2013).

SI develops as a result of fresh groundwater flowing toward the sea, this fresh water discharges at the limit between land and sea, known as submarine groundwater discharge (SGD), and is present either as submarine springs or unconfined seepage. SGD does not only refer to the fresh water discharge (FSGD) but also includes recirculated seawater discharging at the land-ocean boundary (RSGD). The latter is caused by wave set-up, tidal oscillations, thermal and density driven convection. SI and SGD are complementary processes both depending on hydraulic and density gradients (Taniguchi et al., 2002).

The presence of SI can have a negative effect on the availability of potable water for coastal communities, it causes a decrease in the storage volume and contaminates pump-

ing wells. Admixture of less than 1% of seawater leads to water that does not meet drinking

water standards (WHO, 2011; Werner et al., 2013). SGD also has negative effects on its environment, it forms a gateway for contaminants to the ocean, most of which are nutrients from fertilizers and sewage (Taniguchi et al., 2002). This can lead to eutrophication and blooms of harmful algae (Hwang et al., 2005; Lapointe et al., 2005) negatively effecting the biodiversity.

A significant fraction of the Earth's population lives in coastal areas with an estimated 41% of the European population (Eurostat, 2011) and up to 70% worldwide. Studies show that this already large fraction of the population will continue to grow in the future (Neumann et al., 2015), as a result there is a high and increasing demand for drinking water in these regions. The presence of SI poses a severe threat to the availability of fresh water on a global scale (USGS, 2000; Barlow and Reichard, 2010; Felisa et al., 2013). Therefore it is important to understand the mechanisms behind the fresh and saltwater distribution and the effect of anthropogenic activities.

1.1.2 Sea level fluctuations

Many studies have been made to characterize the effect of climate change and in particular sea level rise (SLR) on seawater intrusion (Werner and Simmons, 2009; Oude Essink et al., 2010; Chang et al., 2011). The general idea is that SLR induces a higher saline head which causes salt water to replace fresh water and as a result the landward movement of the interface (FAO, 1997). The reverse is also true, lowering of the sea level causes a relative rise of the freshwater head that will lead to a displacement of salt by fresh water and consequently a seaward movement of the interface (Michael et al., 2005). Important to note is that this shift can also be caused by a change in freshwater head, for example by differing recharge or extraction rates.

One of the major factors influencing SI is hence the groundwater level. According to White and Roberts (1994), the influence of tidal variations on the groundwater level in aquifers depends on the type of aquifer (confined or unconfined) rather than the permeability or hydraulic conductivity. In confined aquifers the influence can extend up to 100 meter landward whilst in unconfined aquifers the effect has a more limited extend of approximately 30 meters. There are essentially two possible effects caused by the tidal activity, some researchers observe a shift of the interface while the thickness of the mixing zone remains constant, the interface moves landwards during high tide and retreats during low tide (Morrow et al., 2010; Gonneea et al., 2013). In contrast it is also possible that the tides do not influence the position of the interface but affect the thickness of the mixing zone, with a sharp interface at low tide but a broad mixing zone during high tide (Inouchi et al., 1990; Kim et al., 2008).

Because SGD and SI are complementary processes and one of the main driving forces behind SGD is tidal pumping, it is a logical consequence that SGD is also influenced by tides. Kim and Hwang (2002) observed semi-monthly fluctuations provided that the aquifer is sufficiently recharged. Taniguchi et al. (2006) also measured semi-diurnal variations and found that the influence is largest just landward of the salt and freshwater interface.

It is clear that the fresh-salt water distribution is effected by tidal fluctuations. Both SI and SGD show significant variations on the seasonal scale but also on the scale of a single tidal

cycle.

1.2 Electrical resistivity tomography in saltwater intrusion and submarine groundwater discharges studies

The salinity difference of fresh- and saltwater induces a difference in electrical conductivity, with the latter being significantly more conductive. Geo-electrical geophysical methods are sensitive to conductivity variations and are thus well suited to study the fresh/saltwater distribution in the subsurface. Nguyen et al. (2009) showed that ERT is particularly well suited to characterize and monitor this distribution and is hence been widely applied, on different scales and in various settings.

Ronczka et al. (2015) used long electrode ERT measurements to monitor saltwater in a phreatic aquifer in Müllrose, Germany. They used existing boreholes and showed that, when using structural a-priori information in the inversion, the salt water lens could be accurately imaged even when using irregular electrode spacings and depths.

Goebel et al. (2017) studied the seawater intrusion in Monterey Bay, central California on a very large scale (kilometres), here several freshwater basins are threatened by SI. Long-offset ERT profiles (electrode spacing of 22.5 m) were collected on the beach, along 40 km of the coast, combined interpretation of ERT data and borehole measurements allowed them to characterize the geometry of the seawater intrusion and revealed complex spatial patterns on the basin-scale.

Paepen et al. (2020) thoroughly studied SGD in the Belgian western coastal plain by combining different geophysical methods. In this case the discharge zone extends beyond the low water line, so to visualize the full lateral and vertical extent multiple measurements are combined. At low water terrestrial roll-along ERT profiles are measured from the low water line to the dunes. At high water, marine continuous resistivity profiling is used, overlapping the terrestrial profile. Combination of these methods image the vertical extent of the SGD, the lateral variations were measured with frequency domain electromagnetic mapping (FDEM). Combination of these three methods proved necessary to acquire a high resolution, 3D image of the SGD.

These are but a few examples, however there are many more cases in which ERT is used to characterize and monitor the salt- and freshwater distribution (e.g. Henderson et al., 2010; Hermans et al., 2012; Koukadaki et al., 2007).

Although ERT has been widely used in SI studies, including monitoring, monitoring on the scale of a single tidal cycle is not common, Swarzenski et al. (2006) did this at Dor Beach, Israel. A profile perpendicular to the beach was measured at low and high tide with the electrodes on the seafloor, also at high water. The used inversion software (AGI's EarthImager) accounts for the salinity and depth of the water. The interface only showed a subtle response to the rising tide, note that there is a small tidal range of only 24 cm. To our knowledge no other studies apply ERT to monitor SI on such a short time scale.

1.3 Objectives and approach

In the scope of this thesis, the fresh-salt water distribution in the western Belgian coastal plain will be monitored, more specifically in the nature reserve "The Westhoek". Here a special case of seawater intrusion is present, a classic salt water wedge exist but is accompanied by a seawater lens above a freshwater tongue under the beach. Contrary to what might be expected, this inverse density distribution is in a stable, dynamic equilibrium. The lens develops when salt water infiltrates the beach on the back shore during high tide, fresh water from flows under the lens and discharges around the low-water line forming the freshwater tongue (Lebbe, 1981; Vandenbohede and Lebbe, 2006; Paepen et al., 2020).

The seawater lens is present in the intertidal zone, which will be challenging to monitor due to its dynamic nature. To our knowledge only Swarzenski et al. (2006) have made a similar study, however, the tidal range in this area is only 24 cm while in the Westhoek this can include the entire width of the beach. Moreover, they are only comparing two time steps while in this work we aim at monitoring full tidal cycles.

The main objective of this thesis is to asses the feasibility of ERT to monitor the resistivity distribution in the intertidal zone during tidal cycles, hence when the electrodes are partially and variably inundated. One major problem is that due to the high conductivity of seawater a large fraction of the current is transmitted in the water layer instead of penetrating the subsurface, leading to strong distortions in the resulting images and a loss of sensitivity at high tide (Henderson et al., 2010). It is thus very important to properly account for the water layer in the inversions. If successful, the methodology could be later extended to long-term monitoring to study the seasonal variations in the fresh/saltwater distribution.

In order to asses the possibilities and limitations of monitoring in a dynamic environment, we will first conduct a synthetic study followed by a comparison with the field data. In the synthetic study, we will asses the influence of the water layer on the inversion results using different approaches and also estimate if it is possible to retrieve the salinity variation by appropriate processing and inversion. In a next step we will verify if field monitoring is possible and if the results are in accordance with the synthetic study. This will be important to learn the limitations for future long term monitoring.

Lastly, the structure of the thesis will be shortly illustrated. Chapter 2 includes a more thorough description of the geology and hydrogeological evolution of the study area, it is described separately from the field data because this specific situation is simulated in the synthetic study. In chapter 3 the theory concerning ERT is discussed, including the approaches used in both the synthetic and field study. Chapter 4 encompasses the synthetic study, the results of which are necessary to interpret the field data, described in chapter 5. Lastly a general conclusion is made in chapter 6 comparing the synthetic and field data and including future recommendations.

Chapter 2

Study area

2.1 Situation and generalities

In the context of this thesis, the saltwater intrusion of the western Belgian coastal plain is monitored, more specifically in the nature reserve "The Westhoek" (Fig. 2.1). The Belgian dunes are part of a long narrow dune strip from Calais (France) to Northern Denmark, the north-west European coastal dunes. These dunes have developed in two stages, the first stage or the old dunes formed 2000 to 5000 years ago, a second stage or the new dunes started developing from the eight century forward and is still active (Vandenbohede and Lebbe, 2006).



Figure 2.1: Nature reserve "The Westhoek" with indication of the study area (blue marker).

The Westhoek is located at the Belgian-French border, with its size of 380 ha, it is one of the largest unfragmented dune areas in Belgium (Vandenbohede and Lebbe, 2007). In this reserve, two tidal inlets were made to allow seawater to enter the dunes during high tide, this forms a unique environment that promotes biodiversity and creates the opportunity for new habitats to develop (Verwaest et al., 2005). A bridge was built over the eastern inlet, as an entry gate, separating the beach from the dune area. Preservation of these artificial inlets would require high maintenance that unfortunately did not take place, currently they are closed due to dune forming, but the bridge is still present. The site that was monitored in the scope of this project is located at the eastern tidal inlet (Fig. 2.1).

The study area is dominated by semi-diurnal tides, with a range of 5 m at spring tide and 3 m at neap tide, the shore is relatively flat with a slope of 1.1 % and has a sandy runnel ridge

morphology, typical for shallow gradient beaches. The beach is between 300 and 450 m wide, making it one of the widest shores in Belgium, the mean high tide comes very close to the dunes (Vandenbohede and Lebbe, 2006) with a distance of approximately 10 to 20 m.

2.2 Geology and hydrogeology

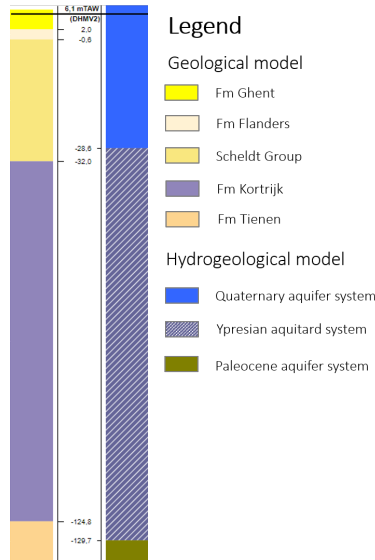


Figure 2.2: Virtual core of the study site, location indicated on figure 2.1 (*Databank Ondergrond Vlaanderen*)

Figures 2.2 and 2.3 show the geology and hydrogeology of the study area, the dunes consist of a 30 m thick sandy deposit, containing the formation of Ghent (Member of De Panne), formation of Flanders and the Scheldt Group (*Databank Ondergrond Vlaanderen*). The De Panne Member are aeolian sand deposits of Holocene age with a fine to medium grainsize, intercalated with peaty, humic or shell layers (Lebbe and De Ceunynck, 1980). The formation of Flanders is also of Holocene age and consists of three distinct facies: coarse sand deposited in tidal channels, fine wadden sand and tidal flat clays (Paepe and Baeteman, 1979). The Scheldt Group is of Pleistocene age and consist of clay, sand and gravel. These three stratigraphic units compose the Quaternary aquifer system with fine sands at the top and coarser sand at the bottom. The aquifer is relatively homogeneous at the border but becomes more heterogeneous towards the east with the presence of inter-connecting clay lenses forming semi-permeable

layers. At least two clay layers are present but there could be more, these two are the ones identified by drillings (Lebbe, 1978). The sandy aquifer is confined at the bottom by the Kortrijk formation, a stiff clay layer of approximately 100 m thickness of early to middle Ypresian age (*Databank Ondergrond Vlaanderen*). In hydrogeological studies it is generally considered as an impermeable layer, as will be the case in this thesis (eg. Vandenbohede et al., 2008; Hermans et al., 2012).

The hydrogeological situation show in figure 2.3 developed during the Holocene. Formation of the dune belt prevents seawater to enter the hinterland so the saline pore water became gradually replaced with fresh recharge water. As a result, a fresh water lens formed under the dunes between saline polder water to the south and seawater to the north, this fresh water lens has been exploited since 1967 by the local water company IWVA (nl: *Intercommunale Watermaatschappij Veurne Ambacht*; Vandenbohede et al., 2008). The origin of the saline polder water can be attributed to the morphology and hydrogeological development of the area, the mean level of the polder is below the high water mark thus a dense drainage system is needed

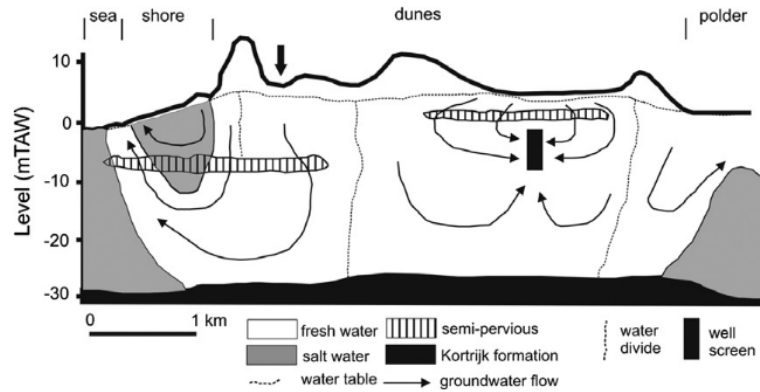


Figure 2.3: Hydrogeological cross section of the study site (Hermans et al., 2012).

to evacuate the excess water. Consequently only a limited amount of water is left to recharge the aquifer and freshening of the polders occurs much slower than the dunes, as a result fossil seawater is still present in the deeper parts of the aquifer (Vandenbohede and Lebbe, 2006; Vandenbohede et al., 2010).

Part of the fresh water flows towards the polder area and discharges along the dune-polder boundary, the other part flows towards the sea, it was believed that this water discharges in the intertidal zone near the low water line, however, recently Paepen et al. (2020) showed that part of the fresh water discharges offshore, below the low water line. Above this freshwater tongue a salt water lens is present that develops due to the infiltration of seawater on the back shore at high tide, during low tide the salt water discharges on the fore shore. Currently the lens does not longer extend towards the sea as indicated in figure 2.3, the freshwater tongue now reaches the surface on the lower beach. This is most likely an effect of the decreasing pumping rate of the IWVA exploitation site, strengthening groundwater discharge. Because of the IWVA groundwater extraction, there are two main groundwater divides present instead of one and they are located north and south of the extraction delimiting water flowing toward the sea, polders and the extraction. Also the presence of the semi-impervious layers have an influence on the flow pattern. One clay layer is located in the centre of the dunes above the extraction, at 0 m TAW, the second is located under the shore and fore dunes at -5 m TAW, these layers have low hydraulic conductivity that hinders vertical flow, leading to enhanced horizontal flow and two local water divides (Vandenbohede et al., 2008).

2.3 Development and evolution of the seawater lens

The unique situation of the inverse density distribution with a saltwater lens above a freshwater tongue is in dynamic equilibrium (Lebbe, 1981). Vandenbohede and Lebbe (2006) studied this situation using numerical modelling, they simulated the evolution of the salt water lens and

found that after 200 years no major changes occurred, confirming the dynamic equilibrium. They also stated that this particular distribution only develops when a wide low sloping shore is present dominated by a large tidal range and low permeable sediments. However, a recent study has shown that the same unique situation is present in "The Paardenmarkt" near Knokke, where the slope is steeper than in the Westhoek (Hermans, 2020, personal communication). The morphology of the salt water lens and thickness of the mixing zone depends on the recharge rate in the dune and the infiltration on the shore, although the simulations are made in steady-state conditions, it is safe to assume that seasonal variations in recharge and tidal range will effect the SI/SGD equilibrium (Vandenbohede and Lebbe, 2006).

In a later study, Vandenbohede and Lebbe (2007) studied the flow cycles in more detail using numerical models but this time with a transient flow model to account for the tidal influence. They found that the discharge/recharge pattern is achieved by two interfering flow cycles. The first one is the fresh water flowing from the dunes and discharging at the low water line creating the fresh water tongue, this process is present throughout the tidal cycle. The second flow cycle is more complex and determines the presence of the salt water lens. During low tide salt water flows from the back shore to the fore shore where it discharges, in the course of a tidal cycle the discharge zone shifts with the position of the waterline.

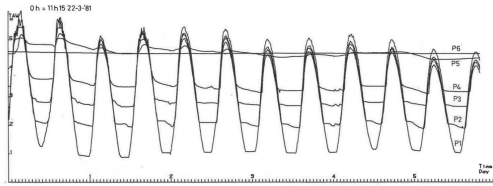


Figure 2.4: Fluctuation of the fresh water head in the shallow observation wells, P1 is located closest to the low water line, P6 in the dunes and the other wells subsequently in between (Lebbe, 1981).

The previous paragraphs clearly indicate the influence of seasonal and tidal variations on the fresh-salt water distribution, variations in freshwater head have been observed in boreholes, confirming this. Lebbe (1981) monitored eight boreholes in the Westhoek, six at -2 mTAW and two at -22 mTAW - all located along the same profile - for six days. Fluctuation in freshwater head was measured in most of the wells (Fig. 2.4), the response is

strongest in the wells located near the low water line and declines when moving further from the sea (P1 to P4), on the back shore (P5) the effect is significantly smaller and in the dunes (P6) it is non-existent. In the deep wells a similar trend is observed. Vandenbohede and Lebbe (2007) made similar observations, wells close to the sea showed a quick and strong response to the upcoming water, further toward the dunes the response had a time lag and was more attenuated. These changes in head will influence the flow and thus salinity distribution but not necessarily to a large extent. Whether or not changes will be captured by ERT will depend on the amplitude of the salinity variations.

Chapter 3

Electrical Resistivity Tomography

3.1 Introduction

The electrical resistivity ρ [Ωm] is a material property characterizing how easily a current can flow through it. The resistivity actually represents how resistant a material is to electrical current, thus a low resistivity means that current can flow easily and vice versa. The electrical conductivity σ [S/m] describes the same process but is the inverse of the resistivity. Resistivity can not be directly measured but the resistance R can according to Ohm's law. The resistance is proportional to the geometry of the material, the constant linking them is the resistivity (Reynolds, 2011).

$$R = \frac{\rho L}{S} \quad (3.1)$$

With L the length and S the section of the sample.

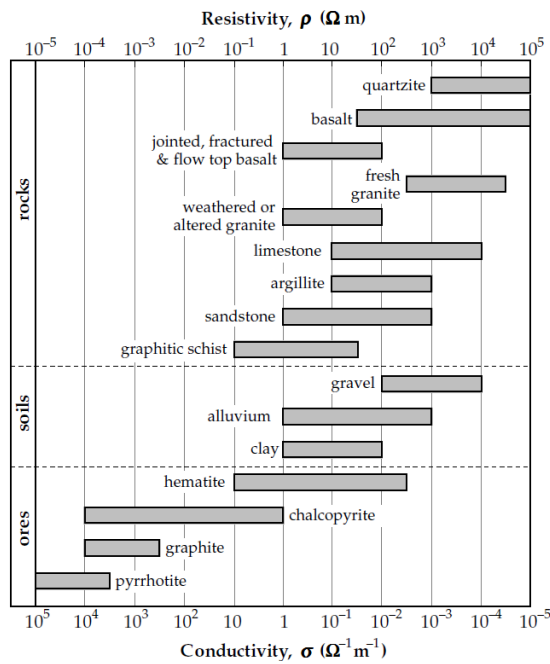


Figure 3.1: Resistivity and conductivity of common geological samples (Lowrie, 2007).

is the movements of ions in the pore fluid, waters with high salinity are hence more conductive. Surface conduction are preferential flow paths present at the surface of grains and develops because of the electrical double layer present at minerals with charges at the surface, this type of

Resistivity methods are widely used in hydrogeology, topics contain aquifer characterization (e.g. Robert et al., 2012; Doetsch et al., 2012), contamination studies (e.g. Grünhut et al., 2018; Casas et al., 2008), riverbed characterization (e.g. Benoit et al., 2019), seawater intrusion (e.g. Ronczka et al., 2015; Nguyen et al., 2009) and submarine groundwater discharge (e.g. Paepen et al., 2020). The method is also used in other study areas like mineral exploration (e.g. Zhang et al., 2018), geotechnical studies (e.g. Sudha et al., 2009), archaeology (e.g. Negri et al., 2008) and forensic science (e.g. Rubio-melendi et al., 2018). And thus proves to be a very versatile method.

The resistivity of a geological sample is highly variable (Fig. 3.1) and can be divided in three mechanisms, metallic conduction occurs within the grains of conducting and semi-conducting minerals. Electrolytic conduction

conduction is especially important for clay minerals (Revil et al., 2012).

Rock forming minerals are generally not conductive so in sedimentary rocks the electrolytic and surface conduction are the dominant mechanisms. When the pore fluid is conductive, for instance salt water, the surface conductivity of clay minerals is proportionally less important than the electrolytic conductivity. The resistivity of sediments and sedimentary rocks is described by Archie's law, the bulk resistivity ρ_b can be calculated when the porosity ϕ , saturation S and resistivity of the pore fluid ρ_w are known (Archie, 1942).

$$\rho_b = \frac{a}{\phi^m} \frac{1}{S^n} \rho_w \quad (3.2)$$

Where m is the cementation exponent, generally between 1.3 and 2.5 for sediments, a is an empirical factor approximately equal to 1 and n is the saturation exponent, usually equal to 2, these factors are mainly controlled by the pore space geometry (Nguyen et al., 2009). For sediments with high clay content a term accounting for the surface conductivity can be added (Waxman and Smits, 1968).

The bulk resistivity is influenced by a wide range of rock and sediment properties. This is both the advantage and drawback of the method, it can be used in various contexts to study multiple subsurface properties but on its own the results are not informative, additional knowledge is necessary to interpret resistivity results.

In this work, the salt- and freshwater distribution is studied during tidal cycles. A change in salinity will influence ρ_w , given that the saturation remains the same the bulk resistivity will change proportionally. However, a change in saturation, will also influence the bulk resistivity so if both are present the outcome might be difficult to predict.

3.2 Principle of ERT

With electrical resistivity tomography (ERT), a current is injected in the subsurface with a pair of electrodes (current electrodes), the resulting potential difference is measured between a second pair of electrodes (potential electrodes). The potential distribution upon current injection is described by the Ampere-Maxwell equation coupled with Ohm's law (Telford et al., 1990).

$$\nabla \cdot (\sigma \nabla \phi) - \nabla \cdot \vec{J}_s = 0 \quad (3.3)$$

With $\vec{J}_s$ [A/m²] the current density from the source, σ [S/m] the electrical conductivity and ϕ [V] the electrical potential.

When considering a homogeneous medium, an analytical solution to this problem can be calculated, for a classical quadrupole setup this is:

$$\Delta \phi = \frac{\rho I}{2\pi} \left(\frac{1}{r_A - r_M} - \frac{1}{r_A - r_N} - \frac{1}{r_B - r_M} + \frac{1}{r_B - r_N} \right) = \frac{\rho I}{K} \quad (3.4)$$

With r the position of the respective electrodes, A and B are the injection electrodes, M and N the measuring electrodes, ρ [$\Omega \cdot m$] is the electrical resistivity and I [A] the injected current.

The denominator describes the geometry of the setup, these terms are thus represented by the geometrical factor K (Reynolds, 2011).

Equation 3.4 is only valid for a homogeneous medium, in field cases this condition is often not met and the apparent resistivity ρ_a is measured instead of the true resistivity ρ . As a consequence equation 3.3 can not be solved analytically and a numerical solution has to be sought using numerical methods (Sasaki, 1994; Negri et al., 2008).

3.3 Inversion

The objective of data inversion is to find a model that explains the observed data so that:

$$d = f(m) \quad (3.5)$$

Equation 3.5 formulates the forward problem where d is the observed data, m is the model parameters and f the non-linear, forward model operator describing the relation between d and m . In ERT the model parameters consist of the resistivity distribution that explains the observed potential difference, the forward model is expressed in equation 3.3. The data and model parameters can assume a wide range of values encompassing several orders of magnitude and are thus usually represented as their logarithms.

The inversion is done by minimizing the difference between the model and the observed data under some *a priori* characteristic of the model, i.e. minimizing the objective function (Tikhonov et al., 1995).

$$\psi(m) = \psi_d(m) + \lambda \psi_m(m) \quad (3.6)$$

The first term in equation 3.6 is the data misfit and the second the model misfit which includes the assumed characteristic of the model, λ is the regularization parameter, balancing the data misfit and the assumed characteristic.

The data misfit $\psi_d(m)$ is generally calculated as the L_2 and can be written as:

$$\psi_d = \left\| \frac{d - f(m)}{\epsilon} \right\|^2 = \sum_{i=1}^N \left(\frac{d_i - f_i(m)}{\epsilon_i} \right)^2 = \|W_d(d - f(m))\|^2 \quad (3.7)$$

Where ϵ_i is the standard deviation at data point i , W_d is the data weighting matrix, a diagonal matrix with $1/\epsilon_i$ as elements. The minimization of the objective function is done through iterations until the misfit reaches a predefined level (Hermans, 2014). In this work, the target misfit level is estimated from noise estimate so that data are fitted within their noise level (Kemna, 2000).

When inversion is executed based solely on the data misfit it would lead to a noisy, implausible often over-fitted model (Tikhonov et al., 1995). The second term in equation 3.6 resolves this problem. A smoothness constraint is often used as the model characteristic and is defined as a spatial derivative of the model, the objective is thus to have minimum roughness (Constable et al., 1987).

$$\psi_m(m) = \int \int \|\nabla m\|^2 dx dz \cong \|W_m(m)\|^2 \quad (3.8)$$

W_m is a matrix evaluating the roughness of m and thus ensuring that the result will be a smooth model. However, a smoothing constraint often results in a model not consistent with the expected geological and hydrogeological processes so that alternative regularization have been developed (e.g. Hermans et al., 2012; Nguyen et al., 2016). However, these alternatives are not investigated in this work.

Equations 3.7 and 3.8 can be substituted in equation 3.6 giving the objective function expressed in terms of the defined operators (Kemna, 2000)

$$\psi(m) = \|W_d(d - f(m))\|^2 + \lambda \|W_m(m)\|^2 \quad (3.9)$$

Equation 3.9 has to be solved iteratively, at each step the regularization factor is optimized through a line search to minimize the error weighted root mean square error. The inversion stops when the error weighted root mean square error (ϵ_{rms}) reaches 1, after a predefined number of iterations or when there is no improvement of the ϵ_{rms} between successive iterations. The inversions of both the field and synthetic data are done with the CRTomo code. This algorithm uses an iterative Gauss-Newton scheme to solve the inverse problem and a finite element approach to solve the forward model (Kemna, 2000).

3.4 Error estimation

Equation 3.9 includes the standard deviation given as a matrix $W_d = \text{diag}(\frac{1}{\epsilon_1}, \frac{1}{\epsilon_2}, \dots, \frac{1}{\epsilon_N})$, where N is the number of data points. The standard deviation ϵ_i , measured resistances R and error on the measurements e_i are related to each other in the following way:

$$\epsilon_i = \Delta \ln R_i \cong \frac{\Delta R_i}{R_i} = \frac{e_i}{R_i} \quad (3.10)$$

The error can be estimated through the reciprocal error (LaBrecque et al., 1996). In a reciprocal measurement the injection and measuring electrodes are switched, theoretically the measurements should be the same (Parasnis, 1988). The error on the data can be estimated by the difference between the normal and reciprocal measurements.

$$|e_{N/R,i}| = R_{N,i} - R_{R,i} \quad (3.11)$$

With the term on the left equal to the error on measurement i , R_N and R_R are respectively the normal and reciprocal measurements. The error can then be used to weight each individual data point.

When an estimate of the error is not available for each data point or each time-step, it might be more practical to derive an error model. In this work, an error model where the magnitude of the error $|e|$ increases linearly with the resistance $|R|$ is used.

$$|e| = a + b|R| \quad (3.12)$$

Parameter a determines the minimum absolute error and b the relative increase, they are determined by plotting the calculated errors against the resistance (Fig. 3.2). Outliers are removed before fitting a linear model, generally $|e| > 10\%$ of $|R|$ but the threshold is subjective. The linear model might be defined as the envelope of the measurement errors (Fig. 3.2; Slater et al., 2000). This approach is conservative as it overestimates the error for all data points, what should limit the artefacts of inversion due to noise but might yield an over-smoothed image. Alternative approaches have been developed to derive a less conservative error model balancing over and underestimation of the error and allowing to recover more details in the inversion (e.g. Koestel et al., 2008). In this work, we simply used the linear trend of the error data set as a less conservative error model.

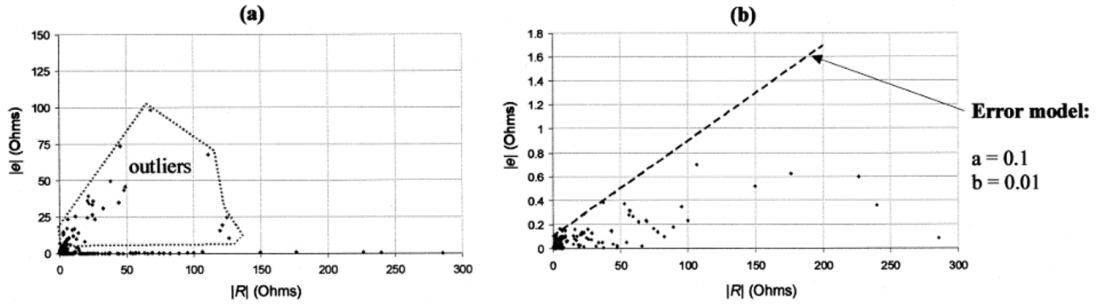


Figure 3.2: Example of determination error model: (a) all measurements, (b) after removal of outliers, the envelope defines the error model (Slater et al., 2000)

3.5 Inversion with reference model

The inverse problem is non-unique meaning that multiple models can explain the same dataset. A way to reduce the amount of mathematical solutions is to include prior information, often through the addition of a reference model (Oldenburg and Li, 1994). Prior information include all relevant information about the study site collected separately from the ERT data set. This can include geological maps, borehole data, seismic data but also previous ERT sections from the same site (Hermans, 2014).

The reference model (m_{ref}) is incorporated in the objective function as an addition to the smoothness constraint, an additional factor is also added that allows to control the influence of the reference model, i.e. the closeness factor.

$$\psi(m) = \|W_d(d - f(m))\|^2 + \lambda(\|W_m(m - m_{ref})\|^2 + \alpha\|m - m_{ref}\|^2) \quad (3.13)$$

The closeness factor α weighs the importance of the reference model during the inversion and is expressed as a diagonal matrix weighting each cell of the model (Dumont et al., 2016). If α is high the inverted model will be closer to the reference model, a low value will have a smaller influence (Caterina et al., 2014). If the last term of equation 3.13 is not added and when using

a homogeneous reference model, the spatial derivative of m_{ref} is equal to zero and the equation would be similar to equation 3.9. Thus, the third factor is added to ensure that the reference model has a influence on the inversion (Hermans, 2014).

The reference model approach can be used to impose the presence of a water layer, as illustrated by Hermans and Paepen (2020). Doing so, it is possible to artificially correct a data set collected at low tide so that it can be compared to a data set collected at high tide. In practice, the low water dataset can be corrected to the high water as follows. First the observed low water data $d_{obs, lw}$ is inverted at the low water level, resulting in a model and corresponding data $d_{calc, lw}$. On the resulting model the water layer is manually added, the data resulting from this resistivity distribution - $d_{calc, hw}$ - is calculated by running the forward model. The correction factor is obtained by calculating the difference between the two data sets: $\Delta_{calc} = d_{calc, hw} - d_{calc, lw}$. Finally to calculate the corrected dataset, this difference is applied on the original data. The described methodology is expressed by the following equation:

$$d_{obs, hw} = d_{obs, lw} + (d_{calc, hw} - d_{calc, lw}) \quad (3.14)$$

Hermans and Paepen (2020) used their proposed methodology to jointly invert land and marine ERT, on both synthetic models as on a field case the method proved accurate even when the seawater layer induces large changes. The presence of the highly conductive layer inherently causes a loss of resolution, to counterbalance this effect they included the inversion results at low water in the reference model. Here the approach will be applied to monitor during a tidal scale, when applying time lapse inversion (see section 3.6) the low water dataset is used as a background model, however the data sets need to be acquired in the same conditions so the low water dataset is corrected to high water conditions according to this methodology.

3.6 Time lapse inversion

Time lapse inversion is widely used in monitoring studies to characterize hydrogeological processes (e.g. Singha et al., 2015), here one is mainly interested in the changes in resistivity rather than the absolute values. Three main procedures exist to inverse time-lapse datasets: independent inversion, time-constrained or reference model inversion and difference inversion (Miller et al., 2008). The latter is used in this study and is hence explained below, the other two approaches used in this thesis correspond to individual inversion and time-constrained inversion of the data and are described by equations 3.9 and 3.13 respectively.

In the difference inversion scheme, the inversion problem is formulated in terms of the variations of the data and model instead of the absolute values (Kemna et al., 2002), when considering these differences equation 3.9 can be written as:

$$\psi(m) = ||W_d(d - d_0 + f(m_0) - f(m))||^2 + ||W_m(m - m_0)||^2 \quad (3.15)$$

Where d_0 and m_0 refer to the background or initial state, these results are thus used as a basis for the following inversions. As with the absolute inversion, the problem is solved iteratively until

the error weighted rms error reaches one, or when there is no improvement. Note that because of the subtraction, there is a decrease in the systematic error (LaBrecque and Yang, 2001).

Accurate reconstruction of the error model is of great importance in time-lapse studies. Lesparre et al. (2017) propose a methodology to model the noise level by using the reciprocal at each time step, their approach is proven to significantly improve the inversion results. However, in monitoring studies, there is not always sufficient time to measure the reciprocal at each time step. Nguyen et al. (2011) stated that when there is minimal difference between the noise in the different time steps, a common error model is preferred, which has been used in multiple studies and proven to be effective (e.g. Hermans et al., 2015).

In this work, reciprocal measurements were not made for each time step so here we assume that the noise level remains relatively constant throughout the monitoring campaign.

3.7 Measuring protocol

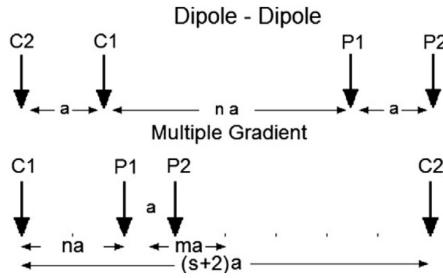


Figure 3.3: Schematic overview of the multiple gradient and dipole-dipole electrode configurations (Loke et al., 2013).

The measured electrical potential depends on the geometry of the electrode array, which is defined by the geometric factor as described in equation 3.4. Dahlin and Zhou (2004) wrote an extensive comparison of 10 electrode arrays most used in surveys based on five numerical models. They recommend the following electrode configurations: multiple gradient, pole-dipole, dipole-dipole and Schlumberger array, all four arrays produce high quality images. Multiple gradient, pole-dipole and dipole-dipole arrays are well suited for multichannel data acquisition meaning that multiple measurements can be made simultaneously, significantly reducing acquisition time. In the scope of this monitoring study

the gradient or dipole-dipole arrays are preferred because of the shorter acquisition time and comparable imaging capabilities.

With a dipole-dipole set-up, a current is injected at the current dipole, the potential difference is measured between the potential dipole, located further along the profile, both dipoles have the same electrode separation a (Fig. 3.3). To describe the geometry, the distance between the dipoles has to be known, the n value is the relative distance between the two dipoles. The geometric factor can hence be written in terms of the above parameters (Reynolds, 2011).

$$K = \pi na(n+1)(n+2) \quad (3.16)$$

The multiple gradient array is essentially a special version of the Schlumberger array (Reynolds, 2011), the potential electrodes are located in between the current electrodes as in Schlumberger but are not necessarily centered. The potential difference is measured at different

locations between the current electrodes (Fig. 3.3). In this measurement protocol a current is injected with an electrode separation of $(s + 2)a$, the potential difference is measured simultaneously at all the potential electrodes pairs with spacing a . The separation factor s represents the maximum number of measurements for a given current injection. The n -factor has the same meaning as in the dipole-dipole array, the geometric factor can then be written as follows.

$$K = 2\pi \left[\frac{1}{na} - \frac{1}{(n+1)a} - \frac{1}{(s+2-n)a} - \frac{1}{(s+1-n)a} \right]^{-1} \quad (3.17)$$

Figure 3.3 also shows the midpoint factor m and is defined as the midpoint of the potential dipole relative to midpoint of the current dipole (Dahlin and Zhou, 2006).

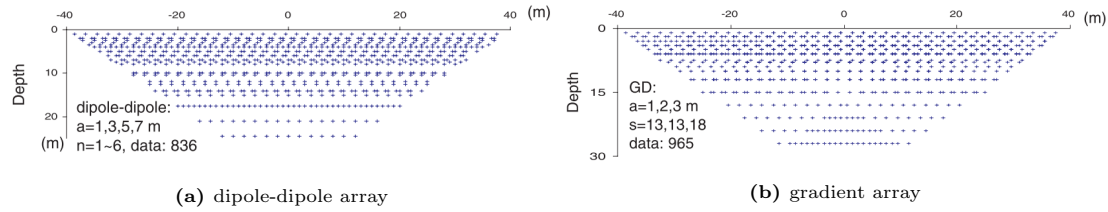


Figure 3.4: Example of data point distribution of a gradient and dipole-dipole array (Dahlin and Zhou, 2004).

A 2D section of the subsurface is obtained by measuring at different locations along a profile and systematically increasing a (Fig. 3.4). For both configurations, the data density is higher in the upper layer because more electrode combinations measure shallow points. Dahlin and Zhou (2006) propose a simple formula to plot pseudosection points for the gradient array:

$$x_{\rho_a} = x_{MN} \text{ and } z_{\rho_a} = \min\{(x_{MN} - x_A), (x_B - x_{MN})\}/3 \quad (3.18)$$

Where x_{MN} is the midpoint of the potential electrode pair, x_A and x_B are the positions of the respective electrodes.

Figure 3.5 shows the sensitivity patterns of the discussed arrays. For the gradient array, the pattern clearly depends on the position of the potential electrodes (n -value). The sensitivity function shows how a change in resistivity will influence the measured potential, thus for a positive change in resistivity, a positive potential difference will be measured in the (+) field and vice versa. When the potential pair is located close to one of the current electrodes, the pattern resembles that of a pole-dipole array, when centered it approximates a Schlumberger array (both not shown here), thus it can be expected that the gradient combines the features of these arrays. Consequently, it has a focussed lateral sensitivity which means that vertical features will be well defined with a relatively sharp vertical sensitivity which means it can also image horizontal layers. The sensitivity pattern of the dipole-dipole set-up also depends on the n -factor, when n is larger than one, it has a more focussed lateral sensitivity although not as sharp as the gradient. Also, the sensitivity is higher towards the ends of the profile but small in the centre, where the majority of data points is sampled. For comparison the Wenner array is also shown, it is clear

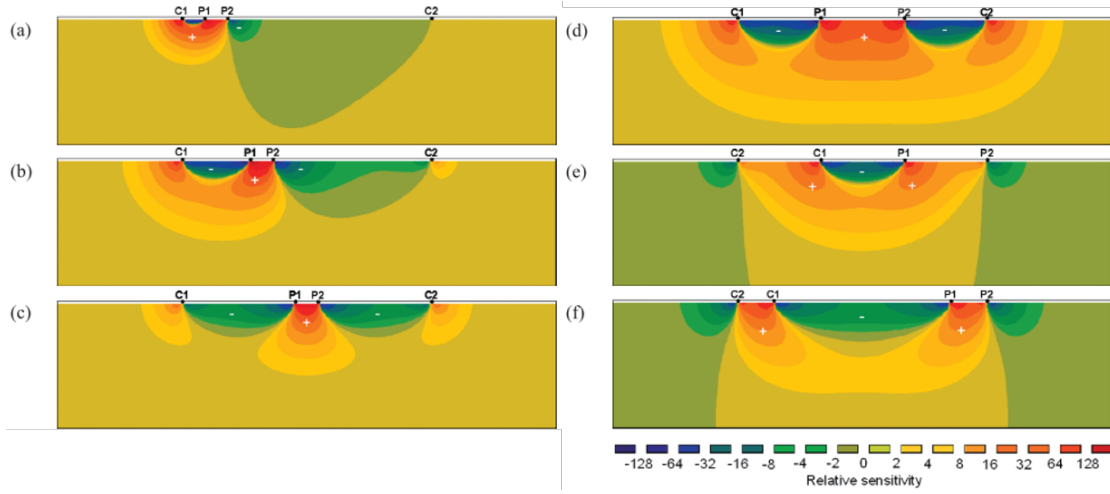


Figure 3.5: Sensitivity pattern of gradient (a-c), Wenner (d) and dipole-dipole (e-f) arrays on a homogeneous background (Dahlin and Zhou, 2006).

that this type has a less focussed lateral sensitivity but is better suited for defining structures at depth (Dahlin and Zhou, 2006).

3.8 Sensitivity matrix

A quick assessment of the image quality can be made by using the data-weighted cumulative sensitivity (Nguyen et al., 2009; Henderson et al., 2010). Kemna (2000) defines the coverage or cumulative sensitivity as:

$$S = \text{diag}\{J^T W_d^T W_d J\} \quad (3.19)$$

Where J is the Jacobian matrix. A high sensitivity implies that a change of the parameter will strongly influence the prediction while a low value will have a small to non-existent influence, hence in a low sensitivity zone is likely badly resolved. The sensitivity can thus give a rough estimate for how well the resistivity is represented by the data set. Note that the sensitivity does not coincide with the resolution, a high sensitivity does not necessarily imply a high resolution.

Other image appraisal tools such as the resolution matrix and the depth of investigation (DOI) can also be used, Caterina et al. (2013) wrote a thorough comparison on the different approaches, here the sensitivity matrix will be used because it is readily available after inversion. Other methods are more expensive and time consuming. Note that the sensitivity given after inversion is represented on a relative, logarithmic scale.

Chapter 4

Synthetic cases

4.1 Methodology

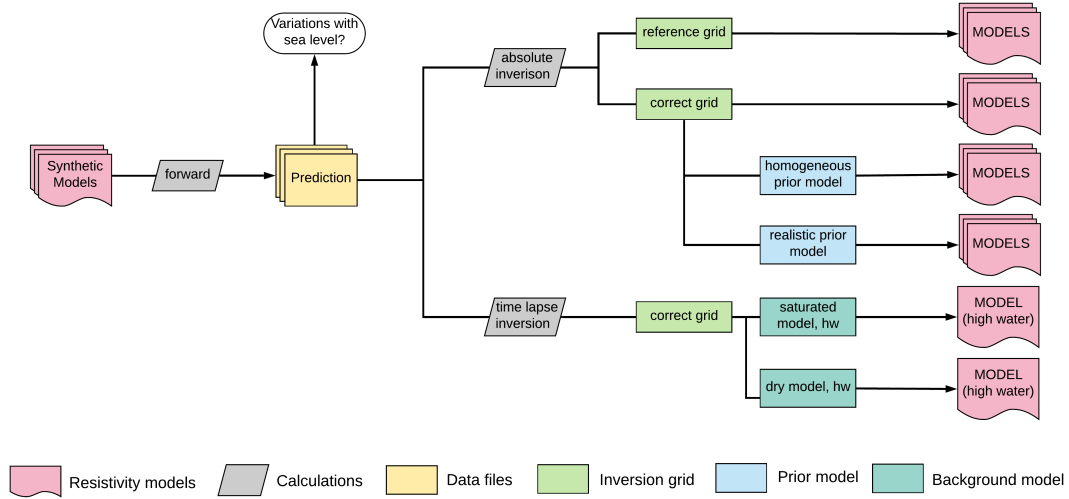
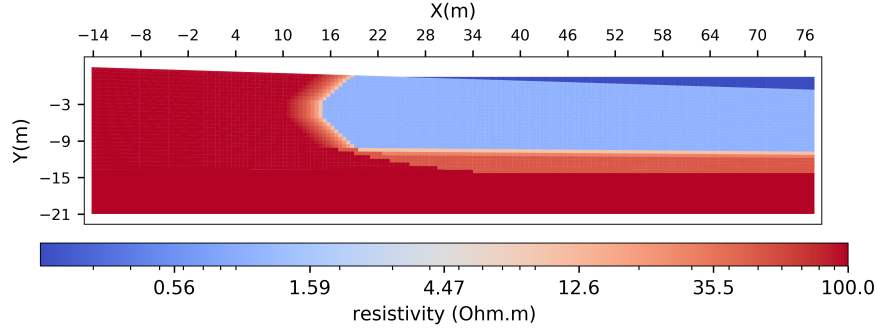


Figure 4.1: Workflow used in the synthetic study, showing the differences between inversion types.

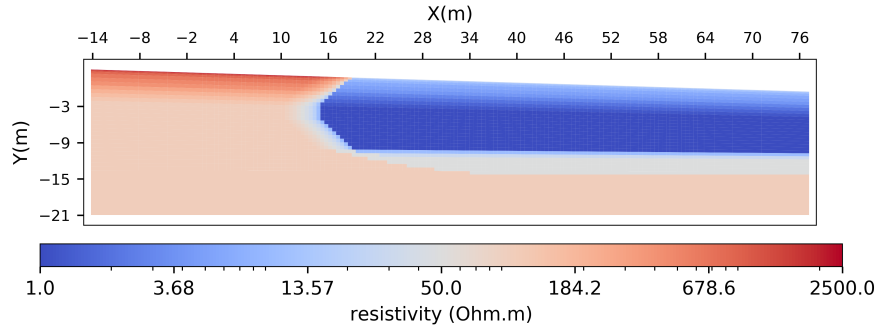
The aim of the synthetic study is to assess the influence of the water layer on the inversion of ERT data in order to know what will be possible and what the limitations are when inverting the field data. For clarity, a simplified workflow is shown in figure 4.1. Models were built with a seawater lens on top of a freshwater tongue, replicating the situation in the Westhoek (Fig. 4.2a). The lens is approximately 10 to 15 m thick (depending on the location on the profile), with the base at -10 m depth. The interface is present at 19 m at the surface and to approximately 15 m.

It will be sufficient to simulate half a tidal cycle, the sea level will be reconstructed every 0.5 m, resulting in 6 models. The models are built using a mesh of 4760 rectangular elements containing 135 rows and 36 columns, electrodes are placed at the second row of nodes so the water level can be adjusted above the electrodes. This allows to manually change the upper layer to account for the sea level and also replicates the situation on the field, where the water layer is present above the electrodes. More on that follows in chapter 5.

Thus, six models were built at six different sea levels to replicate half a tidal cycle, the potential difference resulting from this resistivity distribution is calculated by forward modelling. The data is simulated with a gradient array with $s = 8$ and $a = 1 - 6$, measuring 1392 data points. No noise is added to the data sets so here we simulate a best case scenario. Before inverting the data, the variation of the resistance with the sea level is checked to get an initial idea of the



(a) Basic model at 1.5 m water height.



(b) Model with dry upper layer.

Figure 4.2: Model used in the synthetic study the seawater lens has a resistivity of $2 \Omega.m$, the fresh water $100 \Omega.m$ and the seawater $0.2 \Omega.m$, for the dry layer a saturation of 50% is applied at the top.

influence of the rising tide. The resistance is plotted at different locations on the profile (at both ends and in the middle) and at multiple depths (six in total) by looking at different electrode spacings. For the large electrode spacings less configurations are possible and thus only the data at the ends of the profile were plotted. Note that the first electrode is located in the dunes.

The data sets are subsequently inverted using different approaches to see how it effects the reconstructed resistivity models. The grid used in the inversion is the same one used to build the models and calculate the forward model.

First, the data is inverted independently i.e. absolute inversions (Eq. 3.9) both an extreme and realistic case will be considered. The first meaning the data will be inverted without considering the influence of the water layer in the inversion, thus using the reference level grid. This is expected to illustrate the distortion caused by ignoring the water layer. Consequently the data is also inverted using the correct grid too see to what extent the water layer influences the data.

Next, a reference model is included in the inversion (Eq. 3.13), both a homogeneous and realistic reference model are considered. For the first, a background value of $100 \Omega.m$ is assigned to the subsurface and $0.2 \Omega.m$ to the water layer, corresponding to the resistivity of

respectively fresh- and seawater. It is homogeneous in the sense that the subsurface gets assigned one value for the resistivity. The model shown in figure 4.2a is used as the realistic reference model, adapted to the correct water height. When using reference models, a higher importance can be given to the water layer during inversion by assigning a higher weight to it, in both types a weight of 0.001 is given to the subsurface and 1 to the water layer. Taking into consideration prior knowledge will provide a more realistic solution. For instance, in time-lapse studies, the resistivity models of previous time steps can be used as a reference for subsequent inversion. This can help to overcome the loss of resolution induced by the presence of the seawater layer (Hermans and Paepen, 2020).

At last, a time lapse inversion will also be made using the difference inversion scheme (Eq. 3.15). The high water dataset is inverted by considering the differences with the low water dataset. This method requires the data to be acquired in the same conditions hence the low water dataset has to be corrected to high water, using the methodology described in section 3.5.

Next to the influence of the seawater, the effect of drying in the upper layer induced by the variation of the hydraulic head during the tidal cycle, will also be simulated. Drying will lower the saturation, from Archie's law (Eq. 3.2) we know that this leads to an increase in the bulk resistivity. The model at reference level has been adapted to simulate a dry layer (Fig. 4.2b). A saturation of 50 % was applied at the top, which was gradually lowered over the upper 11 layers, this results in a fairly thick, distinct dry layer which was deliberate to ensure the dry layer had an effect on the data. At high water, we assume that the model is saturated and thus display a lower resistivity.

The differences between profiles and inversion types can be calculated using the following equation.

$$\Delta\rho(\%) = \frac{\rho_x - \rho_y}{\rho_x} * 100 \quad (4.1)$$

With ρ_x and ρ_y the resistivity of profiles x and y , this can be used to quantitatively interpret the inversion results. Given that x is the background, with the used colour scale, red shows a decrease in resistivity in profile y and blue an increase. Also, in all the representations, a cut-off at 100 % (positive and negative) was applied, this is sufficient to show the main changes and also allows to see the small discrepancies. Values between +10 % and -10% are considered artefacts.

When comparing different inversion types, it can also be interesting to look at the model misfit, this shows how close the inversions approximate the real model.

$$M_{misfit} = 100 * \sqrt{\frac{\sum_{i=1}^N \left(\frac{\rho_{calc,i} - \rho_{true,i}}{\rho_{true,i}} \right)^2}{N}} \quad (4.2)$$

For the interpretation, three water quality classes will be distinguished, salt water ($\rho < 2\Omega.m$), fresh water ($\rho > 100\Omega.m$) and brackish water containing the intermediate resistivity values, these are the same classes used to build the model.

4.2 Resistance variations

4.2.1 Results

At all six levels, the influence of the rising tide can be seen as a decrease in resistance as expected from the presence of the conductive water layer above the electrodes (Fig. 4.3, Appx. A). The high conductivity of seawater leads to a low resistivity or resistance, which are proportional (eq. 3.1). Although the general trend is a decrease at certain points an increase can be seen, this phenomenon appears at all levels. The curve is generally smoother on the sea side of the profile compared to the dunes (see Appx. A), and has a quicker response. The remaining figures, the dune side and centre (where available) are given in appendix A. In the dunes, the absolute resistance values are much higher compared to the other locations - which is to be expected from fresh water - but the relative changes are also higher, situating between 80 and 100 %.

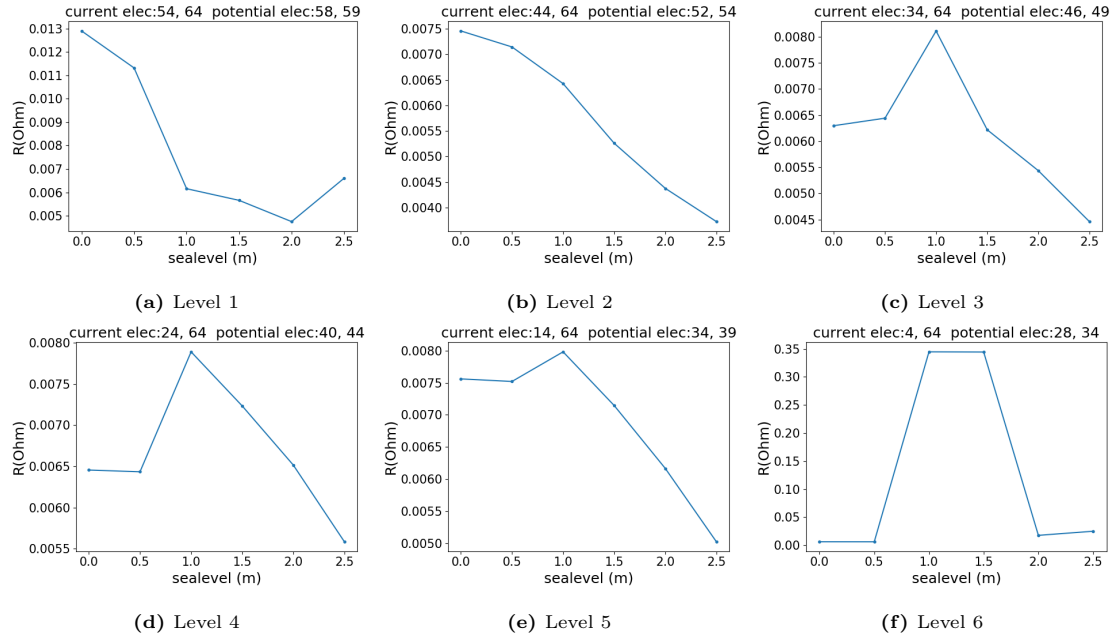


Figure 4.3: Graphs showing the resistance variation at the six levels, on the sea side of the profile.

4.2.2 Discussion and conclusion

As the generally observed trend is a decrease in resistance, the rising sea level clearly has an influence on the observed data, and is not limited to the data configurations investigating the shallow subsurface. The unexpected decrease in resistance at certain points can be ascribed to the sensitivity pattern discussed in section 3.7. In the negative zones, a decrease in the resistivity distribution - as expected here because of the water layer - will result in an increase in

the measured resistance. Thus the addition of the water layer yields different results for different electrode configurations. Given this dependency, identifying resistivity changes from the data alone will be almost impossible and inversion is hence necessary to more thoroughly investigate the effect. Note that here the only factor influencing the data is the salinity of the water layer, when other variations are present, for example a change in subsurface salinity or saturation, this would result in a more complicated trend.

4.3 Absolute inversions

4.3.1 Results

4.3.1.1 Inversion at low water level

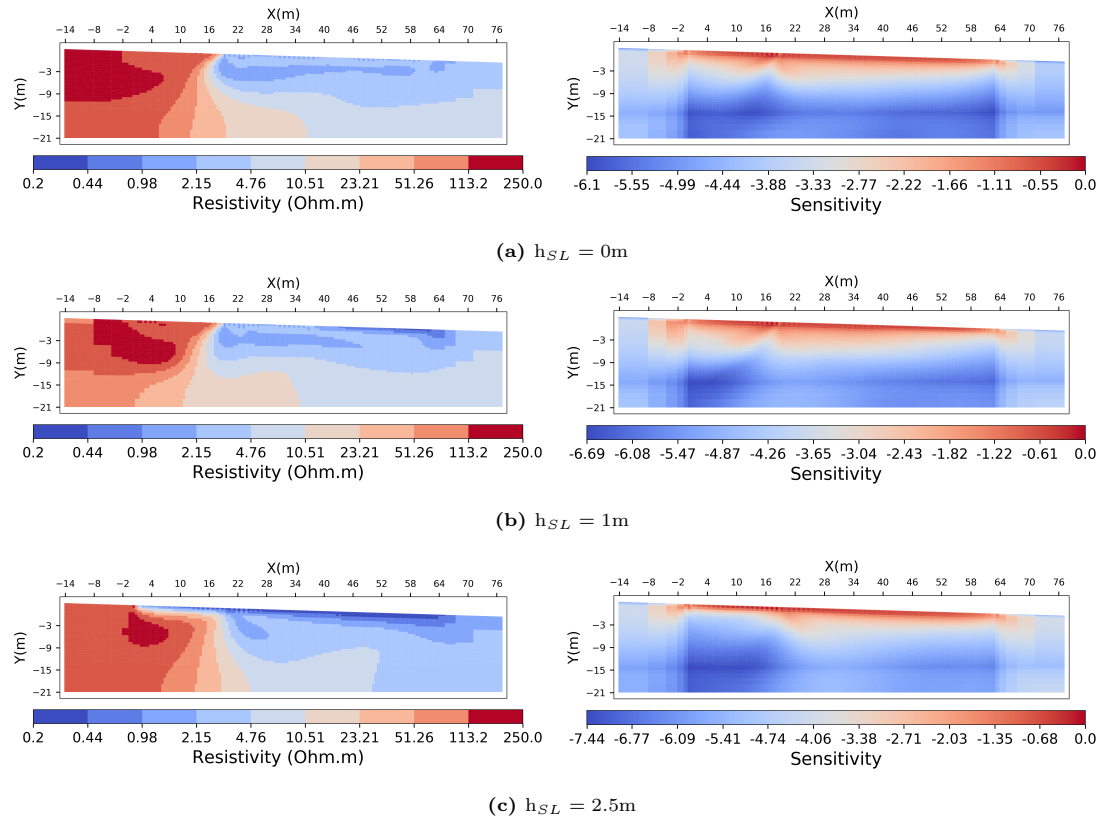


Figure 4.4: Results of inversion using the reference grid with the corresponding sensitivity.

In this section, the data is inverted using the reference level grid, hence not taking the water layer into consideration in the inversion. Note that the models used are obtained by calculating the forward model at different sea levels, so the water layer was present when calculating the

potential difference. A selection of the inversion results are shown in figure 4.4, the two extremes and an intermediate case, the remainder of the figures are given in appendix B.

At the low water level (Fig. 4.4a), the saltwater lens looks similar to the original model, the interface is also present at 19 m at the surface but has a slightly smaller horizontal extent (approximately 1 m less). The base of the lens is irregular but at roughly the same location as in the original model. The freshwater tongue could not be recovered, under the lens there is brackish water present which is likely related to the loss of resolution with depth of ERT. When looking at the sensitivity one can see that it is highest in the upper layer of the model and then rapidly declines with depth, thus the structures at depth have to be interpreted with care, i.e. the vertical extent of the lens and the presence of brackish water underneath. Note that the sensitivity is lower in the vicinity of the interface, this phenomenon is also present in the other profiles, including the ones not shown. As said before the sensitivity is represented on a logarithmic scale, thus the highest value is 0 ($\log_{10}1$), the negative value, for example -7, actually corresponds to 10^{-7} .

At one meter water height (Fig. 4.4b) the seawater lens is also nicely delineated laterally, the interface and extent are the same as in the 0 meter model. The vertical extent of the lens is also roughly the same, with an irregular base at approximately -9 m, a notable difference is the shape of the interface. In the model and inversion with low water level the front has an oval shape and continues as a more or less straight line. Here the front has a more sigmoidal shape instead of a straight line. This unusual shape is present in a zone with very low sensitivity and could hence be an artefact of the inversion. Note that the resistivity is lower in the upper layer ($0.2 \Omega m$) corresponding to the resistivity of seawater, which is only present at the right side of the profile, between 43 and 65 m. Since the water layer is not taken into account during inversion, the inversion algorithm has to mirror it below the surface, yielding lower resistivity than expected.

The results at 2.5 m water height differ significantly from the previous discussed ones (Fig. 4.4c). The presence of the water layer has a significant effect on the data, to accurately describe the situation, the salt water will be split in two types: the salt water lens which is always present and the seawater. The upper layer of the model has a low resistivity which is clearly the result of the water layer, it extends toward 0 m on the profile. This only encompasses the first few rows of the model, and is maximum 1 m thick, a sharp, horizontal transition zone between the seawater and the freshwater aquifer is present. A bit further down on the profile the interface dips downward at approximately 20 m on the profile, coinciding with the location of the seawater lens. The shape of the lens cannot be distinguished on this profile. The sensitivity pattern of this inversion differs from the others, in the upper layer and on the left side it is similar as the other profiles. However, when looking at the area between -2 and 20 m one can see that the sensitivity in the first 5 meters is significantly lower. This coincides with the region where the front of the lens is located. This decrease of sensitivity is due to the recovered low resistivity zone resulting from the water layer. The current flows preferentially in low resistivity zones, reducing

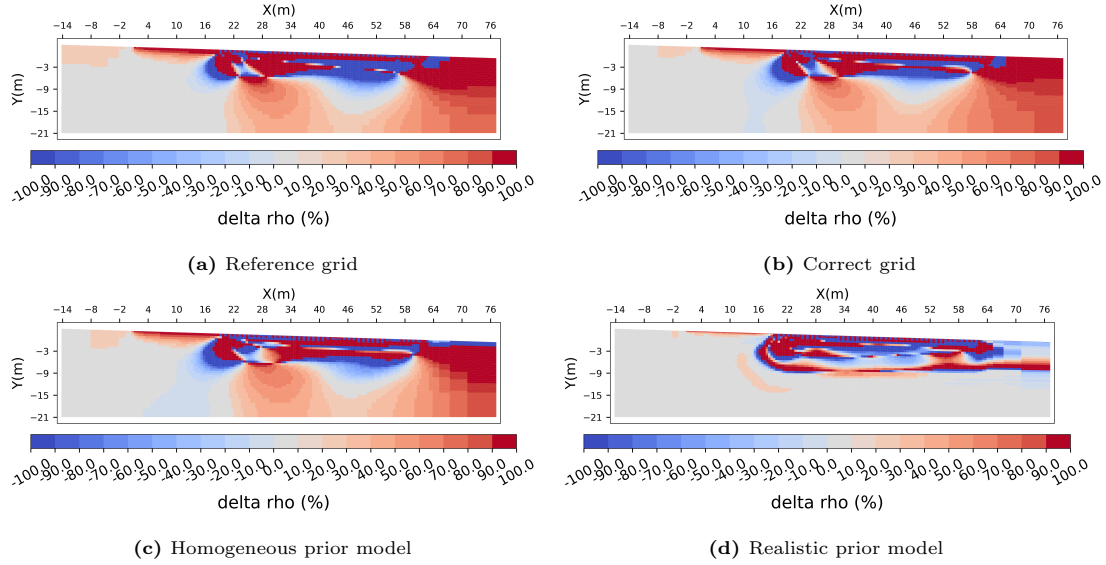


Figure 4.5: Differences between the profile at high and low water for the different inversion types. In all cases the low water profile constitutes the background

the sensitivity of deeper layers.

The remaining results are shortly discussed since they are not shown here (see Appx. B) and have similar characteristics. In terms of geometry, the model at 0.5 meter look roughly the same as the 0 meter model, the only notable difference is the presence of low resistivity in the upper layer resulting from the seawater. The geometry at 1.5 is similar with the only difference that the vertical delineation is not clear, the interface does not continue vertically but dips downward around 46 meter, which is likely an artefact of the inversion. The same can be observed in the 2 meter model, in this model the presence of the water layer extending the lens can be clearly seen, as with the 2.5 model. However, the geometry of the lens still distinguishable. In the higher water models, the low resistivity in the upper layer is more defined with rising sea level. The sensitivity is intermediate between the high water model and the others, it is lower in the area between -2 and 20 m but not as distinct. The unique shape of the lens in the 1 meter model is not present in the other inversions, which is why it was shown here.

The difference between the high and low water profile was calculated with equation 4.1, using the low water profile as the background x and substituting the high water profile in the y -position (Fig. 4.5a), this is also the case for the following inversion types. A red colour indicates a lower resistivity in the high water profile and a blue colour shows an increase. Between 0 and 20 m, near the surface a thin red layer is present coinciding with the water layer surpassing the seawater lens, on top of the lens it is intercalated with blue zones. Since the only difference between the two models is the water layer, they should be identical for the subsurface part with a variation of 0%, meaning that the observed variations are a result of the inversion. In the dunes there are minimal changes, at depth there is a general but small decrease of resistivity. The most

important thing on this figure is the oval shaped blue zone in the centre of the profile, coinciding with the location of the seawater lens, with a red zone roughly in the centre. The general shape of the red zone coincides with the irregular shape of the lens in the high water profile. what indicates that recovering the saltwater lens through inversion is particularly affected by the water level, making monitoring very challenging.

4.3.1.2 Inversion with correct water level

The results of the inversion using the correct grid are shown in appendix B and figure 4.6, note that the differences in the grid used here compared to the previous inversion type are subtle. The upper layer is now horizontal in the zone representing the seawater, i.e. on the right side of the profile the distance depending on the sea level. The lower resistivity values are concentrated in the layer representing the sea water, nonetheless, the smooth inversion tends to enlarge the influence of the conductive layer to underlying rows.

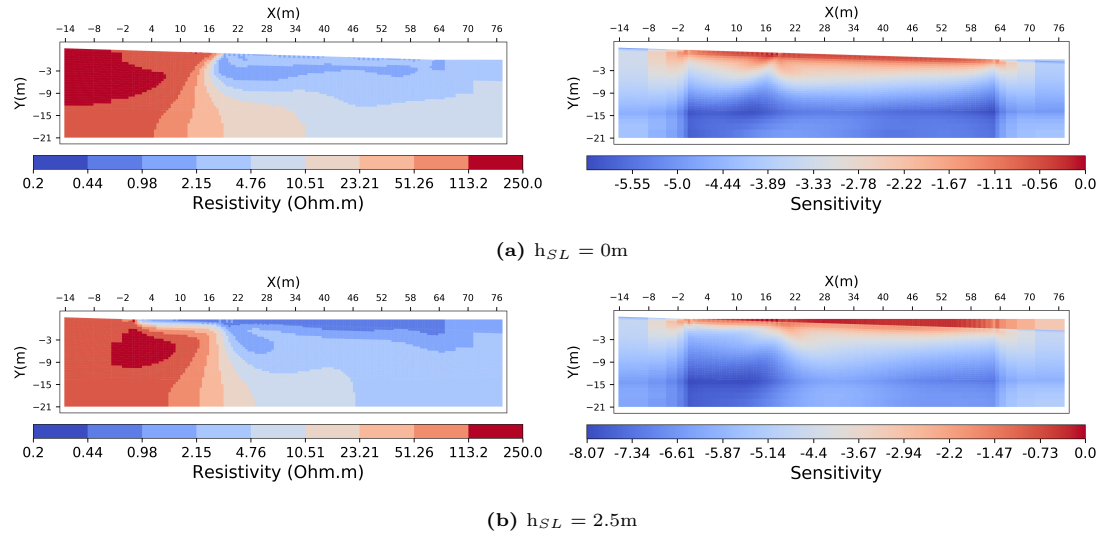


Figure 4.6: Results of the inversion using the correct grid and corresponding sensitivity.

In terms of geometry, the results look similar to the previous inversion type, the interface is consistently present at approximately 18 m for the profiles with sea levels between 0 and 2 m. At 0 and 0.5 m, the lens is well delineated at the bottom with brackish water underneath which is an artefact of the inversion. In the profiles between 1 and 2 meter, the lens is not well delineated vertically, low resistivity values are recovered at depth instead of the high resistivity expected in this zone. This phenomenon was also present in the previous inversion type but not at the 1 m model. For the 2.5 m model, the interface is present at the expected location but the geometry of the lens is completely lost, as was the case for the previous inversion type. The low resistivity in the upper layer resulting from the sea water is present in all the models and becomes more

distinct as the sea level rises, especially when it passes the interface. Nevertheless, the smooth inversion tends to enlarge the lower resistivity of the seawater layer to the subsurface sediments, resulting in an underestimation of the resistivity at shallow depths, although to a lesser extent compared to the previous case.

The sensitivity is practically the same as when inverting the data with the reference grid, the only notable difference is the presence of the water layer, which has a very high sensitivity. The difference between the high and low water profiles (Fig. 4.5b) is also very similar to the previous inversion type with the difference that the red zone in the seawater lens is slightly more prominent between 40 and 48 m.

4.3.1.3 Inversion with homogenous reference model

When using a homogeneous reference model, explicitly integrating the seawater layer, the results look similar to the previous ones (Fig. 4.7, Appx. B), the biggest difference is that the seawater lens is better delineated when using this approach, with exception for the high water profile, as for the previous cases the geometry could not be recovered. The area under the lens again consists of lower resistivity due to the resolution loss at depth. The sensitivity of the subsurface is the same as for the other inversions, the water layer has a notably higher sensitivity. Including the known resistivity of the seawater as prior information helps to better delineate the resistivity distribution; although smoothing across the sea/sediment interface still occur, especially when the thickness of the seawater layer is limited (e.g., 0 to 15 m on figure 4.7b). The differences between the high and low water profile are again similar to the previous ones, with - again - a more prominent red zone in the oval-shaped feature (Fig. 4.5c).

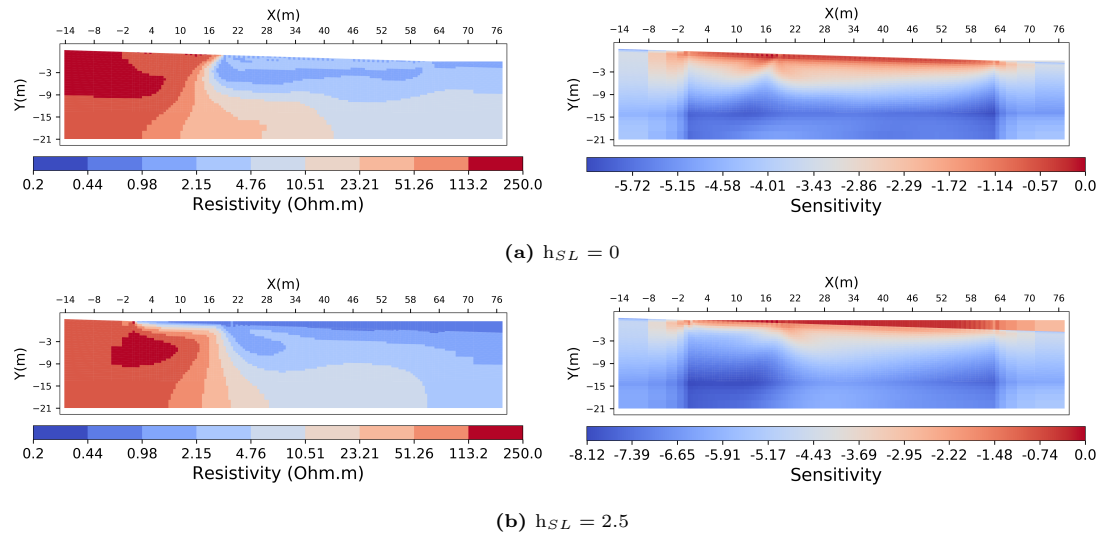
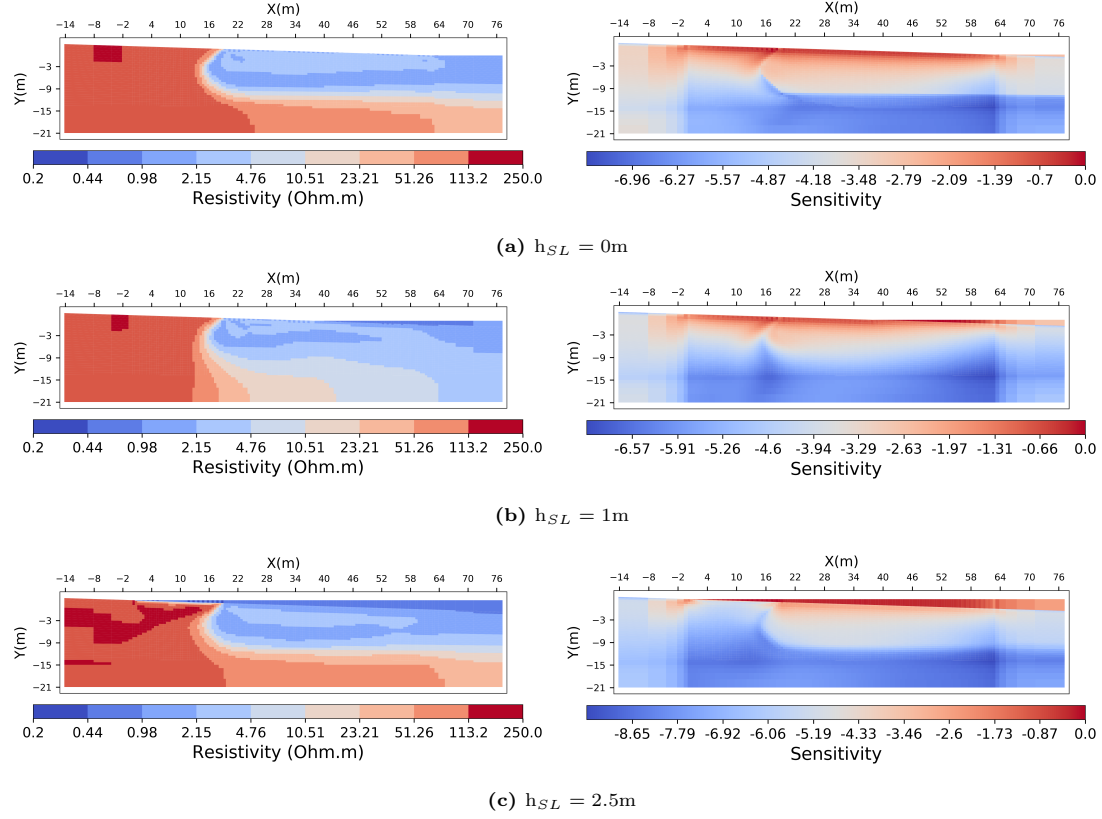


Figure 4.7: Results of the inversion with a homogeneous reference model and corresponding sensitivity.

4.3.1.4 Inversion with realistic reference model

**Figure 4.8:** Results of the inversion with a realistic reference model and corresponding sensitivity.

In a time-lapse study, an earlier measurement can be used as a background model for the subsequent inversion, i.e. a realistic reference model introducing some time constraint in the inversion. Here we explore an absolute inversion where the original model is given as a-priori information (i.e. a best case scenario for which the actual model would be known) and it has a significant effect on the results (Fig. 4.8, Appx. B). In profiles with the sea level at 0, 1.5 and 2.5 m the solution lies very close to the reference model, the lens is very sharply delineated, with minimal artefacts underneath. When looking at the sensitivity the seawater lens can be clearly distinguished with a high values in the lens itself and low values surrounding it. During the inversion of these profiles, few iterations were made to reach the minimal rms error i.e. 2 iterations for $h = 0$ and 1.5 m and 3 iterations for $h = 2.5$ m, this means that the background model is already very close to the solution, which might not be realistic in field cases. Note that even for the high water model, the seawater lens could be recovered, which was not possible in the other approaches. Not all the profiles show this reference model over-fitting, at 0.5, 1 and 2 m water height, the lens is nicely delineated but not to the same extent as the other profiles, the

interface is consistently present at 18 m, as expected. The sensitivity in the lens is fairly high although not as high as in the over-fitted models. Here the difference between the high and low water profile differ from the previous types with overall smaller variations, a subtle decrease is present in the dunes (between 0 and 20 m) and in the mixing zone, following the shape of the lens. The majority of the differences are present in the lens - the morphology of which can be clearly distinguished - but the variations are irregular and there is no real trend or pattern present (Fig. 4.5d). This means that even in a best case scenario (no noise and true reference model) one can expect significant artefact of inversions related to the presence of the water layer. The artefacts would likely be more limited if instead of the true model, the inversion result at the reference level is used as the reference model.

4.3.2 Discussion

The section above describes the different inversion types and the effect it has on the data, here they will be compared to see how the different approaches effect the data.

In most of the models, the sensitivity is roughly the same i.e. high in the upper layer, until approximately -9 m depth, rapidly declining after this threshold, and also lower at the lateral edges of the grid. Note however that the sensitivity decreases when the seawater layer is present, making the resistivity distribution more difficult to recover. When interpreting the results it has to be taken into account that the resistivity above -9 m will be well represented while the area underneath is less reliable. Exceptions are: 1) the models at high water, a zone with low sensitivity is present directly under the upper layer between -2 and 18 m, overlapping with the zone where the front of the lens is present. 2) the inversions made with a realistic reference model, here the sensitivity follows the pattern of the seawater lens for 3 out of 6 profiles.

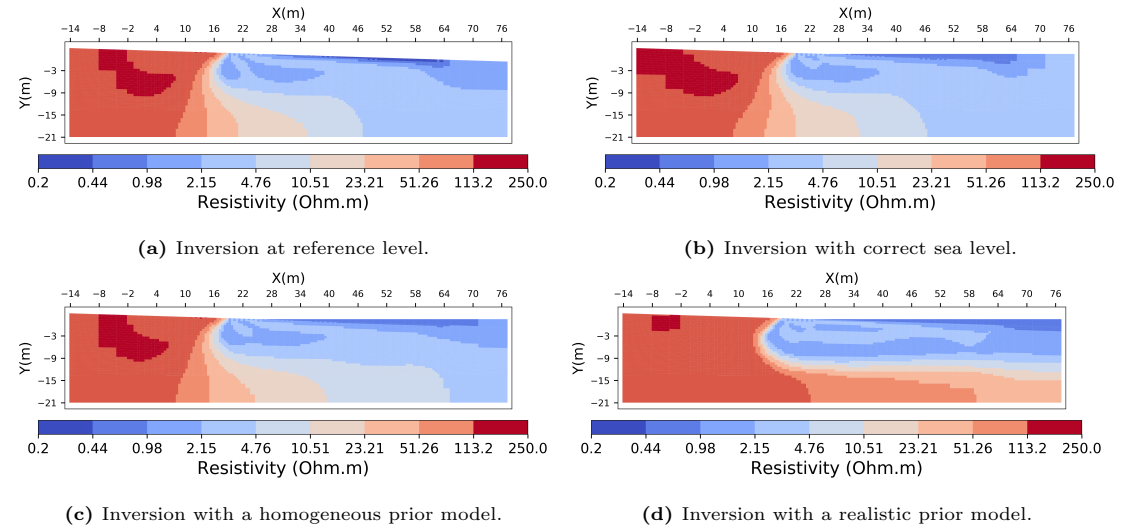


Figure 4.9: Comparison of the four inversion types for the model at 1.5 meter

In terms of morphology we can be brief, figure 4.9 shows the same data inverted in four ways to highlight the differences. The first major thing that stands out is the inversion with reference model, the saltwater lens here is very well defined but this is because the increase of resistivity below the saltwater lens is included in the prior model. This highlights the importance of the reference model in the inversion. In practice, such a model could be obtained from independent prior information or from a profile collected at low tide with higher sensitivity. However, it must be noted that including incorrect prior information in the reference model would yield to a worse solution (Caterina et al., 2014). This option must thus be chosen with care. Note that this phenomenon is not present in all profiles, with water height 0.5, 1 and 2 m this is not the case.

The remaining inversion types do not show this over-fitting. For the models between 0 and 2 meter, the interface is present at 18 m on the surface and the front reaches to 16 m, the lens is slightly shorter than in the original model (± 1 m), thus, laterally the lens is well reconstructed by all three techniques.

At depth, lower resistivity values are recovered due to loss of resolution with depth for ERT measurements, this is more prominent with higher sea levels. Brackish to saline water is recovered here instead of the expected fresh water. The interface should continue vertically, for all approaches this is the case in the first few meters, after this the interface dips straight down. For the inversions without reference model, this happens at approximately 46 m on the profile so the first 30 m of the lens are well reconstructed. When inverting with reference model the interface continues further and dips only at 64 m, meaning that the lens is well reconstructed in the first 48 meter. These values are specific to the 1.5 meter model, shown in figure 4.9 but the same principle applies for the other models. Based on morphology, the inversion with reference model is preferred to the regular inversions because it more accurately defines the seawater lens.

For the high water models the situation is different, in all approaches, the geometry of the lens is nearly completely lost, the only preserved characteristic is the location of the interface at the surface. The front of the lens retreats losing the typical oval shape and the vertical delineation is off-track, again due to the loss of resolution, which is here clearly much more significant. On this profile, the seawater has a negative influence on the quality of the inversion, however with the 2 meter model this is not the case. Here the sea level is also fairly high and surpasses the location of the interface. The most likely explanation is that at high sea levels, thus a thick layer, a significant fraction of the current is transmitted in the water layer, making it harder to accurately reconstruct the sea water lens.

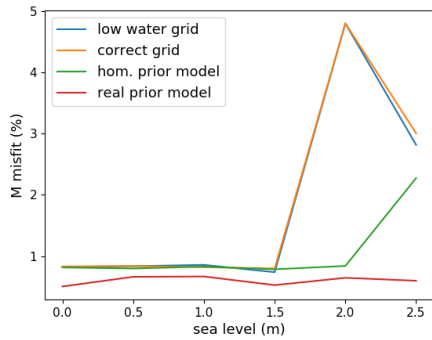
The distortion the water layer causes becomes very clear when looking at the differences between the profiles, with exception from the realistic reference models, all the inversion types show the same general trend. The artefacts at the location of the seawater lens can be linked to the irregular shape of the lens at high water. Note that here two independent inversions are considered.

All the results shown above are represented on a unique colour scale ranging from 0.2

Table 4.1: Range of resistivity values for the different types of inversions, all values are expressed in $\Omega.m$

Sea level (m)	reference level		correct level		hom. pr. mod.		real pr. mod.	
	min.	max.	min.	max.	min.	max.	min.	max.
0	0.52	139.85	0.52	145.47	0.46	161.49	0.38	122.34
0.5	0.23	140.99	0.36	143.13	0.32	122.66	0.30	123.55
1	0.17	282.01	0.30	125.87	0.28	123.13	0.26	121.78
1.5	0.12	126.14	0.28	128.98	0.26	124.55	0.21	120.83
2	0.13	259.92	0.27	258.90	0.25	273.72	0.18	182.22
2.5	0.09	186.38	0.23	162.72	0.23	162.13	0.18	173.61

$\Omega.m$ to 250 $\Omega.m$, however these are not the real resistivity ranges of the inversion results. However, this information can also be interesting and is given in table 4.1. The first thing that stands out is the decreasing trend in the minimum resistivity, for inversion types, it decreases with increasing sea level. For the inversion using the reference sea level, the sea water resistivity of 0.2 $\Omega.m$ is quickly recovered - the 0.5 m model has a resistivity of 0.23 $\Omega.m$ and the 1 m model 0.17 $\Omega.m$ - however, this minimum resistivity continues lowering until 0.9 $\Omega.m$. For the other types, the minimum resistivity steadily drops until it approximates 0.2 $\Omega.m$ at the high water level (0.23 for the correct sea level and homogeneous prior model and 0.18 for the real prior model). Thus, all the approaches were able to recover the low resistivity but more accurately at high tide. As for the maximum resistivity, no clear trend is visible with the rising sea level, the inversion at reference level does seem to recover slightly higher resistivity, probably because it over-compensates for the low resistivity. Further, it is noteworthy that in all cases, the maximum resistivity substantially exceeds the 100 $\Omega.m$ mark.


Figure 4.10: Model misfit of the different inversion types.

Another tool to compare the inversion results is by looking at the model misfit (Fig. 4.10), described in equation 4.2. For the absolute inversions without reference model, i.e. with the reference and correct grid the misfit follows the same trend. It remains constant at approximately 0.8 % until 1.5 m water height, after there is a steep increase to 4.7 % followed by a drop to approximately 3 % at the high water level. The increase takes place when the sea level passes the interface, noteworthy is that the misfit for the 2.5 m model is smaller than the 2 m model, but the geometry of the

lens was completely lost. When using the homogeneous reference model the misfit remains con-

stant at roughly the same level as the previous, followed by an increase to 2.3 % at 2.5 m. Lastly for the real reference model the misfit fluctuates slightly but remains under the 1 % mark.

4.3.3 Conclusion

The quality of the different methods can be assessed in three ways: the visual aspect, the resistivity range and the model misfit, by combining the three the best approach can be determined.

Purely based on the visual aspect, a conclusive preference can not be made. Note that in this specific case, when using a realistic reference model the results are biased because the reference model used is too close to the actual solution, what is not possible for field cases. However, the approach could be used when using a less informative reference model, for example, instead of using the original model one could use the inversion results at the reference level as a prior model for the subsequent profiles, due to time constraints this approach is not tested here.

When considering the resistivity range, the models using the correct grid yield a better approximation of the low resistivity values than when using the reference grid. Solely based on these arguments making the choice whether or not to use a prior model are not evident, when using it the lens is slightly better delineated but the differences are subtle. However, when taking into account the model misfit it is clear that using a prior model yields better results. Where the misfit shows a steep increase at 2 m without a prior model, it remains constant when using it and when the increase does occur (at 2.5 m) it is smaller. This improvement is mostly related to the incorporation of the water layer during the inversion, what helps identifying the sea/sediment interface.

Thus, to revise, the preferred inversion type for this specific dataset is the one using a homogeneous prior model with a background value of $100 \Omega.m$ for the subsurface - which was chosen to resemble fresh water in order to recover the bottom of the seawater lens - and $0.2 \Omega.m$ for the water layer accompanied by a higher weight to ensure that the inverted model displays the known seawater layer. Alternatively, if more information is known about the subsurface, the homogeneous reference model can be refined to further improve the solution. This information can be brought by independent data (e.g. boreholes) or in time-lapse studies by a reference profile collected at low tide when the sensitivity is maximum. However, one should avoid incorporate uncertain information, especially in low sensitivity zones (e.g. Caterina et al., 2014; Hermans, 2014).

Regardless of the inversion type, the water layer generates significant artefacts up to more than 100%. Since the seawater lens has a high conductivity, the absolute change is not as high, nonetheless, this could easily hide changes of subsurface salinity or saturation. It is important to limit the distortions caused by the water layer as much as possible, note that when using the realistic prior model, the effect seems less problematic, thus time-constrained or time-lapse inversion could improve the artefacts caused by the water layer, which is explored in section 3.6.

4.4 Effect of drying during low tide

4.4.1 Results

The idea here is to assess the effect of the water layer, superimposed with a change of subsurface resistivity, i.e. drying in the upper layer increasing the resistivity at low tide. The inversions here are made with a homogeneous reference model as it was best suited for this specific dataset.

The inversion results are given in figure 4.11. The sensitivity is high in the upper 10 m of the model, with lower values on the edges and in the vicinity of the interface, similar as the model without a dry upper layer. The upper 5 meters of the model have an increased resistivity, which is clear above the lens and in the freshwater section, in the transition zone however, it is less distinguished. The thickness of the dry layer is well recovered but the highest resistivity values are not recovered, likely due to the smoothing effect of the inversion. In the original model the maximum resistivity value is $2500 \Omega.m$ whilst in the inversion result is only $1454.07 \Omega.m$. The resistivity values in the centre of the lens are lower compared to the inversion at high water, with the dry layer it is within the $1.44 - 3.88 \Omega.m$ range, at high water between 0.54 and $1.44 \Omega.m$.

The differences between the two profiles are calculated using equation 4.1, the dry model is substituted in the x position and the model at 2.5 m in the y position (Fig. 4.11). The influence of the dry layer is evident in the salt water lens, a thick red layer of approximately 6 meters is present indicating the resistivity in the dry profile is higher. In the fresh water zone, the effect is much smaller. In the zones between -14 and 12 m there is no change in resistivity and again between 45 and 60 meter. In between the two zones there is a slight change in resistivity, corresponding to the mixing zone, the resistivity is again higher in the dry profile. Immediately under the dry layer, there is a blue zone with a shape similar to the salt water lens, intercalated with red zones.

4.4.2 Discussion and conclusion

The inversion could not recover the high resistivity values of $2500 \Omega.m$ because of the smoothing effect, note that this only includes a few cells in the first layer of the model (in the freshwater part). Nevertheless, the inversion clearly showed the presence of a layer with increased resistivity along most of the model. At the mixing zone, the dry layer is not clearly distinguishable, the mixing zone already contains a high degree of variability in a narrow zone, when considering the dry layer on top of that it is likely that the inversion could not completely reconstruct this complex situation. From the difference image, it is clear that the variation of saturation in the upper layer is of a higher order of magnitude compared to the artefacts observed in the previous section, so that it can be qualitatively imaged. Nevertheless, some artefacts of inversion remain in the deeper layers. Thus, when a layer with low saturation is present, this is reflected in the inversion to some extent. When interpreting field data sets, a variation in the upper layer

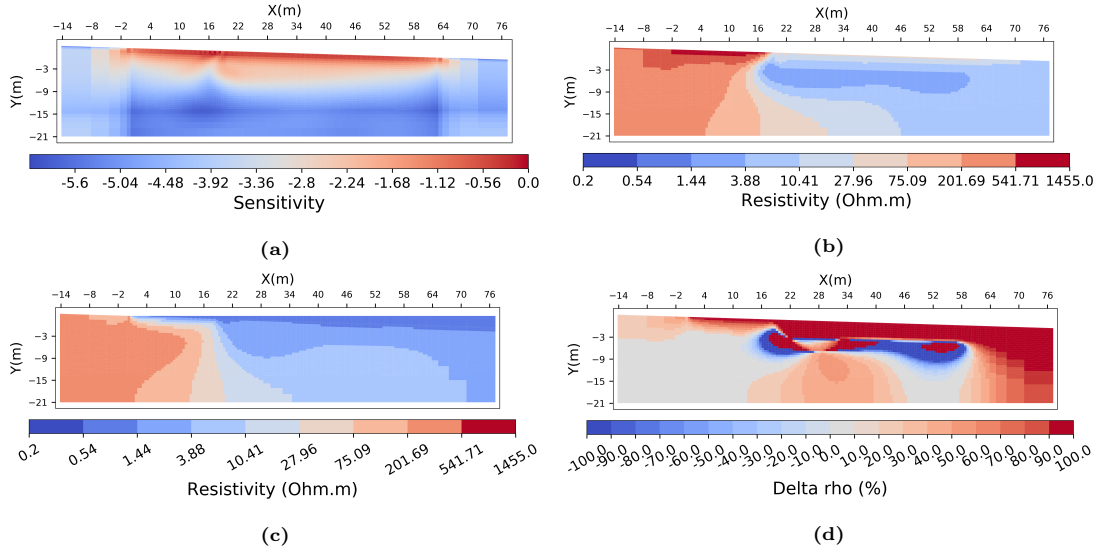


Figure 4.11: (a & b) Result of the inversion with dry upper layer, with the corresponding resistivity. (c) The inversion at high water, for comparison, using the same colour scale. (d) The resistivity difference between the dry and high water models.

could be due to drying so this has to be taken into account. Note that the dry layer in this model is deliberately made thick to see which effect it could have on the data, this is not necessarily the situation on the field.

The figure showing the difference with a high water profile shows some interesting characteristics. First the red layer, representing the higher resistivity at low water becomes thicker towards the right edge of the profile, this is not due to the drying effect. After 64 m there is no data present - this area is the extension of the grid with larger cells to minimize the effect of boundary conditions - and the sensitivity is also much lower, the same applies for the area in front of 0 m. The artefacts present at the locations of the salt water lens are similar to what is observed in the saturated models, i.e, caused by distortion due to the presence of the water layer. The layer with low saturation is superimposed above the effect of the water layer. As stated in the previous section, executing a time-lapse inversion might limit this effect and is explored in the following section.

4.5 Time lapse inversion

4.5.1 Dry model

4.5.1.1 Results

The previous results have shown that the water layer could induce some artefacts in the recovered change of resistivity when independent inversions are used. To overcome this issue, a time-lapse

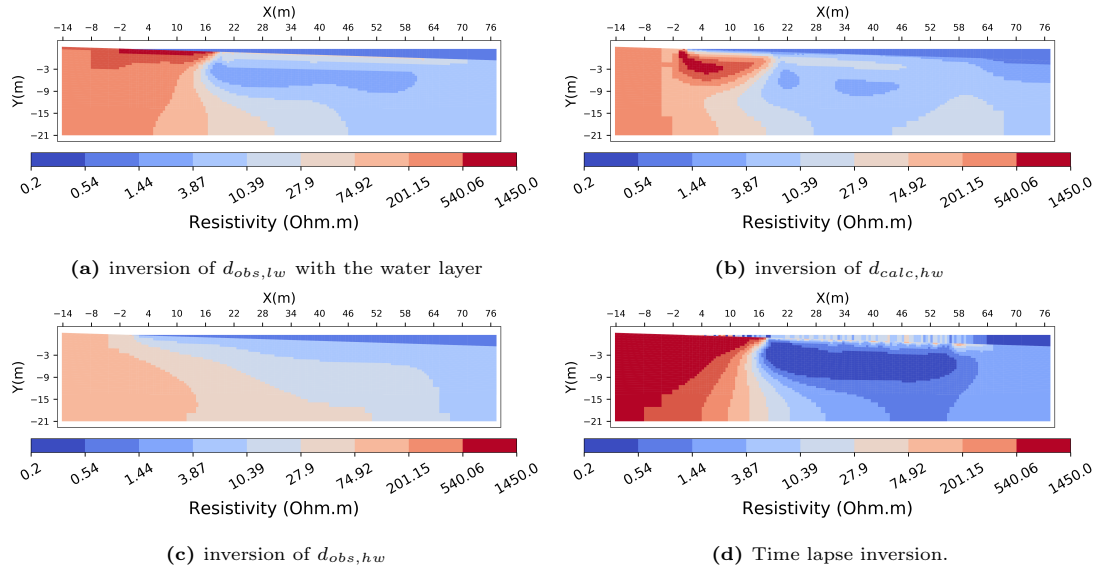


Figure 4.12: Result of the data correction and time lapse inversion for the dry model.

difference inversion is used including a correction for the water layer.

To illustrate the workflow used to acquire the data corrected to high water level, some intermediate results are shown. First, a water layer is manually added to the inversion result at low water level (Fig. 4.12a). Next the corresponding data is calculated by running the forward model, figure 4.12b shows the inversion result of this dataset, note that the geometry of the seawater lens is distorted due to the presence of the water layer. The difference is calculated between the calculated data at high and low water and is subsequently applied to the original - observed - data, giving the low water model corrected to a high water level, figure 4.12c. The result of the inversion of the corrected data set has no remnants of the seawater lens, or the dry layer, illustrating a strong loss of resolution due to the presence of water layer. It looks like a smooth transition from a high resistivity value of $47.04 \Omega.m$ to the low resistivity of the seawater. The quality of the estimated high water data and correction is shown in figure 4.13, from both figures it is clear that the correction is not satisfying, the correlation between the estimated and true data is relatively low, with a lot of spreading, especially for the estimated data where the resistance is severely overestimated for most of the data points. Clearly the correction is less efficient as for the case studied by Hermans and Paepen (2020). Note that with low measured resistances as is the case here, the relative error can quickly be relatively large.

Nevertheless, the corrected data set has been inverted as the background data set in the time-lapse difference scheme. The time lapse inversion was able to recover the basic structures of the model, a simple seawater lens can be seen but the vertical limit is not well defined. At the top of the lens, higher resistivity values are present, that likely correspond to the dry layer. The figures are represented on a unique colour scale, however, the result from the time lapse

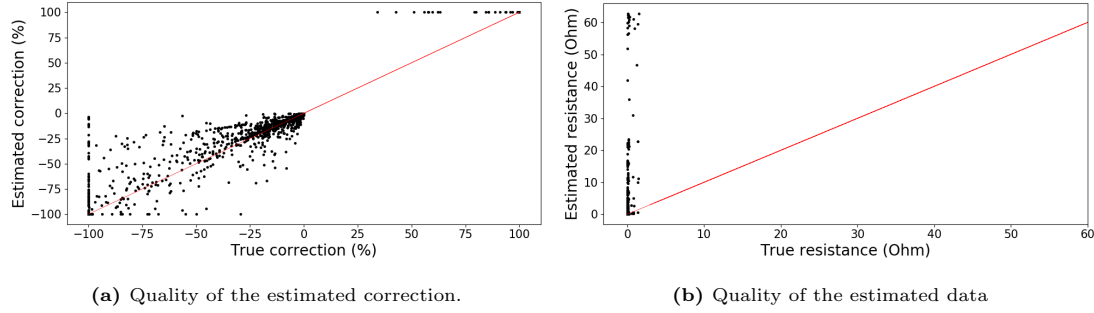


Figure 4.13: Scatterplots representing the quality of the correction and the resulting data, calculated on the dry model.

inversion spans over a large range of resistivity values i.e. from 0 to 407380.03 $\Omega.m$. Lastly, the inversion was finished in 7 iterations, with an rms error of 106.2 % at the end.

4.5.1.2 Discussion and conclusion

We can say with confidence that the correction does not yield a good result when applied on the dry model, the scatterplots do not show sufficiently good correlation between the estimation and true data. The unsatisfying correction is confirmed by the inversion result, giving a smooth image with absolutely no remnants of the original structures. Using the correction as a starting model was expected to improve the results and be less prone to artefacts compared to the independent inversions. Nonetheless, the time lapse inversion failed to converge to a solution (rms of 102.06 %) but some rudimentary structures can still be distinguished, such as a basic sea water lens and remnants of the dry layer. The latter is not realistic since the inverted high water data set is completely saturated. The resistivity range of the inversion is also not realistic, a resistivity of 0 $\Omega.m$ is impossible, this would mean the medium is a perfect conductor and the maximum resistivity is extremely high, this in combination with the very high rms error leads to conclude that the approach is not suited to invert this specific data set.

It seems that the time-lapse inversion scheme fails to converge towards a realistic solution. The difference inversion is supposed to find the model with the smallest possible change compared to the background. Here, it seems that the inversion is mostly modifying the seawater layer, probably because it is the most sensitive part of the profile and fails to identify the changes occurring in the subsurface. This observations requires more research that falls out of the scope of this master thesis.

4.5.2 Saturated model

4.5.2.1 Results

The approach where the dry model is used as the background in a time-lapse inversion is unsatisfactory and the dry layer has a clear influence on the results, the same method was applied on the regular model, i.e, fully saturated, which is explored here.

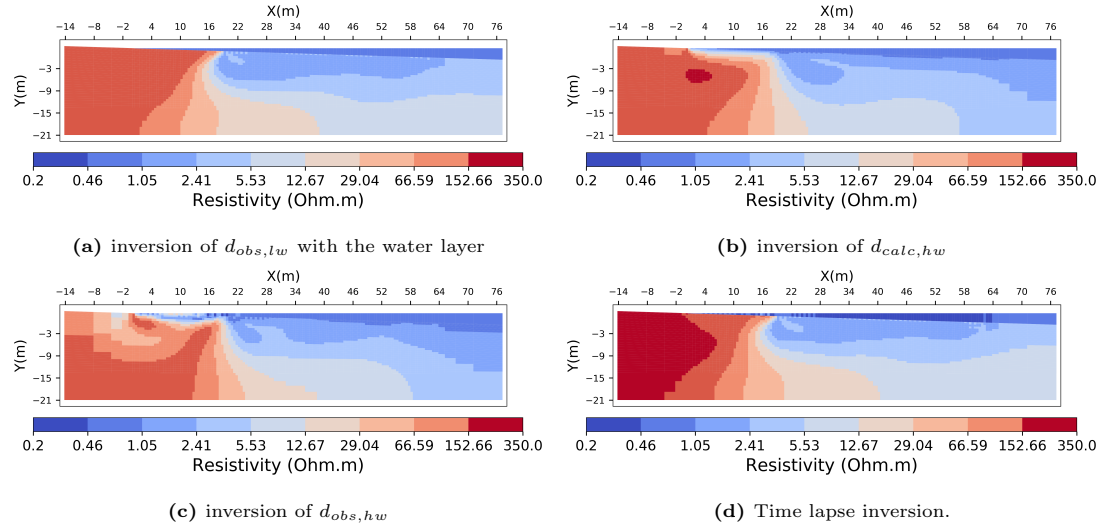


Figure 4.14: Result of the data correction and time lapse inversion for the dry model.

The results of data correction and time lapse inversion are shown in figure 4.14. The inversion of $d_{calc,hw}$ is similar to the absolute inversion of the high water data. The inversion of the corrected data - $d_{obs,hw}$ - nicely delineates the salt water lens with a thick mixing zone but the resistivity under the dunes is lower than expected. This leads to believe that the correction and high water data are relatively well estimated, which is confirmed when looking at the respective scatter plots (Fig. 4.15). The correction plot shows a good correlation with only a few outliers, the resistance plot shows a correlation although less distinct as the first, the data points deviate more from the trend line when the resistance increases. This confirms that the presence of a dry layer at the water-sediment interface is likely the reason for the worse correlation observed in the previous section.

The time lapse inversion was able to clearly reconstruct the geometry of the lens, both horizontal and vertically, the resistivity values lie between 0.086 and 355.39 $\Omega.m$. The inversion was finished in 3 iterations with an rms error of 14.60 %.

4.5.2.2 Discussion and conclusion

Based on a visual analysis, the data correction looks satisfying, this is confirmed with the quality analysis using scatter plots. There is minimal discrepancy between the true and estimated

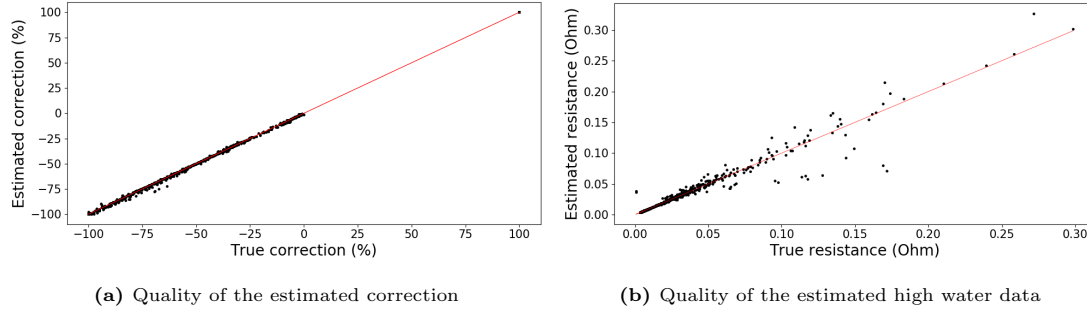


Figure 4.15: Scatterplots representing the quality of the correction and resulting data, calculated on the saturated model.

correction, with the exception of a few outliers. As for the high water data, the correlation is less satisfying, for the low resistance values the points plot nicely on the red line but as the resistance increases, outliers become more abundant. This is reflected in the corrected model (Fig. 4.14c), at the left side, close to the dunes, the resistivity is much lower than what can be expected.

The result of the time lapse inversion clearly delineates the seawater lens, which was not possible with the absolute inversion where the lens had a very irregular geometry. The resistivity values lie within a realistic range and is in accordance with the results from the absolute inversions. Based on the visual aspect, the time lapse inversion yields a satisfactory result for this type of data. However, the inversion did not converge to a solution, the rms error of 14.60 is still too high, as it should be 1. Hermans and Paepen (2020) observed that when applying a correction, the global error is likely to increase due to the lower resistance contained by the data. As for the dry model, the high rms might be related to the fact that modifications are mainly focussed on the seawater layer, to overcome this problem an additional constraint could be developed to make sure no changes occur in the seawater layer and the process can hence focus on the subsurface. However, this is out of the scope of this thesis.

4.5.3 Comparison

When calculating the correction on the model with a dry layer and using this in the time lapse inversion the results are extremely bad, there is only a very low correlation between the true and calculated data. When manually adding the water layer on top of the dry model, an unrealistic situation is created where a thick water layer is present but the sediment underneath is not fully saturated. This violates the conditions for which the correction is supposed to be correct, i.e. a slight change in the model as here a preferential flow path is created on top of the dry zone (Hermans and Paepen, 2020). Remnants of the dry layer are still visible in the final result leading to believe it negatively influences the data correction. When the same procedure was applied on a saturated model, the results are much more satisfying. The data correction is much better estimated, which becomes clear when comparing the scatter plots, for the saturated model

there are but a few outliers while in the dry model there was only a limited correlation present. In their study (Hermans and Paepen, 2020) also calculated the quality of the correction, on a similar synthetic case and the results are comparable to what is seen here, with the saturated model. They did not add a dry layer and thus calculated the correction on a saturated, low water model. Note that here, the quality of the high water data contains much more outliers than in their case. From this it can be concluded that the method of data correction is well suited for model where the upper sediment layer is conductive but not when a significantly more resistant layer is present

Although the results of the saturated model are much more satisfying, and it looks like the time lapse inversion yields a realistic results, the rms error still remains to high, thus further investigation is needed to resolve this problem for example the development of a constraint as explained above, but this is out of the scope of this thesis.

Chapter 5

Field case

5.1 Methodology

5.1.1 Field measurements

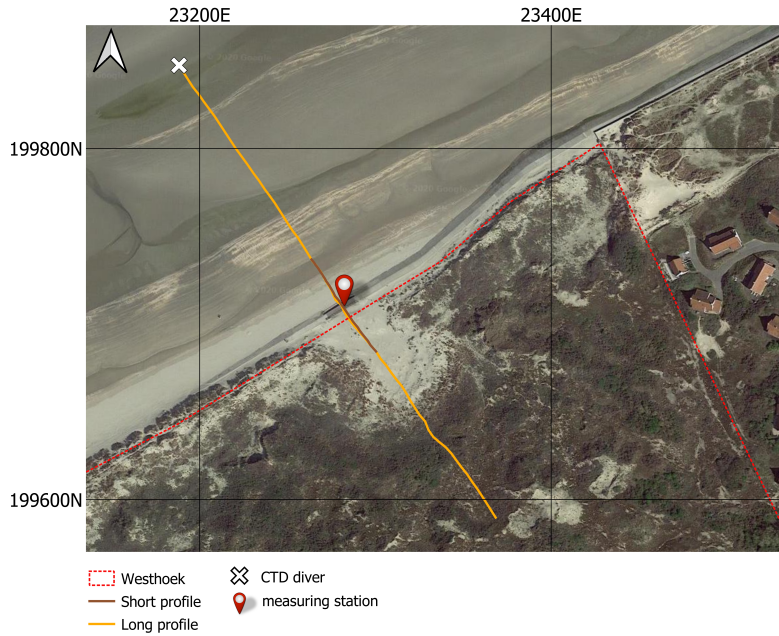


Figure 5.1: Location of the profiles.

Two profiles in the Westhoek were monitored by ERT during the first week of September 2019 (Fig. 5.1). Both are located along the same transect and are centered on the dyke, corresponding to the high water line. Both profiles consist of 64 electrodes, one profile with 5 meter and one with 1 meter electrode spacing. The long profile is used to detect deeper variations whilst the short profile provides a higher resolution at the high tide limit.

For the sections of the profiles in the dunes, the electrodes are metal rods hammered in the ground and connected to an electrical cable (Fig. 5.2a). To limit the contact resistance and efficiently inject electrical current, seawater was poured on the electrodes in the dunes. On the beach the cables are present in a dynamic environment, a set-up as used in the dunes could probably not withstand the influence of the waves so a marine or borehole cable was used instead. Such a cable has electrodes sealed within it allowing it to be used under water or underground without additional connection so there is no need to place metal rods in the ground. To ensure the cable remained secure, it was buried in a shallow trench of approximately 20 to 30 cm deep

(Fig. 5.2b).

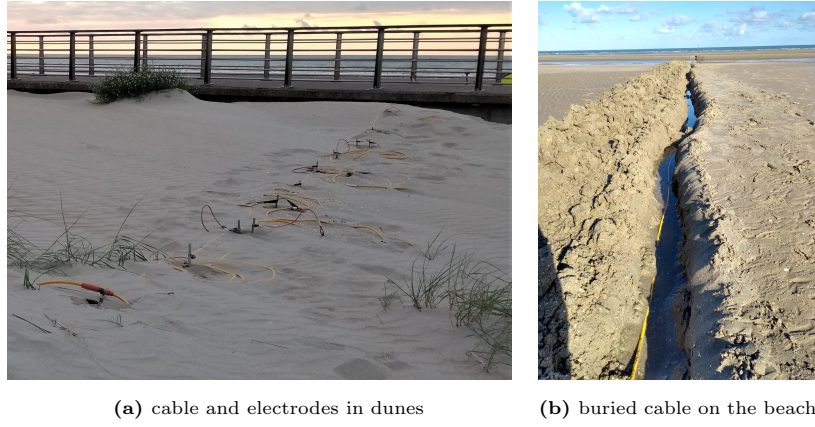


Figure 5.2: Examples of the different types of cables used in the experiment.



Figure 5.3: Field test set-up and equipment.

The ABEM Terremeter was used to monitor resistivity on the site, this is a state-of-the-art acquisition system that can measure resistivity, self potential and induced polarization (ABEM Instrument AB, 2012). The equipment was set up under the bridge to protect it from rain (Fig. 5.3a). During the resistivity measurements, CTD data was simultaneously gathered to know the exact time the electrodes are inundated. The diver was located near the first electrode of the long profile, as indicated in figure 5.1 and to ensure stability, the device was secured to a concrete panel (Fig. 5.3b). The model used during the campaign is the DI 272 manufactured by Eijkelpamp (Eijkelpamp: Soil and Water, 2016), the diver measures conductivity, temperature and pressure every eight seconds.

Setting up the equipment was a time consuming process, the trench had to be dug by hand at low tide and refilled before high tide and electrodes had to be placed in densely vegetated dunes. As a result the first day was used to set up and test the equipment, including testing both measuring protocols. On the field, the results of these test were compared to determine the

protocol best suited for the study site and objectives.

Table 5.1: Repeatability error and measuring time of the to proposed arrays.

	Gradient	Dipole-dipole
Var (%)	1.90	5.69
Duration (h)	0:52	1:02

The repeatability error (later discussed in more detail) is significantly higher for the dipole-dipole profile and there was also a significant amount of non-physical negative resistivity, thus, the gradient array was used for the remainder of the survey. Note that the measuring time of the gradient is 10 minutes shorter than the dipole-dipole which makes it

more suitable for monitoring in a dynamic environment. The inversion scheme assumes constant conditions during measurements, any possible variations will induce an additional error during inversion, thus a shorter acquisition time is desired when possible.

The monitoring campaign started on 3 September at 6:50 and ended on 5 September at 12:00, during this time 38 normal profiles were measured (long and short) and 3 reciprocals using a gradient array with $s = 7$ and $a = 1 - 7$, measuring 1372 data points. Sea level measurements from the closest station (Nieuwpoort) are also shown on the plot, the measurements are made every five minutes and given in mTAW, four full tidal cycles were recorded during the survey (Fig. 5.4).

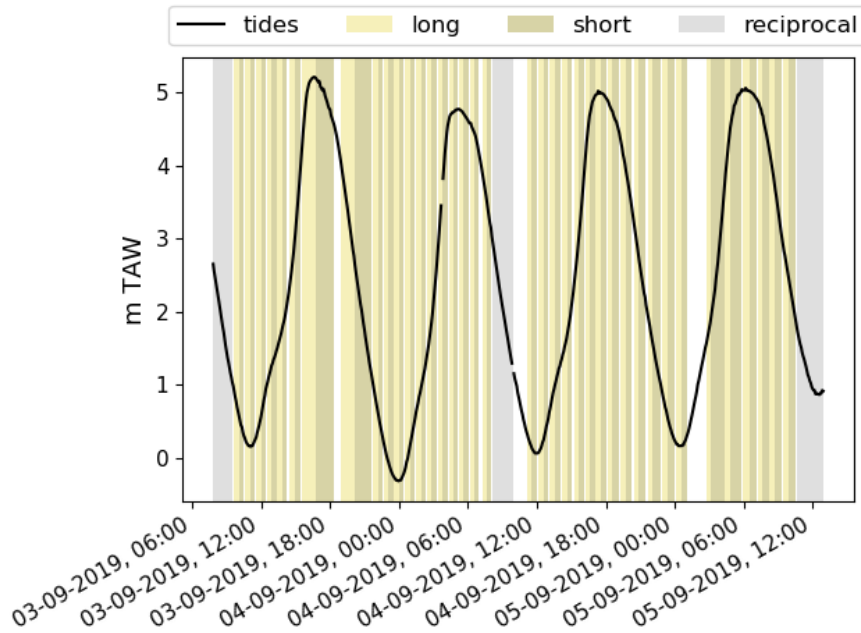


Figure 5.4: Measuring intervals and sea level measurements from Nieuwpoort. The first electrode of the long profile has an elevation of 2.559 mTAW and for the short profile 4.488 mTAW.

5.1.2 Data processing

The sea level data measured in Nieuwpoort will likely differ slightly from the situation in De Panne, since the cities are located approximately 13 km apart. To ensure the correct sea level is used, the CTD data will be plotted and compared to the measurements from Nieuwpoort. The conductivity is expected to give the exact time at which the sea reaches the first electrode, from the pressure the water height above the diver can be derived. The temperature log is expected to be less informative.

Before the monitoring data sets are inverted, the data quality is evaluated to ensure the profiles are viable for inversion. This can be done by looking at the variation coefficient. During measurements data points are stacked, the variation coefficient is equal to the standard deviation between stackings divided by the mean resistance for the data point. The same data point is measured repeatedly until either the variation coefficient reaches a predetermined minimum or the maximum number of stackings is reached. In this case, we used a minimum of 2 stacks and a maximum of three stacks, with a threshold of 1%. The low number of stacks is justified by the desire to acquire data as fast as possible given the dynamics of the tide cycle. The threshold is therefore used to filter data points of low quality. The variation is a relative value, the absolute variation is obtained by multiplying with the resistance.

As with the simulations, the variation in resistance during a tidal cycle is plotted on certain part on the profile and at multiple levels. This will give a first impression on to which effect the tide has on the subsurface resistance.

The last step before the data can be inverted is to assess the noise level of the data, as described in section 3.4 there are two ways to approach this, either by reconstructing the envelope of the remaining errors or by using the trend line. The two approaches will be tested and compared to determine which is best suited to invert this specific data set.

5.1.3 Inversions

The grids used for the inversions are shown in figure 5.5, apart from the elevations and the scale the grids are the same. They are build of 4760 rectangular elements, containing 35 rows and 136 columns, similar as the grid used for the synthetic study. As with the grid used in the synthetic study, the cells also get larger towards the edges and the electrodes are placed on the second row of nodes, allowing to accommodate for the changing tide with the first row of nodes. The first row of elements has a thickness of 10 cm for the long and 5 cm for the short profile, below the thickness is equal to half the electrode spacing, thus 2.5 m for the long and 0.5 m for the short profile and there are two cells between subsequent electrodes. Note that the first layer is much thinner than the remaining ones, although it may not seem like it, in figure 5.5 the electrodes are located on the second layer.

Three types of inversions are carried out: first an absolute inversion with homogeneous reference model, it is homogeneous in the sense that the subsurface gets one background value for

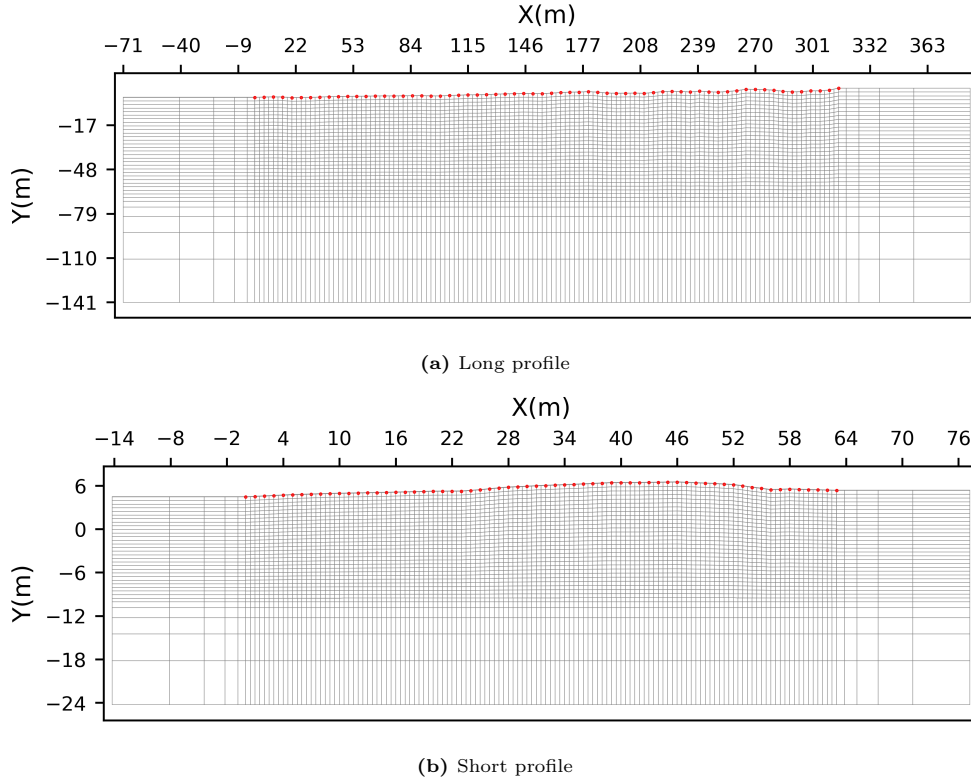


Figure 5.5: Grids used for the inversions, the electrodes are indicated by the red dots.

the resistivity however, the water layer get a different background value and weight, as with the synthetic case. The data is also inverted with a realistic reference model, the profile measured at the start of the tidal cycle is used to invert all subsequent profiles, an approach situating between the absolute and time lapse inversion. Given the unsatisfactory results obtained with the synthetic case, it was decided not to pursue with the time-lapse inversion. Note that for the visualization, the colour scale is limited to of $4500 \Omega m$.

Interpretation of the data is done semi-quantitatively based on relative resistivity variations, three classes are distinguished: fresh water ($\rho > 100 \Omega m$), salt water ($\rho < 5 \Omega m$) and brackish water, containing the intermediate resistivities. Note that high resistivity values ($\rho > 300 \Omega m$) correspond to unsaturated sediments. The differences between profiles are calculated using equation 4.1. The formula is not only applied to asses the difference between high and low water but also to calculate the difference between different low water profiles, when the electrodes are not covered with water, the upper layer of the beach will dry during ebb which will also influence the resistivity as shown in the synthetic model. The earliest measured profile will always be profile x and the latter profile y , a lower resistivity in profile y will result in a positive difference, represented by red, a higher resistivity in y will result in a negative difference and a blue value. As with the synthetic model, a cut-off at 100% was applied and values between

-10 % and +10 % are considered to small to be interpreted.

5.2 Pre-processing

5.2.1 CTD data

Figure 5.6 shows the data of the CTD diver plotted with the measurements from Nieuwpoort as the background. Contrary to what was expected, there is no correlation between the conductivity data and the tidal fluctuations (Fig. 5.6a), the electrodes of the diver likely remained wet when the beach was not fully saturated. The temperature log (Fig. 5.6b) does have a correlation with the tides, temperature at the diver increases at high tide and drops during low tide. This means that seawater temperature was higher than the temperature surrounding the diver. Since the summer of 2019 was extremely warm and the diver was buried in the sand, this seems plausible.

The pressure measured above the diver can be used to estimate the height of the water column above the device, these calculations are made automatic but they assume a water density of 1 kg/m^3 (Eijkelkamp: Soil and Water, 2016). Assuming a density of 1.025 kg/m^3 for sea water, a correction factor of 0.9756 was applied to account for the higher density. The z coordinate in the vicinity of the diver is known so the water height could be reconstructed in mTAW (Fig. 5.6c). The tide rises 20 minutes earlier in De Panne, which was expected. Further, the sea level is higher although this difference is not constant. The flat parts do not represent stable sea levels but means that the equipment was not inundated at the time.

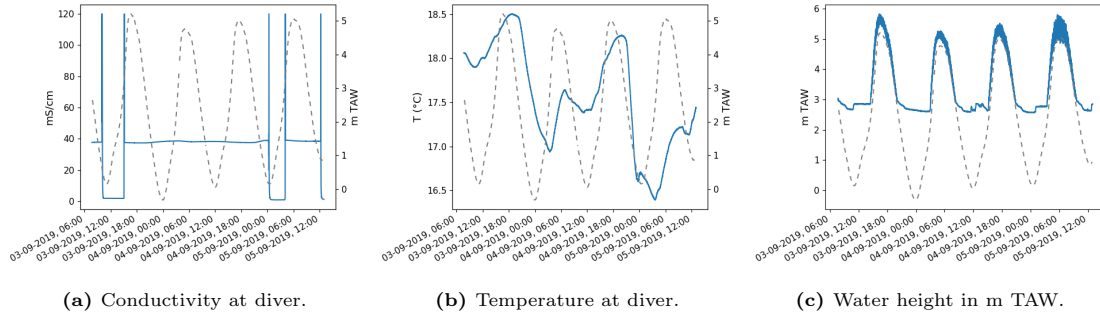


Figure 5.6: Data of the CTD diver, with the sea level measurements of Nieuwpoort as background.

5.2.2 Data quality

The variation coefficient (Fig. 5.7a) increases after the second tidal cycle, for the short profile the increase is significant, for the long profile the increase is more subtle but present nonetheless. The absolute variation (Fig. 5.7b) shows a similar trend, there is a significant increase in variation for the short profile starting around the end of the second cycle. For the long profiles however, the variation remains relatively constant, with a slight increase towards the end.

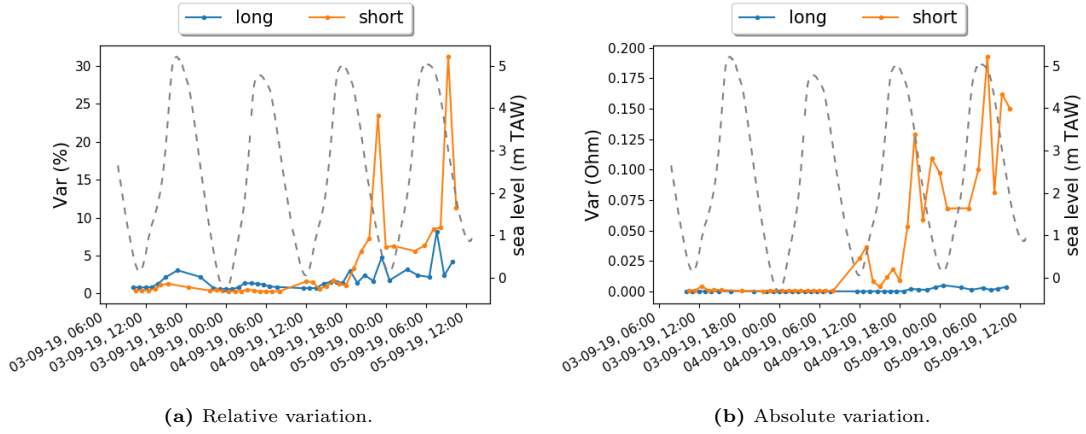


Figure 5.7: Mean variation at data points for all the profiles.

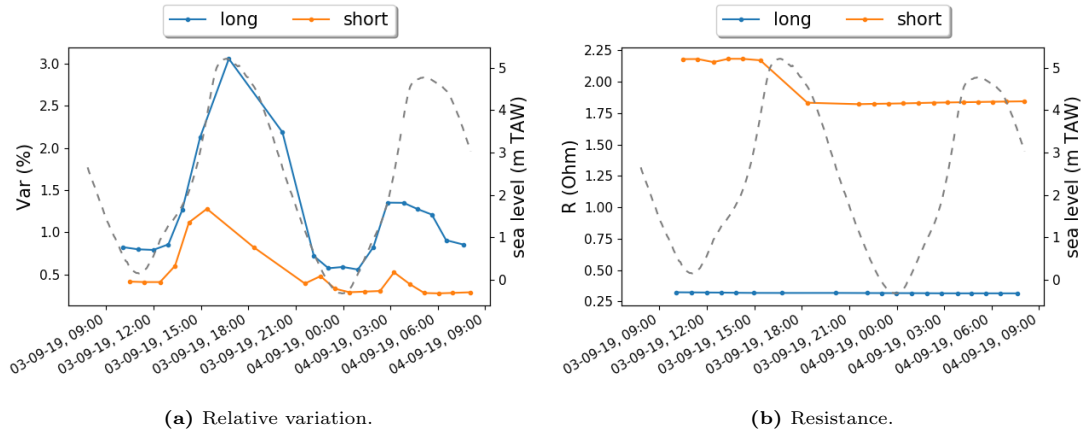


Figure 5.8: Variation coefficient and resistance during the first two tidal cycles.

When looking at the relative variation during the first two tidal cycles there is a clear correlation with the tidal fluctuations, the variation coefficient increases when the equipment is covered with seawater. Figure 5.8a is a zoom on the first two tidal cycles to clearly show the described correlation. This trend is not present in the absolute variation, with the presence of the sea, the average resistance should be lower hence the constant absolute variation. However, figure 5.8b shows the mean resistance but there is no correlation with the tides, note that the figures represent average values so the expected variability is most likely cancelled out by averaging.

The fact that the quality of the measurements declines significantly after a specific point gives reason to believe that there might be a problem with the equipment. The first hypothesis was that corrosion occurs at the electrodes due to the inundation with salt water, however, the electrodes are made of stainless steel so this seems unlikely. To get a better understanding of the problem, the bad data points were plotted in a pseudosection (Fig. 5.9) using equation 3.18.

Short profile 37 was chosen because it has a high relative and absolute variation. Two types of data were plotted, first the resistance, distinguished between positive and negative values, it is impossible for the subsurface to have a negative resistance so these are obvious errors. Secondly the standard deviation (i.e., absolute variation) of the measurements - as described above - was plotted, the threshold between good and bad data points was set at 0.005Ω and was chosen based on figure 5.7b. Both figures show a similar trend, where the majority of the bad data points is located in the dunes, this confirms that corrosion at the electrodes is not the reason for the high variation. Note that the resistance is higher in the dunes - due low saturation - so it is logical that the absolute variation is higher in this part and might contribute to what is defined as bad data points in figure 5.9b. However, from the negative data points alone it is clear that the problematic data points are situated in the dunes. A possible explanation is that the electrodes in the dunes have a bad contact resistance, at the start of the campaign seawater was poured over the electrodes in the dunes to ensure good ground contact but this was not repeated later in time. The electrodes on the beach are frequently inundated with seawater which ensures a good contact resistance. This could explain why the bad data points are limited to the dunes, unfortunately the contact resistance log was not saved so it cannot be confirmed.

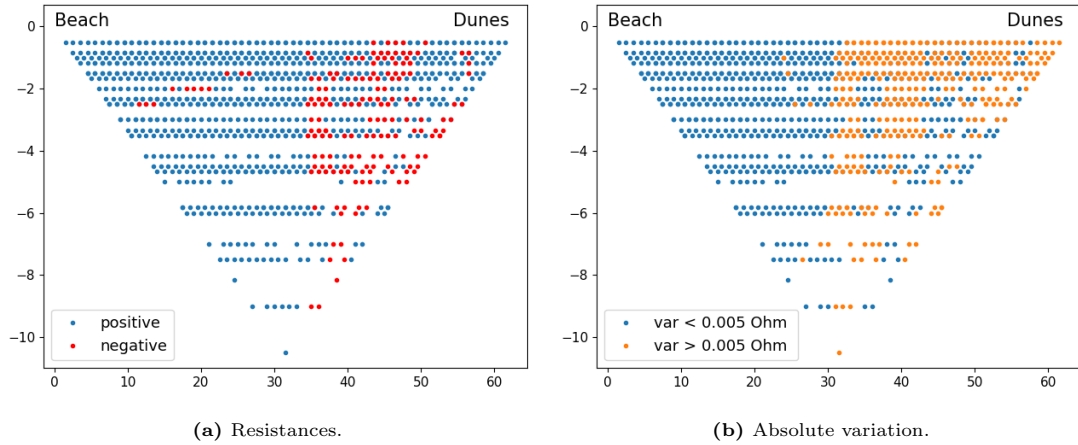


Figure 5.9: Pseudosections of the bad data points for short profile 37.

5.2.3 Profile selection

The previous section extensively evaluates the data quality. For the short profiles it is clear that only the ones measured during the first two cycles will be inverted, both the relative and absolute variation increase significantly from this point forward. As for the long profiles, the relative variation also rises although not as much and the absolute variation remains relatively stable. Thus, more long profiles could be used for the inversion. However, as previously stated, each type of profile has its specific purpose and they complement each other, so the same profiles will be inverted for both types.

The profiles measured during the first two tidal cycles, i.e. 1 through 20 are good candidates for the inversions. For technical considerations, profiles measured during rising and falling tide are not considered for the inversions, only profiles measured during relatively stable conditions will be inverted. This is because the electrodes flooded with seawater vary during the measurement, taking this variability into account can be done by using a 4D inversion scheme where the data is simultaneously inverted accounting for the time of acquisition for each quadrupole (Kim et al., 2009) or by interpolating (Doetsch et al., 2012) but this is out of the scope of this thesis.

To accurately select the profiles, the CTD data is plotted together with the measuring time (Fig. 5.10). The profiles measured at low water are 1, 2, 3, 4, 5, 9, 10, 11, 12 and 13, they coincide with the flat parts of the CTD water height. For the high water profiles, 7 and 16 will be inverted, the water height during measurement of 16 is relatively stable, for profile 7 this is not the case but it is the only one measured at high water for the first tidal cycle so the water height will be considered constant. Note that the effect is higher with the short profile.

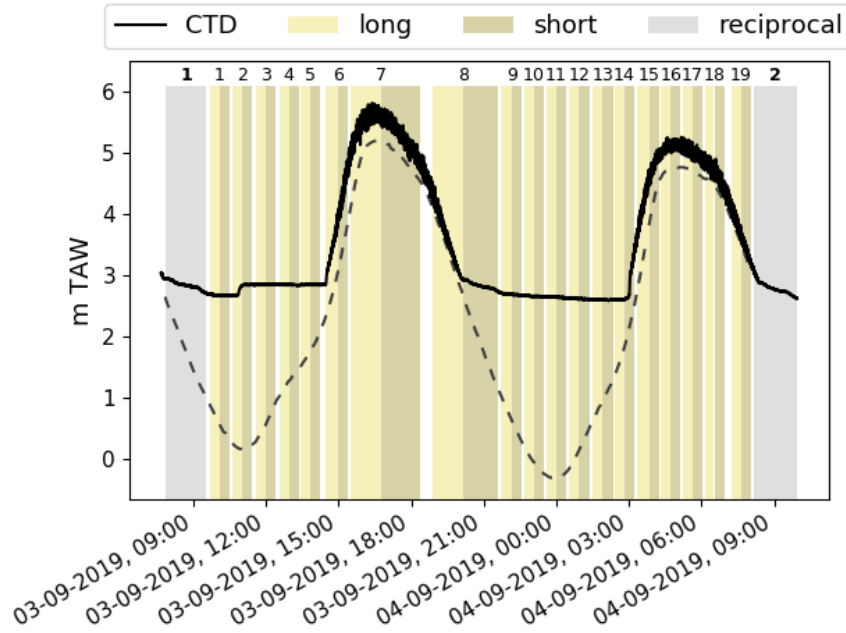


Figure 5.10: Measuring time for the profiles with good data quality, with the water height from the CTD data and Nieuwpoort measurements background.

5.3 Resistance variations

As with the synthetic data, the resistance variations at multiple levels and on different locations of the profiles were plotted for the selected profiles, the two tidal cycles were plotted separately

(appx. C), figures 5.11 and 5.12 show a selection. The situation here is slightly different than in the synthetic study, here the first profiles were measured at low water, and the last one at high water, in the synthetic study there was a systematic increase in the sea level of 0.5 m. We could expect that the resistance remains constant during the dry period, or slightly increases because of drying in the upper layer, followed by a drop in resistance because of the rising water. This is observed in only one plot: on the centre of the short profile, at level 1 during the first tidal cycle (Fig. 5.12a).

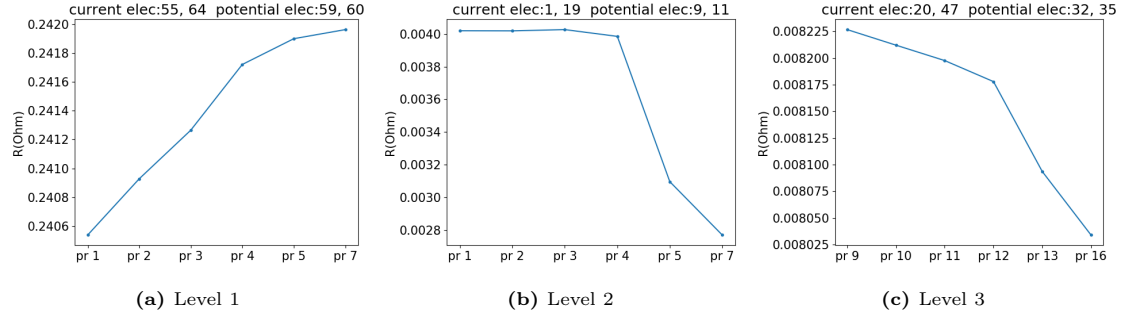


Figure 5.11: Resistance variations of the long profile.

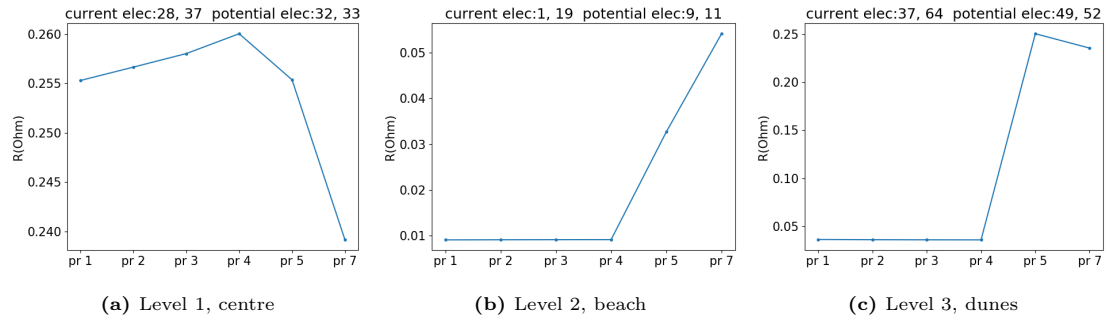


Figure 5.12: Resistance variations of the short profile.

For the long profile, in many cases a more or less stable resistance is observed during the dry period followed by a decrease at high tide. This suggests that no strong variations are captured during the tidal cycle, with exception of the influence of the water layer. Most of the dune plots of the long profile show a continuous increase in resistance, with the occasional negative spike. Because the dunes are not effected by the seawater this is likely an effect of drying. At last, some plots show sudden spikes and in others no clear trend can be distinguished, the sensitivity pattern influences how measurements react to a change in resistance - as explained in the synthetic study - and can thus cause these inconsistencies.

The situation for the short profile is different, where in the long profile many plots are relatively stable followed by a sharp decrease at high tide, the reverse is true for the short profile. In all the plots located close to the sea, the resistance is stable at low tide but increases

significantly at high tide. Also here it can be assumed that only few variations are present during the tidal cycle with the exception of the influence of the water layer at high tide. The increase in resistance - instead of the expected decrease - can likely be ascribed to the sensitivity pattern. For the other locations the same can be said as for the long profile i.e. some show a decrease, some an increase and spikes are abundantly present.

5.4 Error estimation

5.4.1 Results and discussion

Section 3.4 briefly explains reciprocal theory, the idea is that the reciprocal and normal measurements measure the same subsurface and hence should be identical. Thus, the profiles have been measured right after each other to ensure that the resistivity distribution in the subsurface remains as constant as possible. Three reciprocal profiles were measured, all at low water, two of which are situated in the section that will be used for the inversions (Fig. 5.10). It was impossible to measure reciprocals for each profile because it is too time consuming and due to the changing environment. Reciprocal and normal 1 are both measured at low water and can hence be used to calculate the error model. The second pair is reciprocal 2 and normal 19, the normal profile is measured while the tide was retreating, section 5.2.3 explains that these types of profiles are not used in the inversion, the same difficulties will be introduced if this profile is used to calculate the error model. Thus, the error level applied during inversions will be calculated only from the first reciprocal and will be kept constant for all profiles. This is not an ideal solution given that the data quality is varying (see above).

Table 5.2: Error parameters for the different types of models: envelope (env.) and trendline (trend.) and with different threshold (TH) values.

Type and TH level	Long			Short		
	a (Ohm)	b (%)	outliers (%)	a (Ohm)	b (%)	outliers (%)
10 % env.	0.0100	5.00	8.16	0.0150	5.00	0.80
10 % trend.	- 0.00387	3.13	8.16	0.000170	0.98	0.80
5 % env.	0.00100	3.80	16.98	-	-	-
5% trend.	-0.00275	2.31	16.98	-	-	-
2.5 % env.	-	-	-	0.0100	2.100	9.04
2.5% trend.	-	-	-	-0.00251	0.960	9 .04

In section 3.4 two methods are introduced to reconstruct a linear error model, in the first case this is the envelope function encompassing all the remaining errors (Slater et al., 2000), this is a conservative approach where an overestimation is made for all data points. The second option is to reconstruct the trend line, based on Koestel et al. (2008), which is less conservative and

hence more prone to artefacts. Both approaches will be tested to assess which one is best suited for this dataset. The models are based on the remaining errors, hence after removal of outliers, generally a cut-off value of 10 % is applied but this value is subjective, thus different threshold values will also be explored. Table 5.2 summarizes the error parameters (a and b) reconstructed by applying different approaches, an example is shown in figure 5.13, the remaining figures are given in appendix D.

In the following section the inversion results of the different error models are quickly compared to determine the best suited error model, a more thorough description of the results themselves follows later.

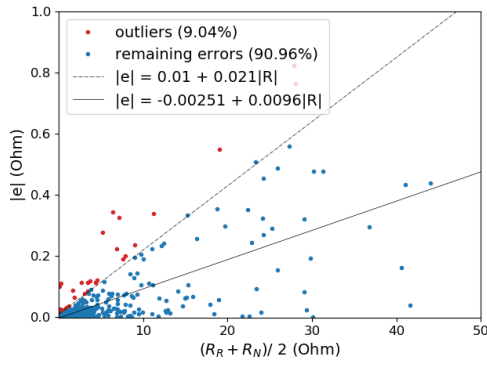


Figure 5.13: Reconstruction of the error model for the sort profile at a threshold value of 2.5 %.

First the traditionally used cut-off value of 10 % for the reciprocal error was used, when applying the resulting values in the inversion, the envelope approach yields an over-smoothed image for the long profile (Fig. 5.14a), a very rudimentary seawater lens can be seen but it is not clearly delineated. Fresh water is present under the dunes, as expected. The trend line approach gives a more satisfactory result (Fig. 5.14b) the seawater lens is clearly present in the centre of the profile with fresh water underneath. Under the dunes, there is brackish water present, which can be a remnant of the infiltrating seawater

from when the inlet was working. There is also brackish water present under the freshwater, which is probably related to a change in lithology, i.e. clays from the Kortrijk formation (Paepen et al., 2020), which are more conductive than the sands on top. Note that the trend line model has a negative absolute error, this is not realistic so a value of 0 ohm was applied in the inversion, this approach is used for all negative errors. From the two described options, the second is certainly preferred however, an absolute value of 0 ohm is unlikely. Because the percentage of outliers is low (8.16 %) the threshold value is lowered to see which effect this has on the inversion, the next lower threshold that has a notable effect on the error model is at 5 %. This has a significant effect on the data inverted with the envelope model (Fig. 5.14c), the seawater intrusion is much better delineated compared to the 10% threshold, this means that being more strict for the threshold allows to derive a globally less conservative error model, even with the envelope. As for the trend line the models look similar (Fig. 5.14d), the absolute error is also negative with the lower threshold. Note that the models inverted with the trend line model have a much higher resistivity range, probably due to over-fitting.

The results for the short profiles are given in figure 5.15, the 10 % threshold yield good results, for the envelope the salt water lens is well delineated and a small freshwater lens is clearly

present (Fig. 5.15a). In contrast, with the trend line the freshwater lens is not as clearly defined, also brackish water is present at depth and in the dune area showing that more artefacts are present when using the trend line approach. Because for the long profiles a lower threshold was tested, the same is done here, also only 0.8 % of the data points are outliers so the threshold can be easily lowered. At 2.5 % both models look similar compared to their 10 % counterparts, the resistivity range is slightly higher for both models.

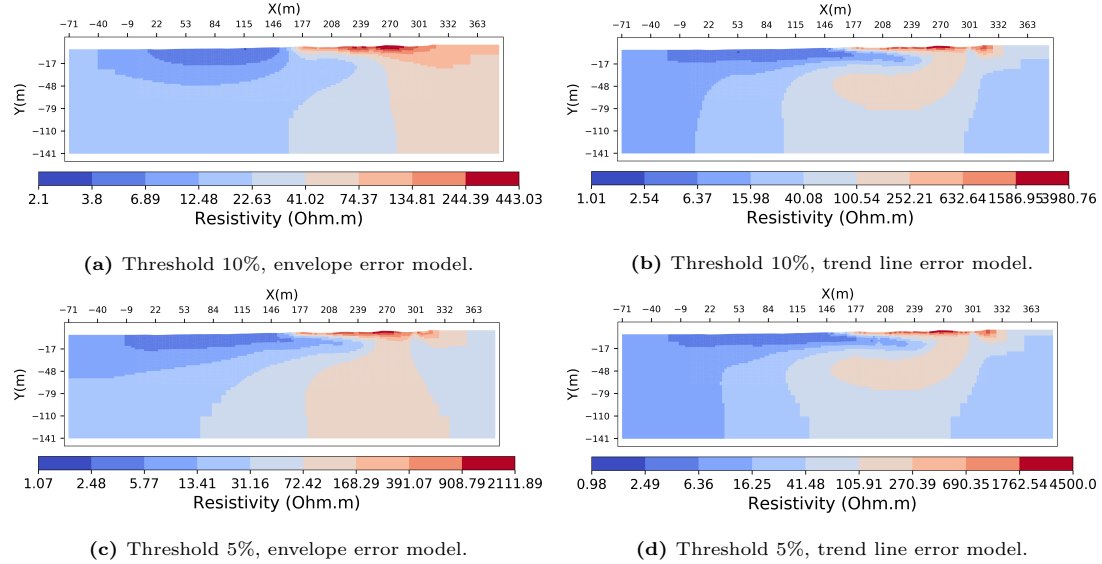


Figure 5.14: Comparison of different error models for long profile 1.

5.4.2 Conclusion

For the long profile, both the trend lines and envelope at 5 % could be used, however the trend lines give a sharper vertical delineation of the seawater lens and are thus preferred. The trend lines do contain more artefacts but are mostly present at the edges of the grid, this has to be taken into account when interpreting the results.

Table 5.3: Error parameters used for the inversions.

	Long	Short
TH (%)	5.0	10
a (Ohm)	0.001	0.015
b (%)	3.8	5.0

For the short profiles it is clear that the trend line approach is not satisfactory, structures cannot be clearly distinguished and the data is over-fitted. For the envelope both threshold values give a good result with a comparable visualization capacity. Since there is no clear difference between the thresholds, the preference goes to the traditional 10% level, because the resistivity range is slightly smaller

To be consistent, the same approach will be used to calculate the error model for the long and short profile, thus either the envelope or trend line. Since using the trend line is strongly discouraged for the short profile,

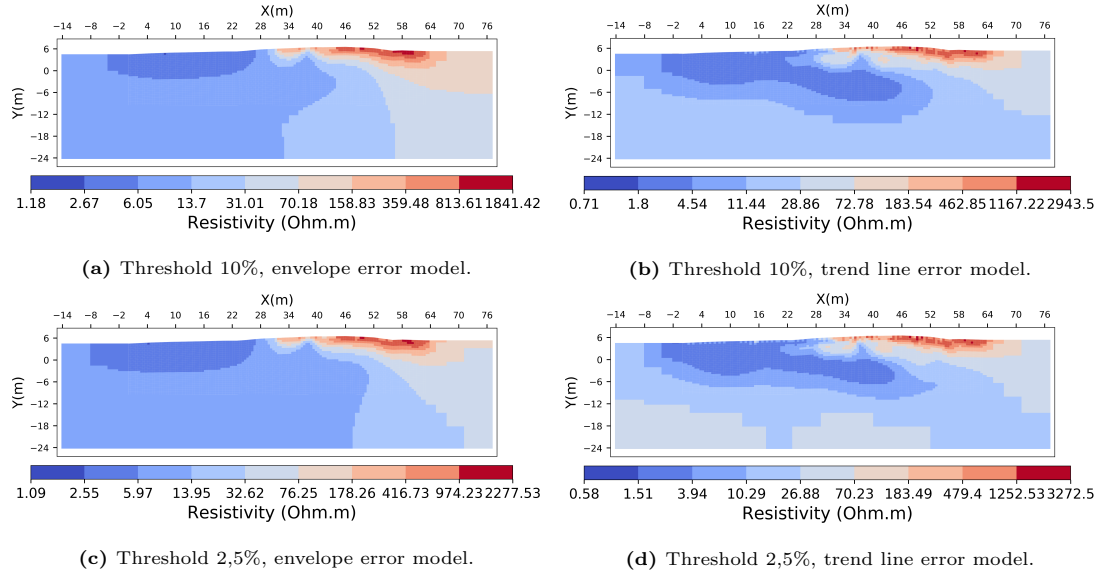


Figure 5.15: Comparison of different error models for short profile 1.

the envelope will be used to reconstruct the error model, the used parameters are summarised in table 5.3. Although the trend line is the preferred approach for the long profile, the envelope still gives satisfactory results. Given the varying the quality of the data, a conservative approach is safer for the inversion of the subsequent profiles.

The previous section illustrates the importance of a suitable error model, a small error does not necessarily mean that the inversion results will be better as clearly illustrated by the short profile. The errors here are smaller in the trend line approach but the higher errors from the envelope are visibly better, the smaller errors result in an over-fitting of the data, creating more artefacts. These artefacts resulting from over-fitting are also clearly present in the long profile.

To conclude, in this study the more conservative envelope model is preferred with a lower threshold value to allow for a less conservative model when needed. However, since the data quality varies with time (see section 5.2.2) more reciprocals are likely needed to reconstruct a more accurate error model (Lesparre et al., 2017). To avoid the problem of acquisition time, there are a few possible alternatives that can be used in future monitoring campaigns; a subset could be acquired for instance 10 % of the full profile, the second option is to measure a reciprocal data point immediately after the normal but this also reduces the efficiency of acquisition. Lastly, a combination of the two could also be applied.

5.5 Absolute inversion with homogeneous reference model

5.5.1 Results

5.5.1.1 Long profiles

Figure 5.16 shows two examples of the inversion of the long profile, one at low and one at high water. The sensitivity, also shown in figure 5.16 is high in the upper layer, the zone of high sensitivity is thicker under the dunes compared to the beach. Current preferentially flows in low resistivity zones, lowering the sensitivity of deeper layers it is thus logical that the zone with high sensitivity is thicker under the more resistive dunes compared to the beach. The sensitivity is sufficiently high until a depth of approximately 20 m, deeper results have to be interpreted with caution, also at the edges of the grid the sensitivity drops, as with the synthetic model. Note that extra cells were added to the edges of the grid to limit the influence of boundary conditions of the data, coinciding with low sensitivity zones. Results should only be interpreted between 0 to 315 m. The inversion results look similar in terms of geometry, the profile at high water contains lower resistivity values, due to the presence of the highly conductive seawater, also the lens is slightly better delineated at depth compared to the low water profile. The remaining profiles also look similar and are hence not shown, what is more informative is the resistivity range of the inversion results and the differences between the profiles. The differences are calculated with respect to the first profile of each tidal cycle, thus, in the first cycle differences are calculated with respect to profile 1, in the second cycle to profile 9.

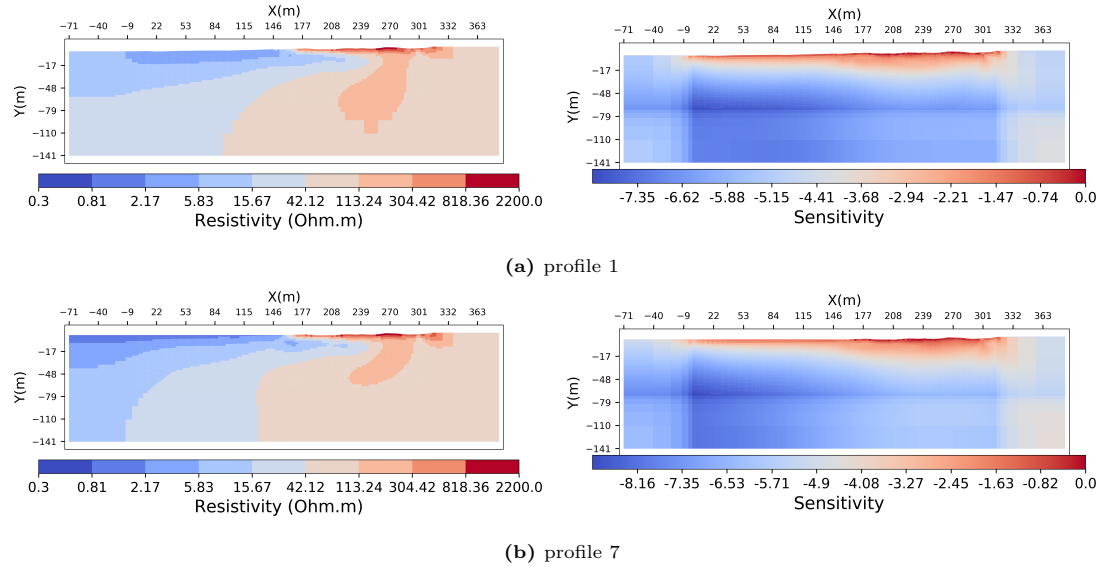


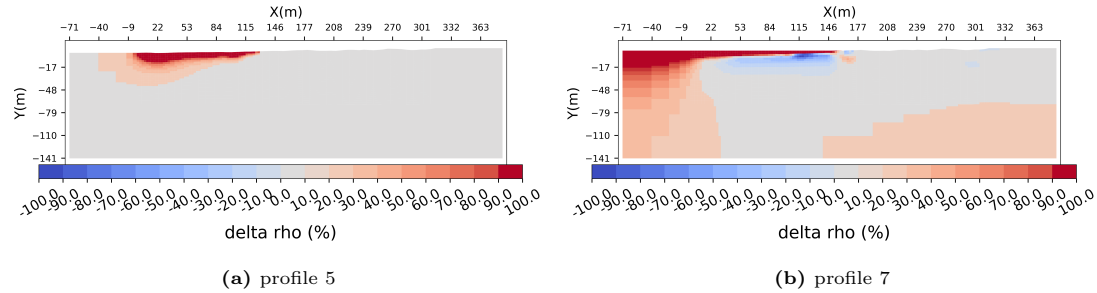
Figure 5.16: Examples of the inversion results of the long profile using a homogeneous prior model.

Table 5.4: Resistivity range field data inversions, the dashed line separates the tidal cycles.

Profile	Long		Short	
	min.	max.	min.	max.
1	1.07	2092.59	1.18	1841.72
2	1.07	2017.08	1.18	1837.10
3	1.07	2132.46	1.17	1885.57
4	1.05	2069.69	1.18	1966.53
5	0.50	2129.67	0.42	1876.86
7	0.34	2395.79	0.35	2004.62
9	1.07	1602.22	0.34	2230.87
10	1.07	1746.73	0.34	2258.50
11	1.07	1596.35	0.38	2229.83
12	1.07	1742.65	0.35	2247.36
13	0.64	2098.20	0.35	2265.30
16	0.36	2220.15	0.30	2199.23

Table 5.4 shows the minimum and maximum resistivity of the inversion results. In the maximum value there is no obvious trend present but for the minimum values there is. When looking at the first tidal cycle, i.e. profiles 1 through 7, one can see that the resistivity remains at a fixed value of $1.07 \Omega.m$ until the profile 3, at profile 4 there is a small decrease ($1.05 \Omega.m$) followed by a steeper decrease in profiles 5 $\Omega.m$ and 7 of 0.50 and 0.34 $\Omega.m$ respectively. A similar trend is present in the second tidal cycle.

As for the calculated differences (Fig. 5.17, Appx. E), in profiles 2 and 3 no significant changes occur. For profile 4, there is a small cluster of red cells present around 22 m on the profile, meaning there is a slight decrease in resistivity in profile 4 compared to 1. At profile 5, the same change is also present but more substantial, the area where a negative change in resistivity occurs is 138 m long and has a depth of -48 m on the beach closer to the sea, decreasing to -10 m further towards the back shore (Fig. 5.17a). For profile 7, a layer is present on top of the model with lower resistivity, immediately underneath, between 22 and 146 m a lens is present with higher resistivity compared to profile 1, a negative resistivity change is also present at depth under the dunes (Fig. 5.17b). The results of the second tidal cycle are similar and hence not further discussed.

**Figure 5.17:** Resistivity differences for two subsequent long profiles.

5.5.1.2 Short profiles

A selection of the inversion results of the short profiles are given in figure 5.18. The sensitivity is again higher in the upper layer, the zone of high sensitivity is thicker in zones of high resistivity,

as with the long profile. Here there were also cells added on the edges of the grid, in this case the results should only be interpreted between 0 and 63m. A clear salt water lens is not present, what can be distinguished is the mixing zone, the front of which is present between 52 and 58 m on the profile, a small freshwater lens is present in the centre of the profile. The structures are well defined in the first four profiles but starting from profile 5 (see Appx. B), the inversion is more prone to artefacts what might indicate a change in noise level. Brackish water is present under the dunes, where in the earlier profiles there was freshwater present in this zone. Also the freshwater lens is not as clearly defined and a low resistivity zone is present at depth between 40 and 52 m and increases in size for the subsequent profiles (Fig. 5.18b). In later profiles, a freshwater tongue seems to intrude the brackish water. Although the high water profiles are more prone to artefacts, the influence of the water layer can be seen as lower resistivities are present in the upper layer compared to the other profiles.

The minimum resistivity in the first four profiles has a constant value of $1.18 \Omega.m$ (table 5.4) from profiles 7 through 16 there is no clear trend visible but the values are all circle around $0.35 \Omega.m$. This seems to indicate that the cycles are not similar and more changes occur during the first cycle. As for the maximum resistivity there is no clear trend visible, as was the case for the long profiles.

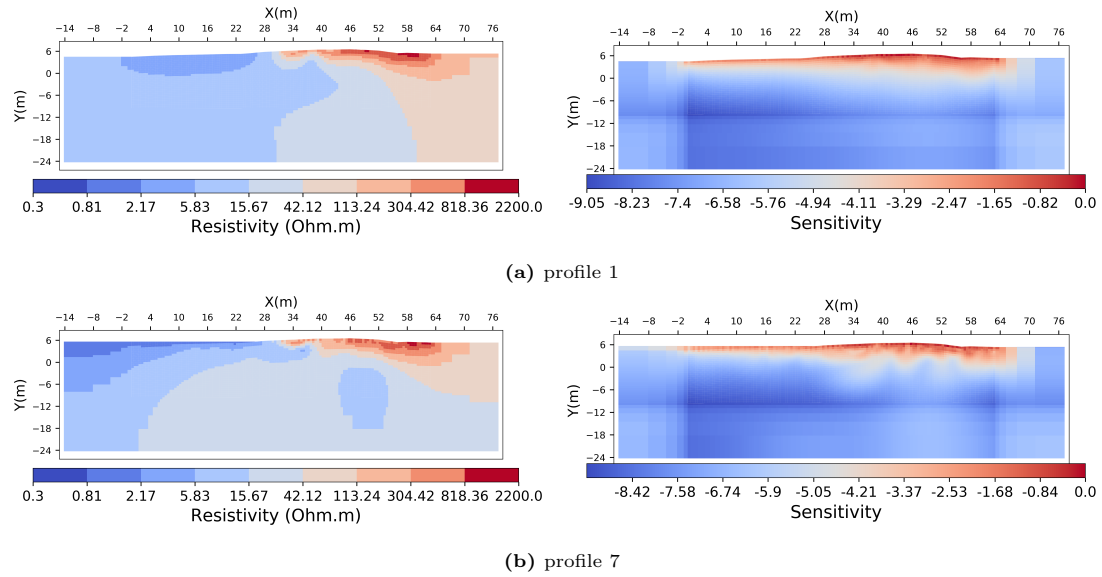


Figure 5.18: Examples of the inversion results of the short profile using a homogeneous prior model.

Figure 5.19 shows some examples of the calculated differences in the first tidal cycle for the remaining figures we refer to appendix E. As with the long profiles, the differences were calculated with respect to the first profile of the tidal cycle, hence profiles 1 and 9 for respectively the first and second tidal cycle. For profiles 2, 3 and 4 all changes are present in the range considered as artefacts. For profile 5 there is a decrease in resistivity (compared to profile 1) between -2

and 22 m, note that this red zone has a similar shape as in long profile 5, probably indicating a re-saturation with water. Further, there are some small changes, in the centre of the profile a positive and under the dunes a negative change. In profile 7, there is a decrease in resistivity present in the upper layer with immediately underneath a dark blue lens representing an increase. Also here there are smaller changes present, similar to those in profile 5 but more pronounced. As for the second tidal cycle in profiles 10 through 13, there are minimal changes, there are some larger differences in the shape that roughly outlines the red lens in profile 5. In the high water profile, there are more changes present in this region, with negative changes near the surface and positive changes deeper.

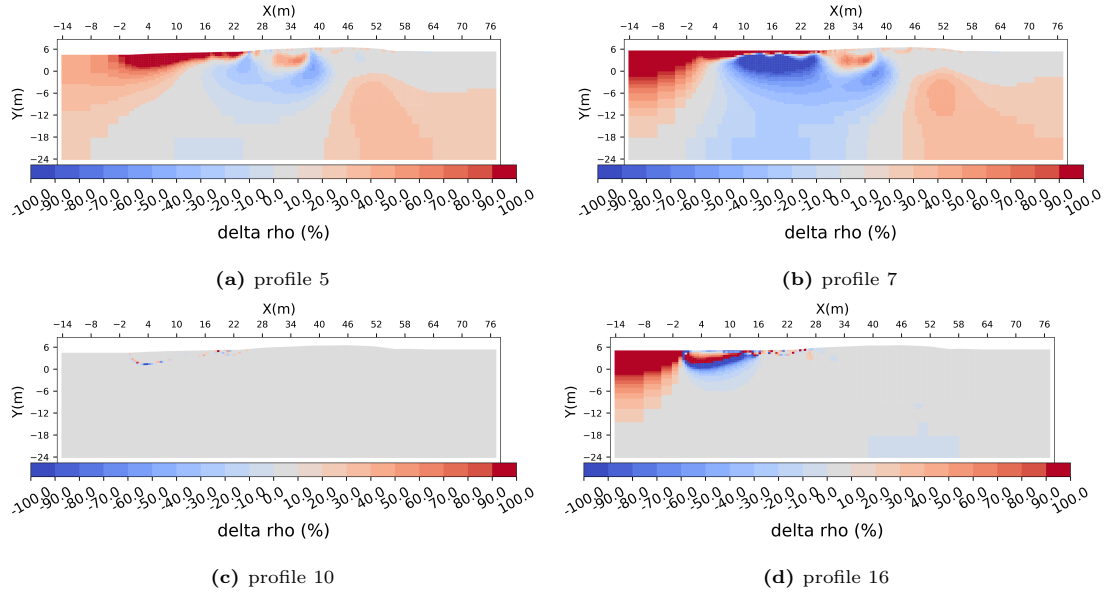


Figure 5.19: Resistivity changes of different short profiles.

5.5.1.3 Discussion and conclusion

For the long profile the resistivity values and differences are much more informative than the figures of the inversion results themselves. The trend seen in the minimum resistivity, i.e. stable at low water followed by a decrease at high water can be clearly related to the rising tide. Note that profiles 4 and 5 are not inundated during measurements indicating that the sediment is saturating when the tide is rising. The same trend can be seen in the figures depicting the differences (Fig. 5.17) where in profile 5 there is clearly a decrease in resistivity that is thicker near the sea, for profile 4 the decrease is very subtle - as well as the drop in minimum resistivity - this indicates the start of the saturation at the beginning of the profile. Note that only one hour passes between profile 4 and 5 so the saturation happens rather quickly. At profile 7 there is a low resistivity layer present on top of the sediments, corresponding to the seawater layer. The

blue zone underneath, where the resistivity is higher than in the first profile is likely an artefact of the inversion. The synthetic study clearly stipulates that the water layer causes distortions in the inversion results, which will likely be the case here as an increase due to drying or freshening is not expected during the rising tide. For the second tidal cycle the results are similar and hence represent a second cycle of saturation with seawater. For the long profiles, the changes induced by the rising tide are mostly related to the saturation and hence resistivity in the upper layer, no changes or fluctuations are evident in the seawater lens, likely because the resolution is not good enough to capture possible variations or because they are too small. Note that for profiles 2, 3, 10, 11 and 12 there are no significant differences, however, during ebb the sediment dries so we could expect to see an increased resistivity in these profiles since in the synthetic study the dry layer is well recovered. On the long profile, no indication of a dry layer is present, in the synthetic study a fairly thick dry layer was reconstructed so it is possible that when only a thin layer of increased resistivity is present, it is not recovered by the inversion, especially considering the large electrode spacing another possibility is that the drying happens fast.

For the short profile a small freshwater lens is present at the surface between 30 and 38 m and is present on all profile, this could be related to a small freshwater discharge zone corresponding to the phreatic aquifer, to the author's knowledge this phenomenon has not been reported in literature. A clear salt water lens is not present as in the long profile which is related to the scale of the short profile. Here much smaller electrode spacings were applied to acquire a more detailed image, with as a result that the depth of investigation is significantly smaller compared to the long profile and hence not deep enough to capture the full vertical extent of the seawater lens. As for the long profile, the resistivity range and relative differences are more informative than the resistivity images. For the minimum resistivity during the first cycle, a similar trend is observed as in the long profile, i.e. stable at low tide followed by a decrease at high tide. As for the differences, a zone with lower resistivity is present in profile 5 with a similar geometry as in the long profile and is likely also the result of re-saturation of the sediment during high tide (before the water reaches the equipment). Further, a small oval-shaped zone with lower resistivity is present between 30 and 38 m, note that this is the location of the freshwater lens, this suggests that with the rising tide, the lens might become slightly more brackish however not significantly because it could still be clearly distinguished on the high water profile. This zone is surrounded by a blue zone representing a higher resistivity and could be an indication of changes occurring in the transition zone between salt and freshwater. When looking at the differences in profile 7 a similar situation arises but is superimposed with the same types of artefacts found in the long profile, the presence of the water layer causes distortions in the inversion results, in the long profile this is expressed as a substantial - blue - zone with higher resistivity and is here also present. Note that the presence of the water layer can also be clearly seen as the red layer on top of the model. Note that also here, no remnants of a dry layer are recovered. The situation for the second tidal cycle is different from what described above. In the first cycle a trend similar to what is observed for the long profiles is present, i.e. a stable resistivity followed

by a decrease at high water. Although, in the second cycle this remains approximately stable, this could suggest there is a difference between both cycles. As for the differences, during the first cycle clear patterns arise (discussed below) while for the second cycle this is not the case, only at the high water profile significant variations are present but they follow a completely different pattern. We observed that starting from profile 5 forward the results are more prone to artefacts during the first tidal cycle this effect is rather small but in the second cycle it is much more pronounced. The likely explanation is that the error model is not suited for these profiles as it is based on the first profile, this could also explain why there are no significant variations during the second tidal cycle, i.e. because of a badly estimated error model, the data is more prone to artefacts and hence the resistivity distribution could not accurately be reconstructed (Lesparre et al., 2017).

Thus sum up, in both the long and the short profile the influence of the rising tide can be seen in the inversion results, a re-saturation of the sediment occurs before the water reaches the equipment, lowering the resistivity. The presence of the water layer causes distortions in the results, as thoroughly described in the synthetic study, and is expressed by a zone with much higher resistivity immediately under the surface. For the long profile no smaller variations can be distinguished but in the short profile it can be seen that the rising tide causes subtle changes that might be related to variations in the salt- and freshwater distribution. For the short profiles it was chosen to limit the interpretation to the first tidal cycle because the second tidal cycle is more prone to artefacts and this could lead to a misinterpretation of the data.

5.6 Time constrained inversion

5.6.1 Results

5.6.1.1 Long profiles

For the time constrained inversion, i.e. with a realistic reference model, the used prior model is the result of the inversion of profile 1 using a homogeneous reference model. In terms of the visual aspect, the inversion results look similar as when inverting the data using a homogenous prior model (Fig. 5.20, Appx. F), the sensitivity is also higher in the upper layer with a thicker sensitive zone under the more resistive dunes. Note that also here the data should only be interpreted between 0 and 315 m.

The iteration process finished quickly for the low water profiles - after only one iteration - this suggests that the prior model explains the data well, which was expected since in the previous section it became clear that minimal differences are present between these profiles. For the high water profiles more iterations were made to converge to a solution, due to the influence of the water layer these dataset differ more from the prior model.

In terms of the resistivity range (Tbl. 5.5), similar observations are made compared to the previous inversion type. During a tidal cycle, the minimum resistivity remains relatively

constant and then rapidly declines when the water rises and approaches the equipment. The decline in resistivity at the low water profiles are related to re-saturation of the sediment due to groundwater flow during high tide.

Table 5.5: Resistivity range field data inversions with a realistic reference model, the dashed line separates the tidal cycles. The results for the inversion of the prior model are also added, for comparison.

Profile	Long		Short	
	min.	max.	min.	max.
1 (<i>prior</i>)	1.07	2092.59	1.18	1841.72
2	1.04	4433.84	1.12	1978.94
3	1.03	4262.61	1.10	2030.38
4	0.98	4376.65	1.15	1854.95
5	0.46	4562.45	0.44	2522.31
7	0.33	2993.24	0.34	2752.30
9	1.04	4297.69	0.41	2679.64
10	1.04	4286.58	0.41	2673.02
11	1.04	4285.22	0.42	2624.04
12	1.04	4274.01	0.42	2659.96
13	0.65	2272.61	0.43	2621.36
16	0.34	2720.30	0.30	2738.33

were also present in the preceding profiles. For the high water profile, the difference looks very similar to what is observed in the previous inversion type and are hence not shown the main characteristics are a strong influence of the water layer and a zone of higher resistivity underneath due to the distortions caused by the water layer but here the presence of the small variations is not present. The second tidal cycle shows the same trends and are hence not further discussed.

As for the calculated differences the results differ slightly from the previous inversion type (Fig. 4.5d, Appx. F). For all low water inversions, the same variations are present i.e. a thin layer of lower resistivity between 22 and 208 m under the surface, with at some parts blue values on top, indicating a higher resistivity. Under the layer with lower resistivity, between 155 and 220 there is an oval shaped zone with higher resistivity present (Fig. 5.21a), this is the case for all low water profiles that are not yet influenced by the rising tide. In profile 5 (Fig. 5.21b) the influence of the rising tide can be seen as the sediment gets saturated and can be seen as a zone with lower resistivity between -9 and 120 m that gets thinner towards the back shore. Note that the re-saturation effect is superimposed on the smaller variations that

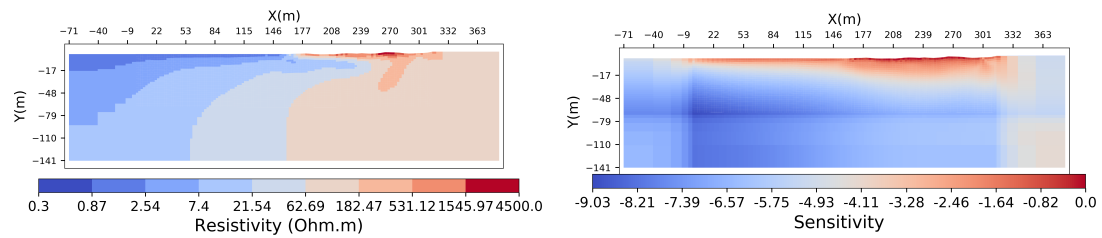


Figure 5.20: Result of the time constrained inversion for long profile 7 with the corresponding sensitivity.

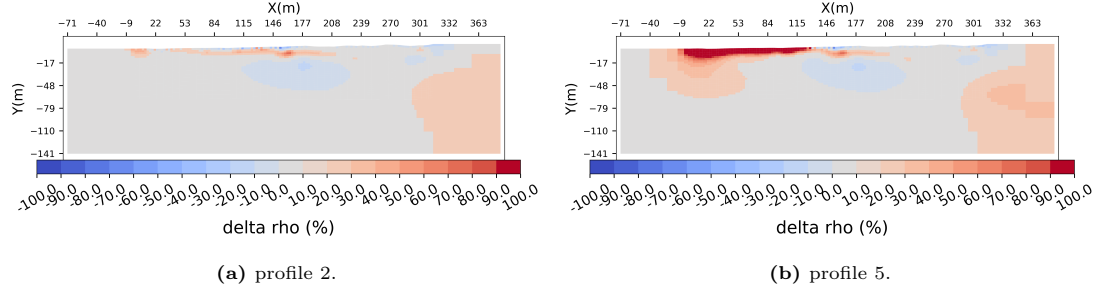


Figure 5.21: Relative differences for the long profiles, with reference to the prior model.

5.6.1.2 Short profiles

As for the long profile, the inversion was made by using the inversion result of profile 1 as a prior model for all subsequent inversions. The interpretation should again be limited to the zone between 0 and 63 m, because the edges of the grid are added to limit effect of boundary conditions and hence contain no data. The inversion results look similar as the previous inversion type (Fig. 5.22, Appx. F), also here the sensitivity is highest in the upper layer and extends deeper in the more resistive dunes. Profiles 2, 3 and 4 look practically the same as their counterparts using a homogeneous reference model, the top of the seawater lens is well defined, more specifically the mixing zone. When looking at the resistivity between 6.24 and 17.16 $\Omega.m$ it forms a clear front at 52 m, note that the zone under -6 m is not considered since the sensitivity here is very low. The small freshwater lens is also well defined on these profiles (e.g. Fig. 5.22a).

From profile 5 forward (Fig. 5.22b), geometry of the seawater lens is lost, the shape of the mixing zone is very irregular compared to profile 2, it seems as if a tongue of higher resistivity intrudes this zone, which gets larger in later profiles (e.g. Fig. 5.22c). This could point to a variation in the salt and freshwater distribution but could also be an artefact, caused by an unfit error model, as was the case in the previous inversion type. The small freshwater lens is still present in these profiles albeit less well delineated which could point to a badly reconstructed error model. In profile 5 the resistivity is lower in the upper layer between 0 and 15 m, pointing to a re-saturation with seawater during high tide when the sea has not yet reached the equipment, in profile 7 the presence of the seawater layer is evident as a layer with low resistivity on top of the sediment.

In terms of the minimum resistivity (Tbl. 5.5) during the first tidal cycle, the expected trend is present. The resistivity is relatively constant for the first four profiles and then rapidly declines at profile 5 due to the re-saturation and then further drops in profile 7 when the sediment is inundated with the highly conductive water. This trend is not present in the second cycle, as was the case for the previous inversion type. So it is likely also the effect of the unfit error model.

The relative differences in resistivity are all calculated with the prior model as background, in all profiles the resistivity is lower in a significant fraction of the profiles with values between

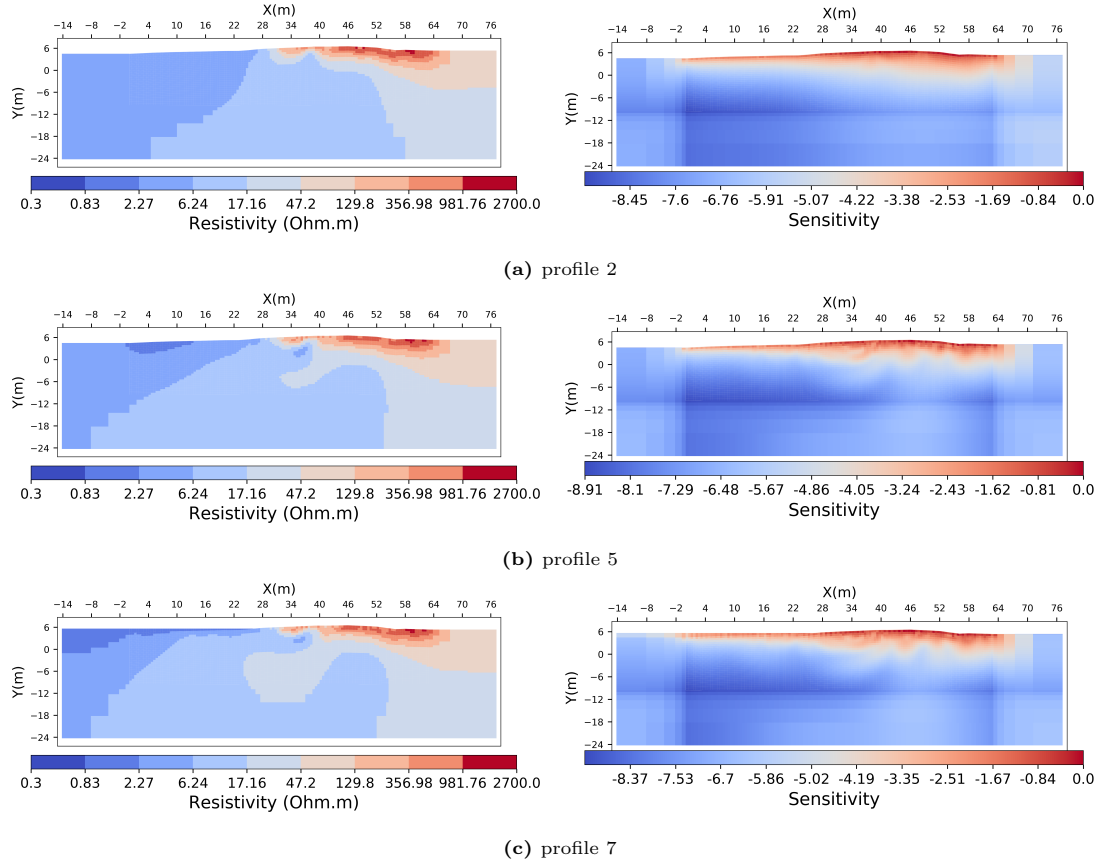


Figure 5.22: Result of the time constrained inversion for short profiles 5 and 7.

10 and 50 %. For profiles 2 and 3 the pattern looks quite unusual where in the upper layer there are no or limited changes in two elongated zones (e.g. 5.23a), note that this shape roughly resembles the sensitivity pattern. Thus, for the zones with high sensitivity no detectable changes occur in profiles 2 and 3. Profile 4 does not show this unique shape of virtually no changes but still contains significant changes at depth (see Appx. F). For profile 5 the changes are similar to what is observed in the previous inversion type, with the addition of the changes at depth as observed in the previous profiles. The changes include: a lower resistivity in the upper layer corresponding to the re-saturation, a zone of lower resistivity between 30 and 38 m corresponding to the location of the freshwater lens and a zone of higher resistivity surrounding it (Fig. 5.23b). The same can be said for profile 7, the changes are similar as in the previous inversion type meaning the water layer is clearly present and caused strong distortions, with the addition of a lower resistivity at depth and on the edges of the profile (see Appx. F). The second tidal cycle is not as expected, all the profiles show a similar pattern as what is seen in profile 5, profile 16 has the additional effect of the water layer on top.

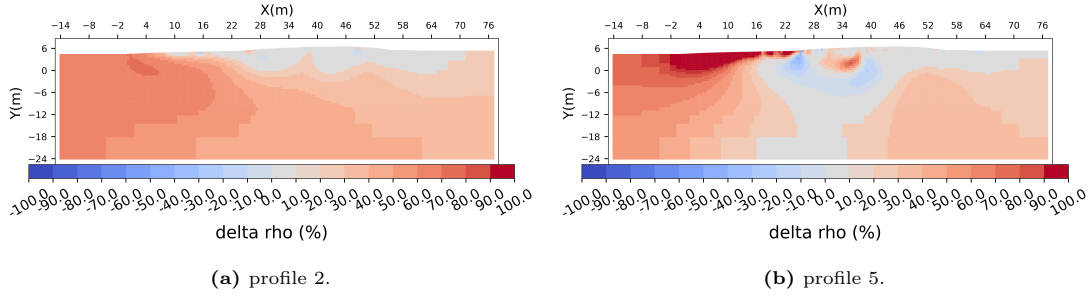


Figure 5.23: Relative differences for the short profiles, with reference to the prior model.

5.6.2 Discussion and conclusion

For the long profile, we can be brief since the results are very similar to what is observed with the independent inversions. As the tide rises, the sediment gets re-saturated before the sea covers it, which causes a significant decrease in resistivity that can be seen in both the absolute difference images and the minimum resistivity values. At high water the seawater layer causes clear distorting in the inversion result that manifests itself as a zone with significantly higher resistivity immediately under the surface. The same process happens during the second tidal cycle. The small variations present in the profiles could be an indication that profile 1 is drier than the subsequent profiles, however, since the same variations are present in all the profiles it is likely an artefact of the inversion. The inversions were finished after only 1 iteration for the low water profiles suggesting the prior model provides a good estimation for the resistivity in subsequent profiles, which significantly reduced the computational time.

For the short profiles, the results also look similar to what is observed when using a homogeneous reference model with the addition of slightly lower resistivities present in all profiles. The re-saturation and inundation causing distortions are clear during the first tidal cycle, so is the lens of slightly lower resistivity suggesting the freshwater lens becomes more brackish at high water. The second tidal cycle is again more complicated, the relative differences suggest there are variations in the salt- and freshwater distribution. However, there is no trend present in the minimum variation and the variations are similar in all profiles, it is likely that the observed variations are artefacts rather than real variations. This could be due to the fact that the error model is based on the reciprocal measured at the start of the campaign which is likely not suited for these profiles.

In terms of interpretation the same things are observed as when inverting independently (i.e using a homogeneous reference model) as when using a time-constrained (i.e. using a realistic reference model), however, using the prior model could help identify more subtle variations. Note that the dry layer is also not detected here, however a re-saturation is observed so the drying does occur but it likely happens quickly.

Chapter 6

Overall conclusion and future recommendations

The main objective of this thesis was to assess the ability of ERT to monitor the seawater intrusion on a very short time scale. Since, to the authors knowledge, a similar study has not been done before this came with some challenges. First, because of the highly conductive seawater layer, a large fraction of the current is transmitted into the water layer instead of penetrating the surface, which can cause strong artefacts in the inversion results which is a well known effect (Henderson et al., 2010). Next, the monitoring is executed in the intertidal zone, in this dynamic environment the electrodes are partially, and variably inundated during measurements introducing another difficulty.

First we studied the effect a water layer can have on the inversion results with synthetic models, different inversion methods were applied to assess the effect on the data and see which approaches are well suited. The presence of the water layer has a strong effect on the data when it is significantly thick. For the models at the low water levels the geometry of the seawater lens could be recovered fairly well, however at the high water level, the geometry was completely lost and the lens assumed a unusual shape. The influence of the water layer is also clear when considering the relative differences in which large artefacts were present in the vicinity of the lens. Thus, the water layer significantly effects the data both qualitatively and quantitatively in a manner difficult to anticipate. By explicitly adding the water layer by means of a reference model, the inversion results considerably improved and is hence highly recommended when inverting similar data. In the case here, using a realistic reference model yielded in an over-fitting because the prior model was too close to the solution, using the inverted data as a prior instead of the original dataset might improve this and is also more realistic since that is what can be used in field experiments. The results from the time-lapse inversion were unsatisfactory, in both cases, i.e. the dry and saturated model, the solution did not converge to a rms error of 1 but to respectively 106.2 % and 14.60 %. The algorithm likely focusses on the changes in the water layer instead of the subsurface. To overcome this an additional constrained could be developed to ensure no changes occur in the water layer so that the algorithm computes the changes in the subsurface. Apart from the fact that the time lapse inversion was displeasing, in the saturated case the geometry of the lens could be recovered which was not possible when inverting independently, thus time-lapse inversion on this dataset is promising but needs further investigation. Extra attention should also be given to the correction on the model with a dry layer, the corrected dataset here was of very poor quality but in order to accurately predict what can be expected with field data sets this should also be considered. Note that in this study a best case scenario was replicated, hence without the additions of noise, in future studies this should be added to get a more realistic

representation of field data.

On the measuring campaign, four full tidal cycles were measured, however, the data of only the first two cycles was well suited to invert for the latter two the repeatability error increased significantly. The majority of the bad data points was located in the dunes so this decrease of data quality was attributed to a poor contact resistance. When similar campaigns are executed in the future, especially when the objective is to monitor on a long scale, water should preferably be poured on the electrodes daily to avoid this. Two profiles were monitored, one long profile with large electrode spacings to detect deeper variations and one short profile with smaller electrode spacings to detect more subtle variations near the front of the seawater lens. The data was inverted using a reference model so the water layer could be added explicitly, as from the synthetic study it was clear this yielded a much more satisfactory result. Both a homogeneous as a realistic reference model were used and they yielded similar results. As the tide rises, the subsurface gets saturated with the conductive water, this can be seen in both types of profiles as the resistivity drops to a value approximating seawater before the sediments are covered with water. When the water layer reaches the equipment, significant distortions are caused in the inverted dataset, as can be expected since this has been observed in the synthetic case as well as by Henderson et al. (2010). For the long profile, this trend is present in both tidal cycles but for the short profile that is not the case. The absence of the observed trend is likely not the indication of a process but point to a badly reconstructed error model. To obtain an accurate error model, reciprocal measurements should be made for each time step (Lesparre et al., 2017), however when monitoring in a rapidly changing environment this might be too time consuming. During our field campaign, three reciprocals were measured a first improvement could be made by measuring a reciprocal at each low tide - which was not done here - another option is to measure a partial reciprocal for each normal profile, that way an error estimation can be made for each individual profile without compromising the acquisition time too much. At first glance, the time-constrained approach was able to detect more subtle variations, however, as they are consistent throughout the profiles it could also be artefacts, note that here we monitor on a very short time scale so any possible variations might be too small to capture. In both the long and the short profile no changes in the fresh and saltwater distribution are definitively present, a time-lapse inversion might be able to capture small variations but first the methodology has to be refined using synthetic models before applying it to field data.

To sum up, the presence of the water layer causes clear distortions in the inversion results as observed in the synthetic model and confirmed by the field data, this could make the detection of variations in the interface challenging given the magnitude of the distortions, which are up to 100%, that could easily mask any possible changes. Nonetheless, monitoring using ERT looks promising providing that certain things are investigated more thoroughly, for instance the effect of a dry layer and the time lapse inversion, when these methodologies are refined, a time-lapse inversion can be applied in which any possible variations might become clear.

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Appendix A

Resistance variations synthetic models

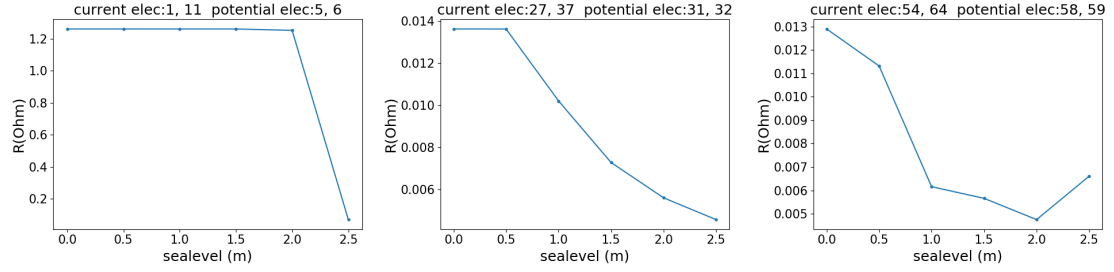


Figure A.1: Resistance variations of the synthetic model at level 1.

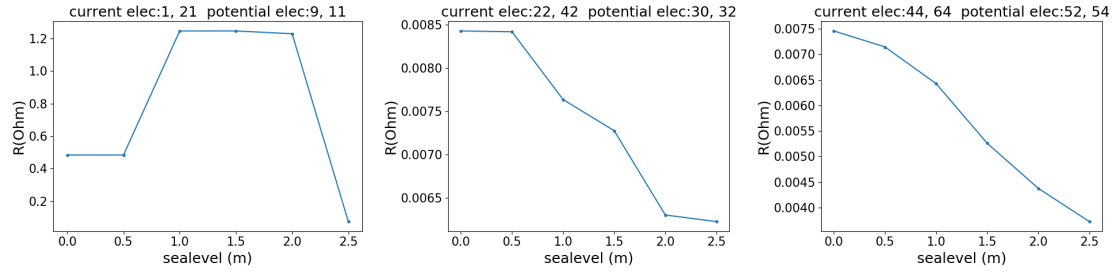


Figure A.2: Resistance variations of the synthetic model at level 2.

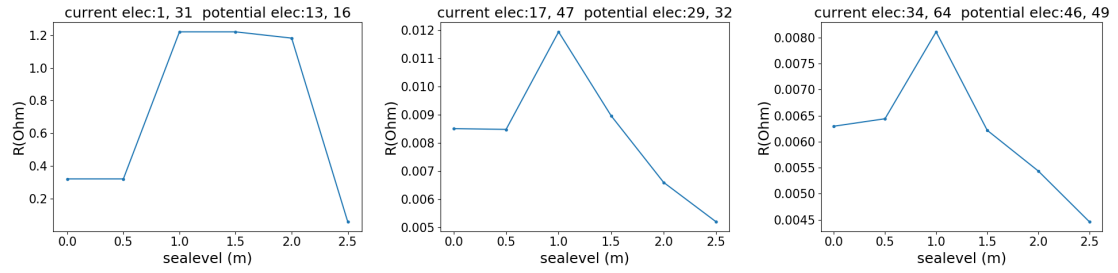


Figure A.3: Resistance variations of the synthetic model at level 3.

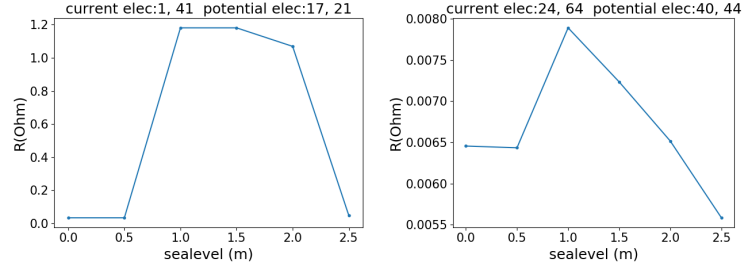


Figure A.4: Resistance variations of the synthetic model at level 4.

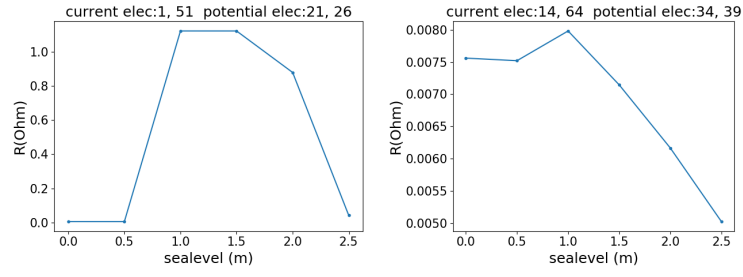


Figure A.5: Resistance variations of the synthetic model at level 5.

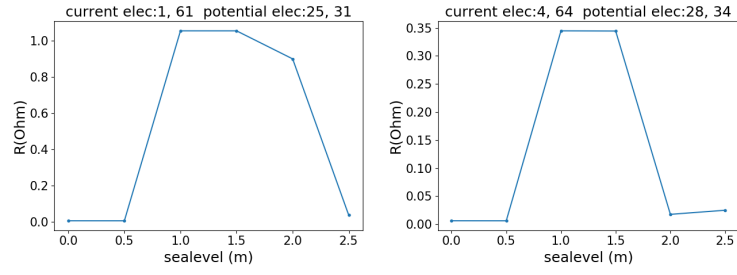


Figure A.6: Resistance variations of the synthetic model at level 6.

Appendix B

Synthetic models: inversion results

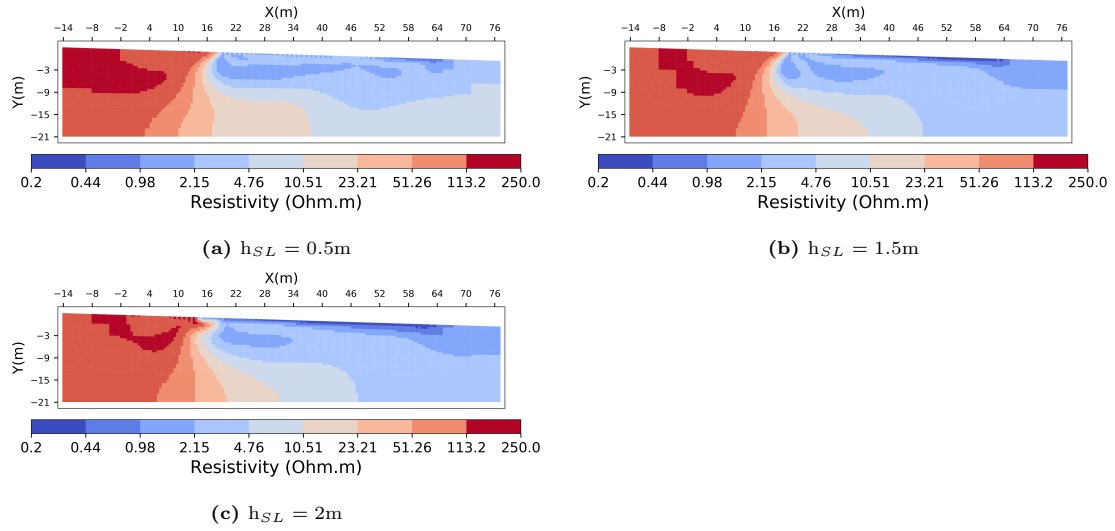


Figure B.1: Inversion with reference grid, additional figures.

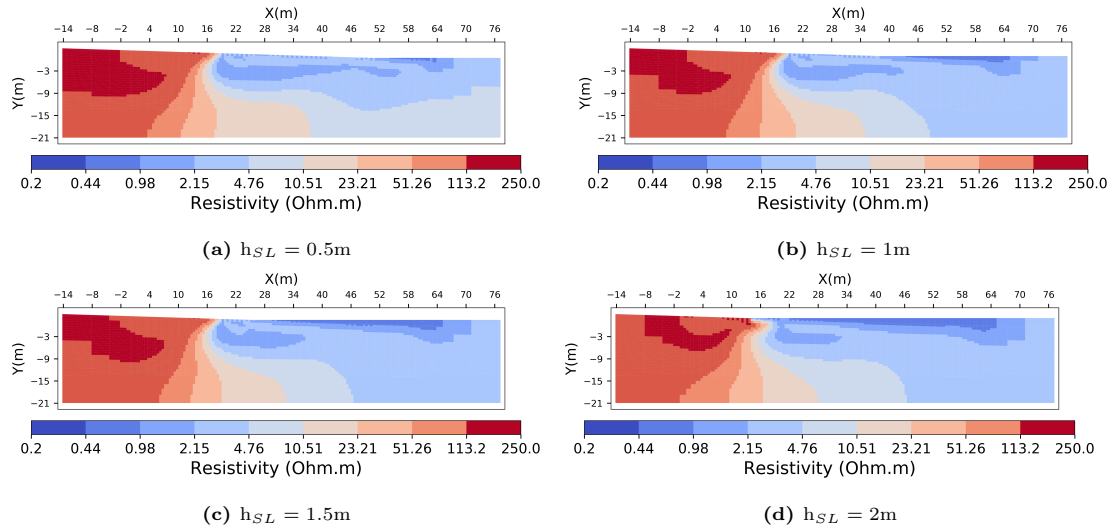


Figure B.2: Inversion with correct grid, additional figures.

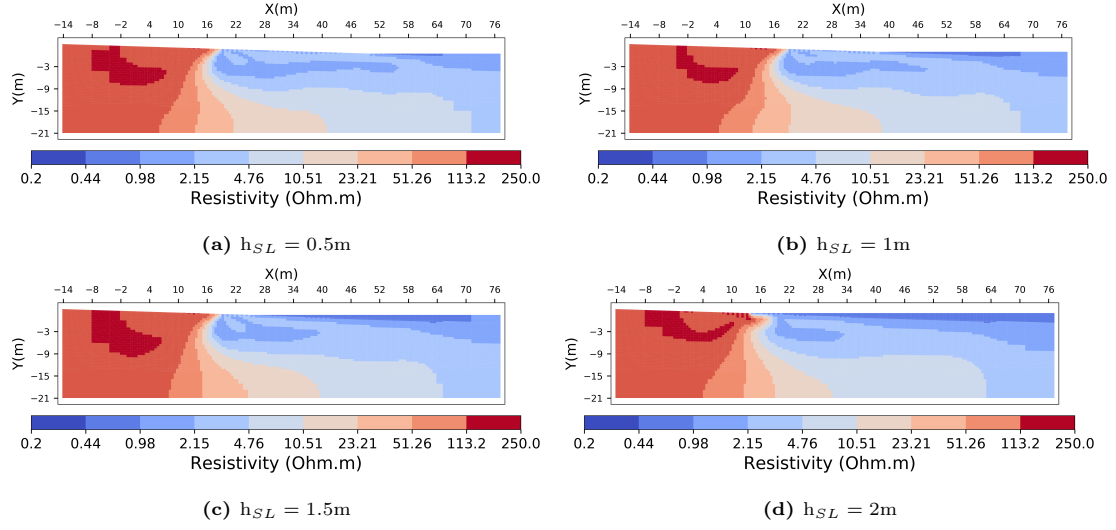


Figure B.3: Inversion with a homogeneous reference model, additional figures.

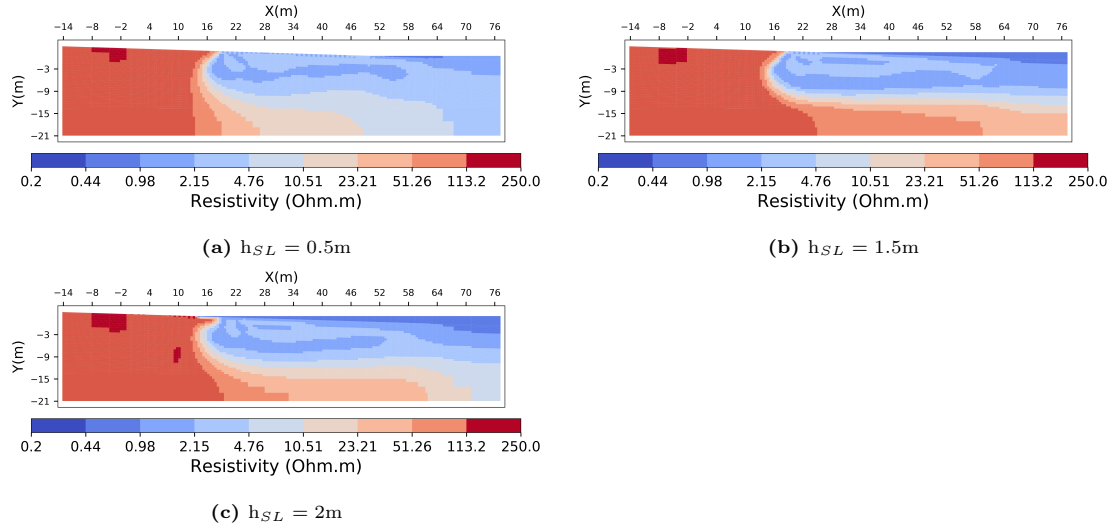


Figure B.4: Inversion with a realistic reference model, additional figures.

Appendix C

Resistance variations field data

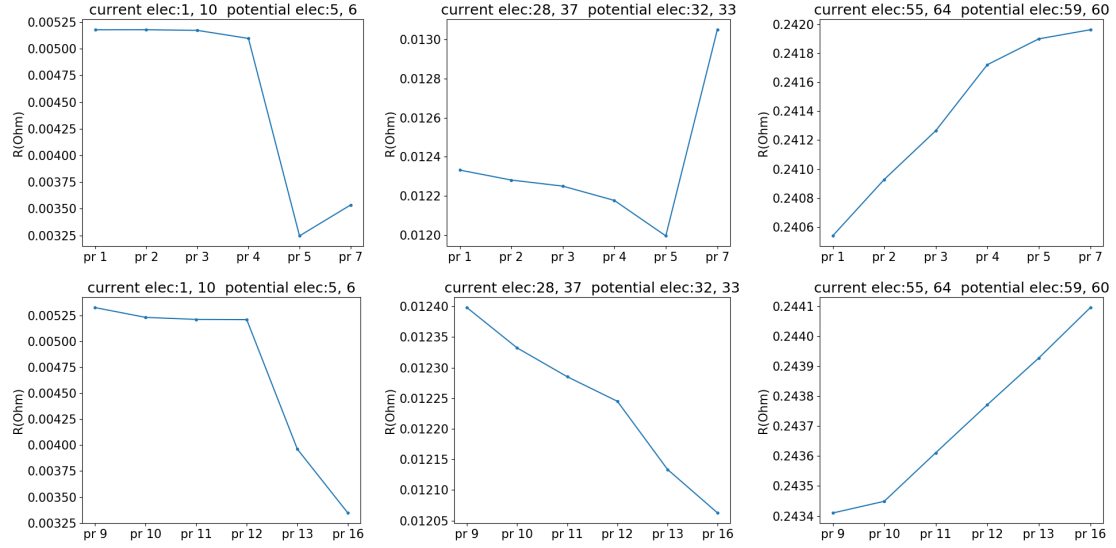


Figure C.1: Resistance variations of the long profile at level 1.

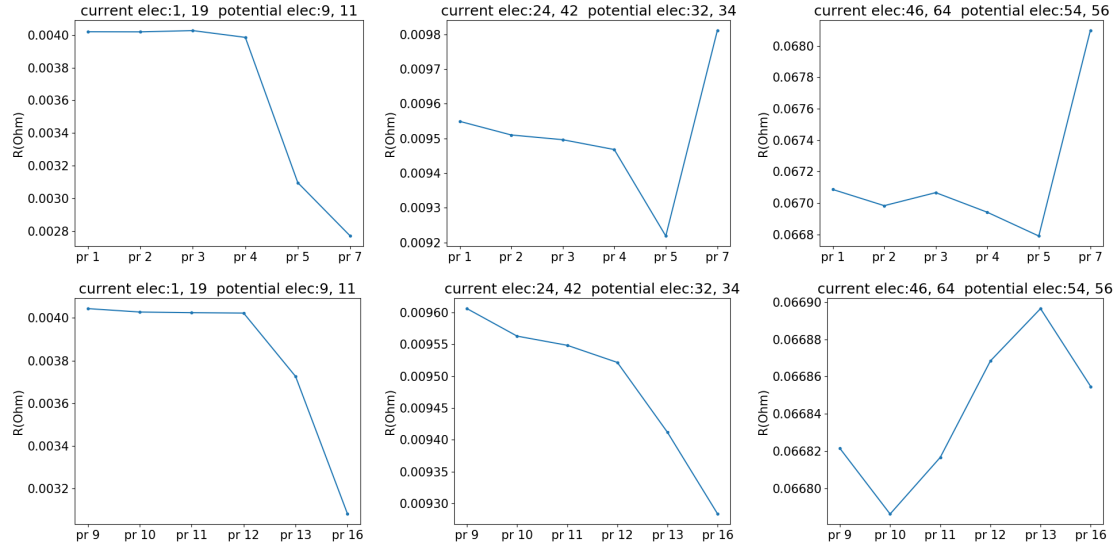


Figure C.2: Resistance variations of the long profile at level 2.

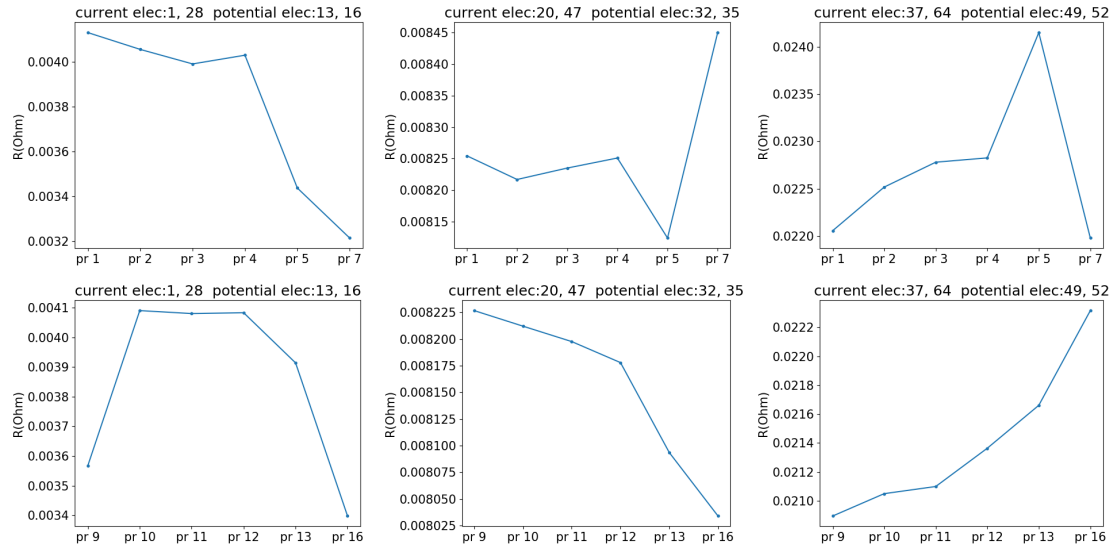


Figure C.3: Resistance variations of the long profile at level 3.

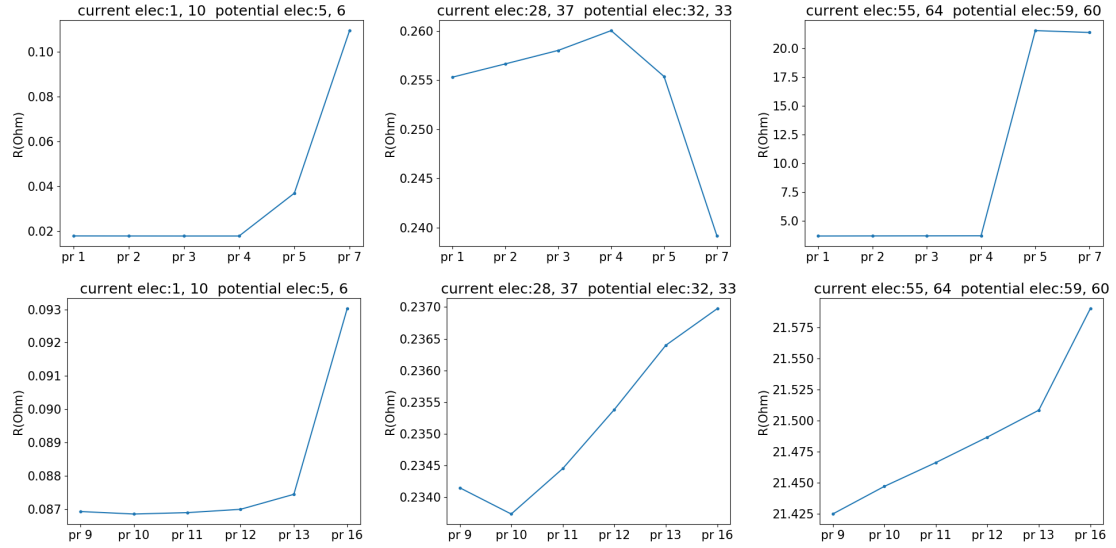


Figure C.4: Resistance variations of the short profile at level 1.

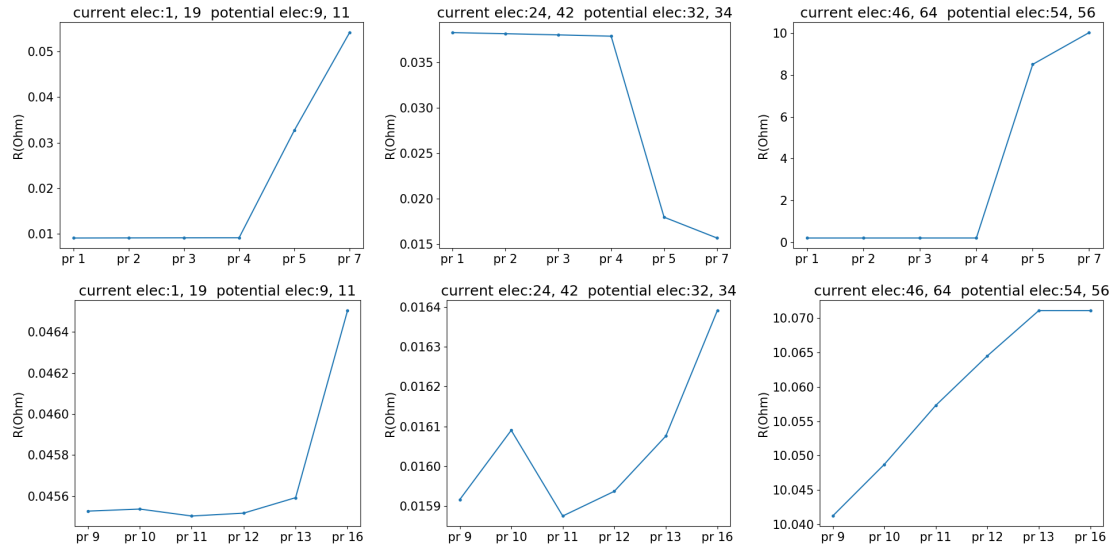


Figure C.5: Resistance variations of the short profile at level 2.

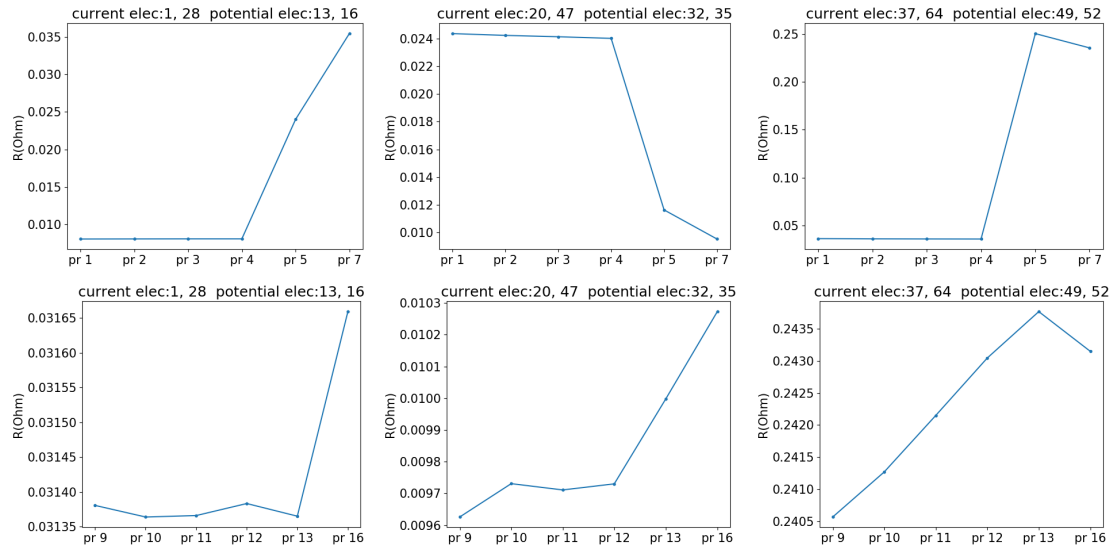


Figure C.6: Resistance variations of the short profile at level 3.

Appendix D

Reciprocal error: model reconstruction

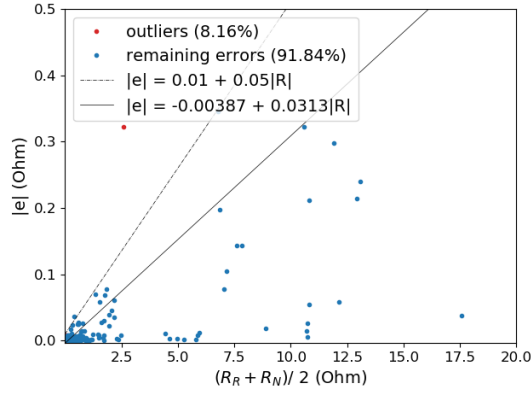


Figure D.1: Long profile error model, 10 % threshold.

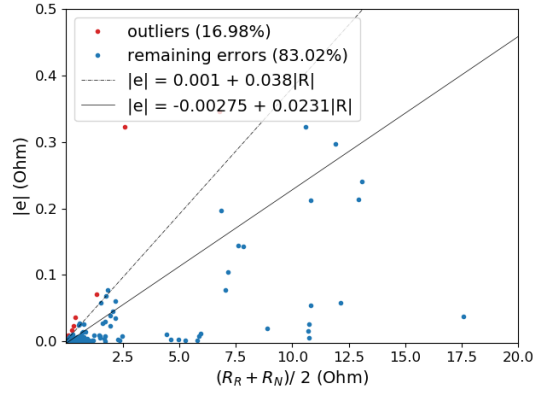


Figure D.2: Long profile error model, 5 % threshold.

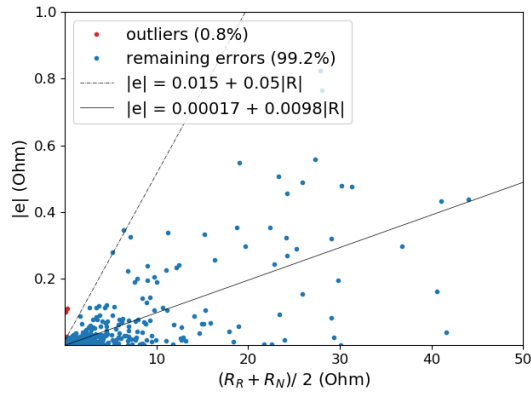


Figure D.3: Short profile error model, 10 % threshold.

Appendix E

Field data: inversion with homogeneous reference model

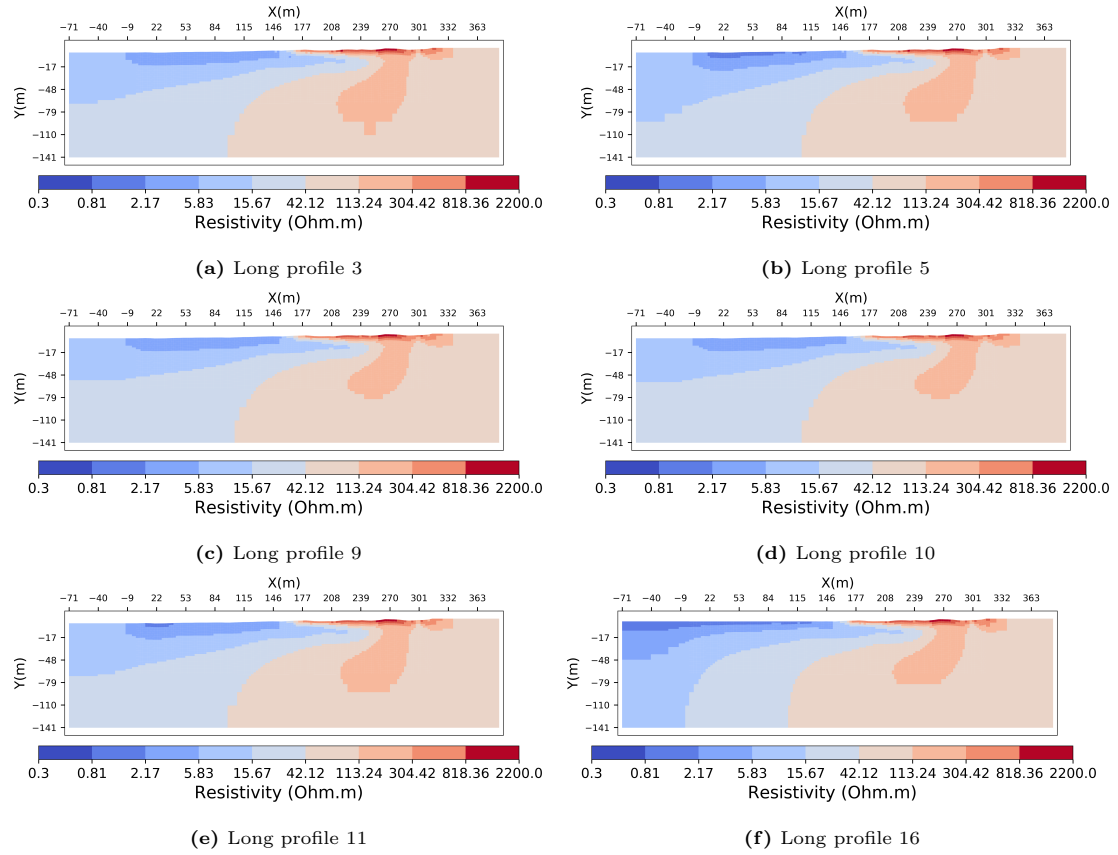


Figure E.1: Inversion results of the long profiles using a homogeneous reference model, additional figures.

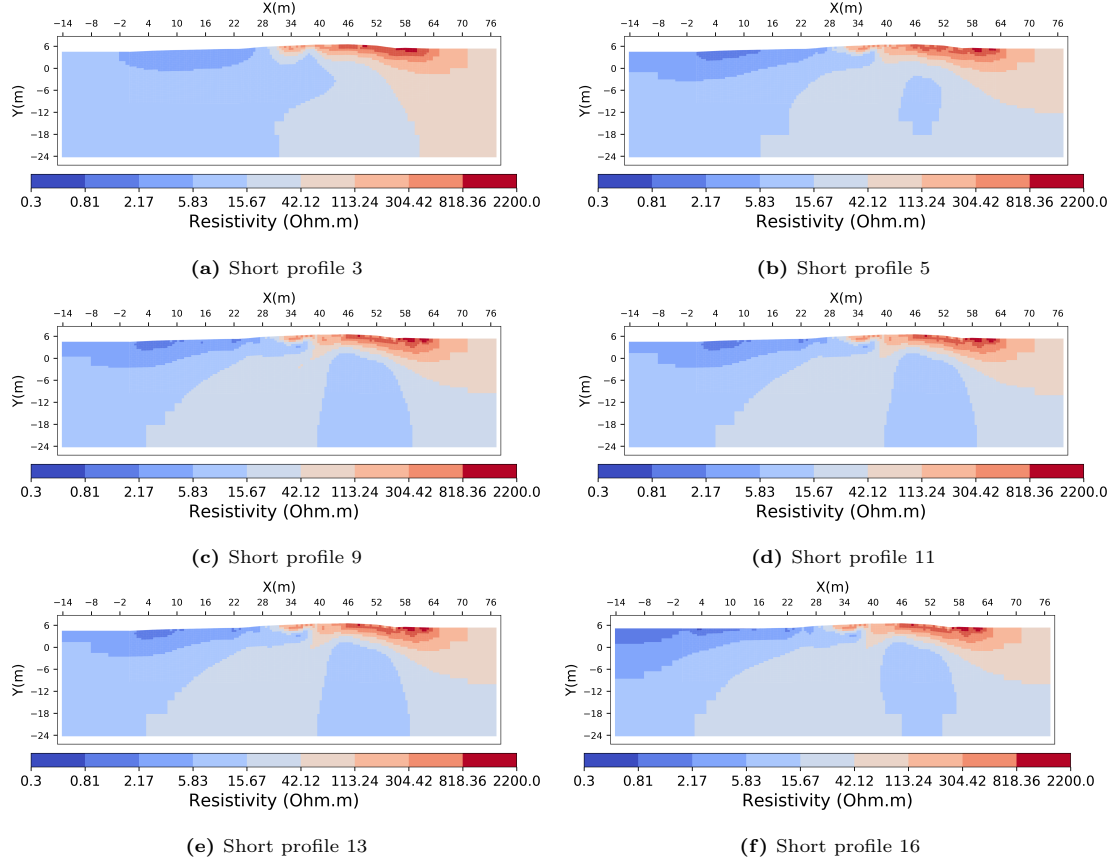


Figure E.2: Inversion results of the short profiles using a homogeneous reference model, additional figures.

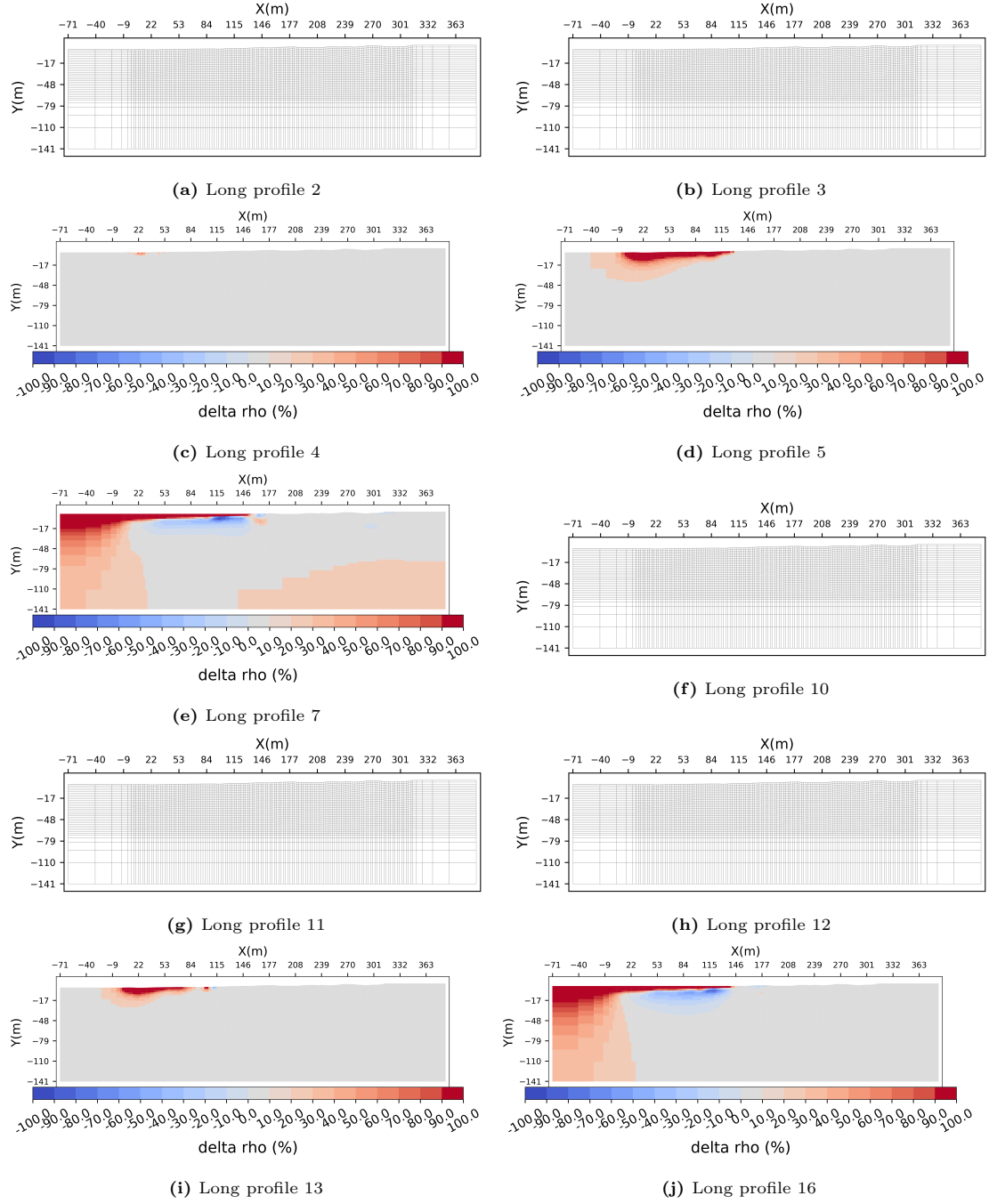


Figure E.3: Delta rho of the long profiles, the difference is calculated with the first profile of each cycle as background for the remaining profiles, i.e. profile 1 is the background for profiles 2-7 and profile 9 for 10-16.

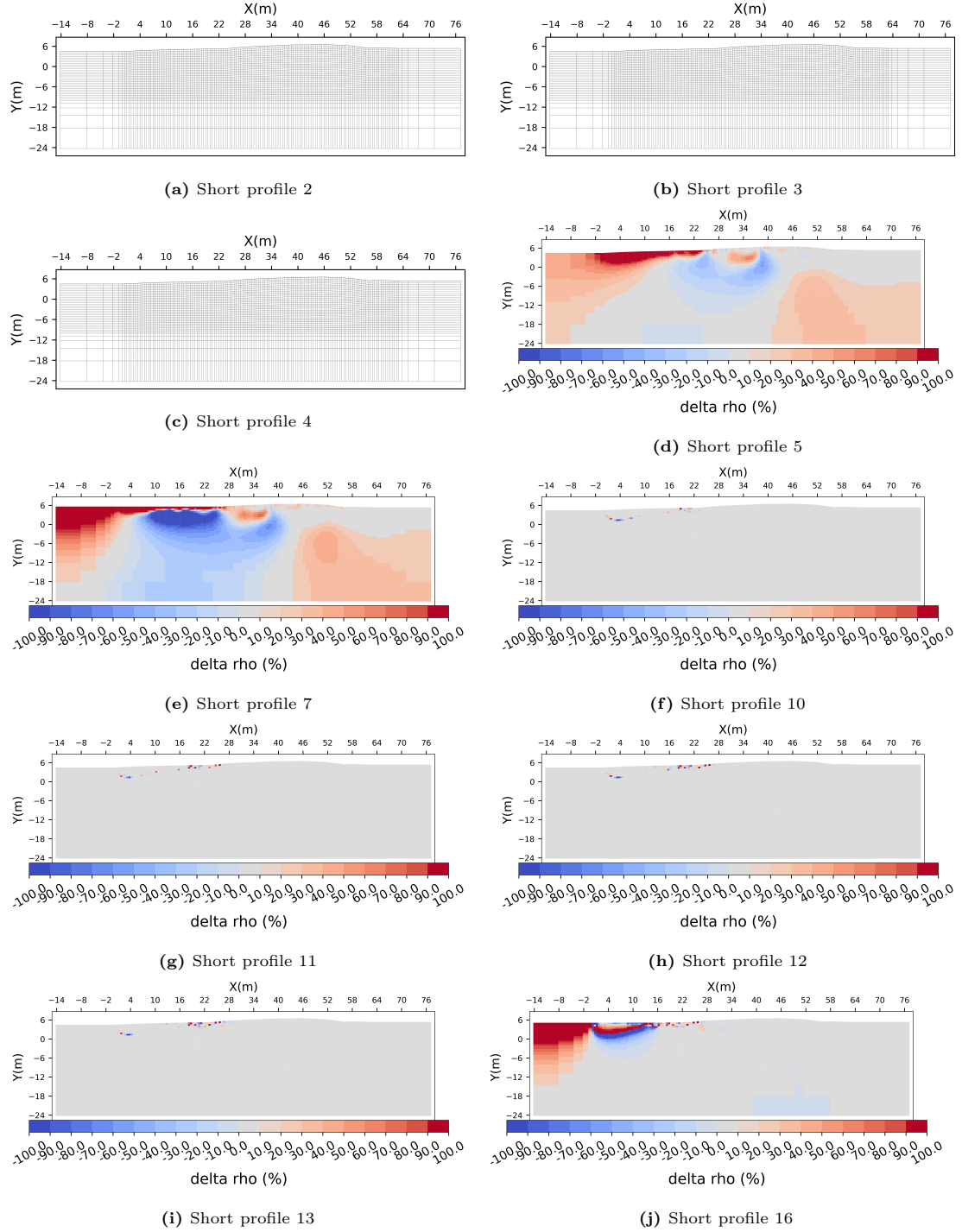


Figure E.4: Delta rho of the short profiles, the difference is calculated with the first profile of each cycle as background for the remaining profiles, i.e. profile 1 is the background for profiles 2-7 and profile 9 for 10-16.

Appendix F

Field data: time-constrained inversion

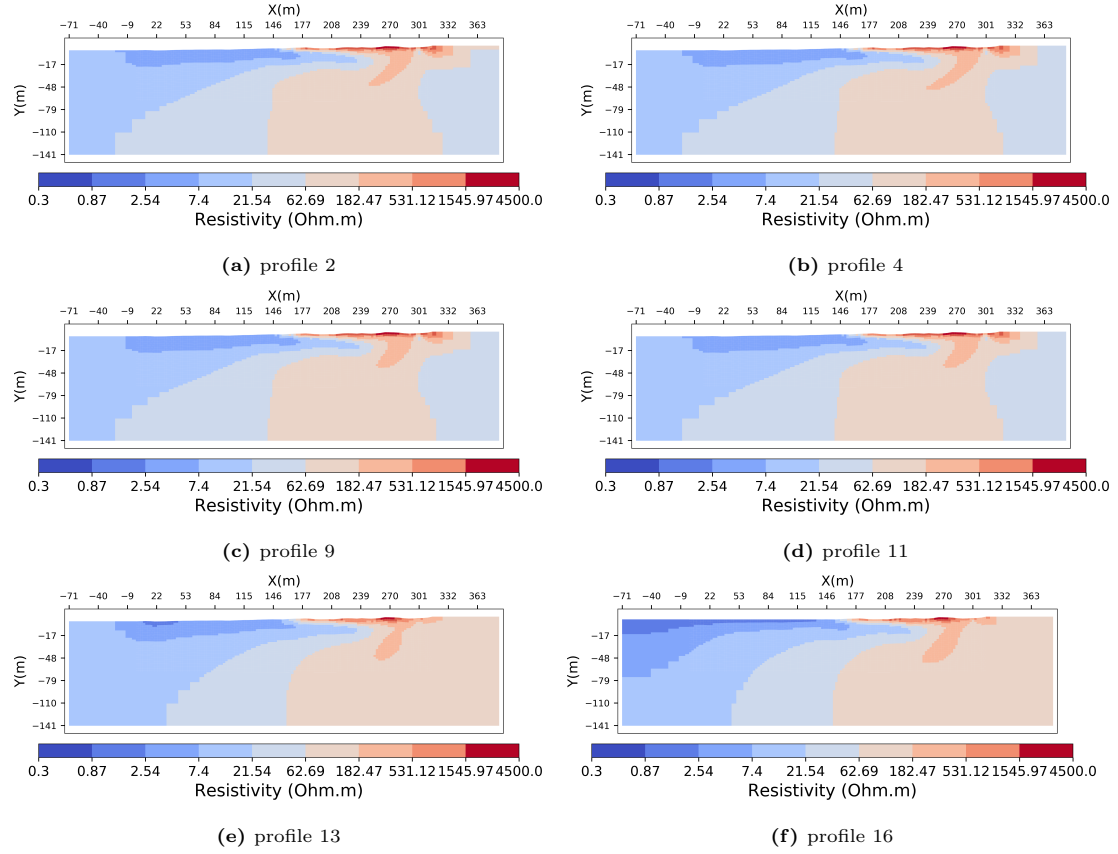


Figure F.1: Results of the time constrained inversion for the long profile, additional figures.

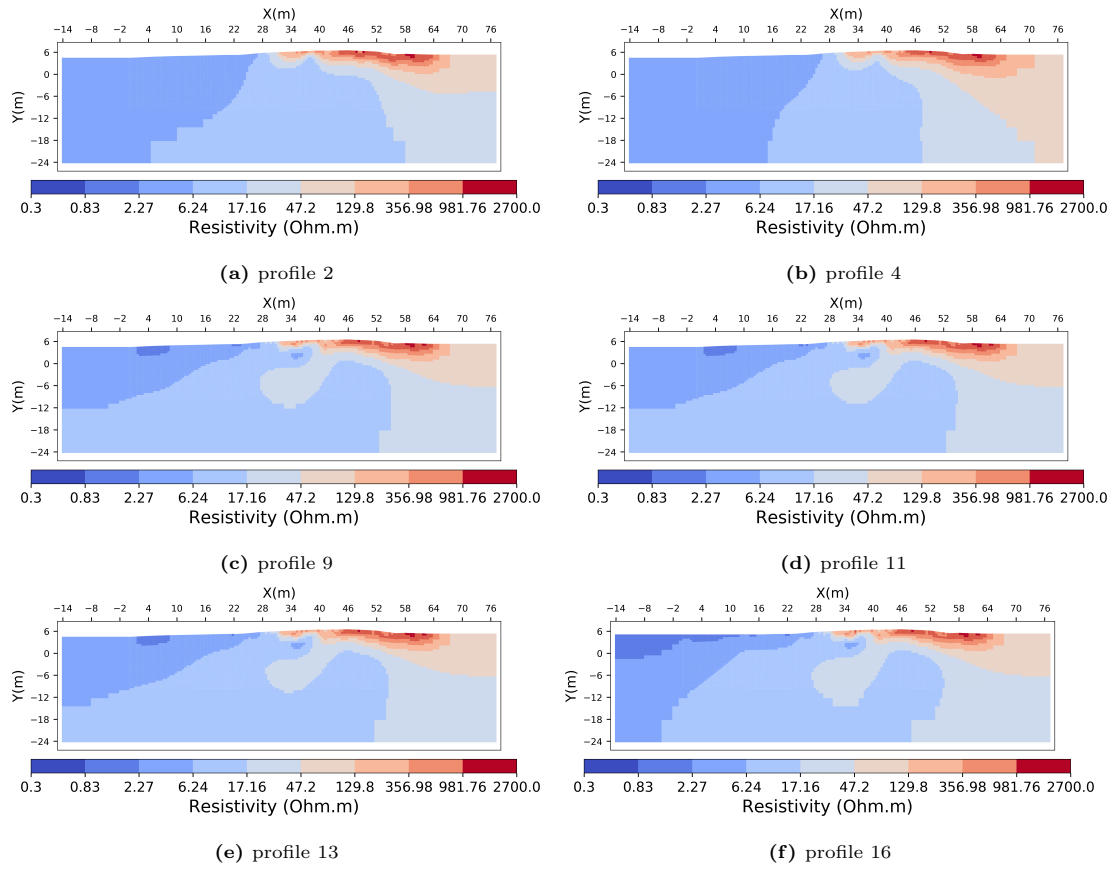


Figure F.2: Results of the time constrained inversion for the short profile, additional figures.

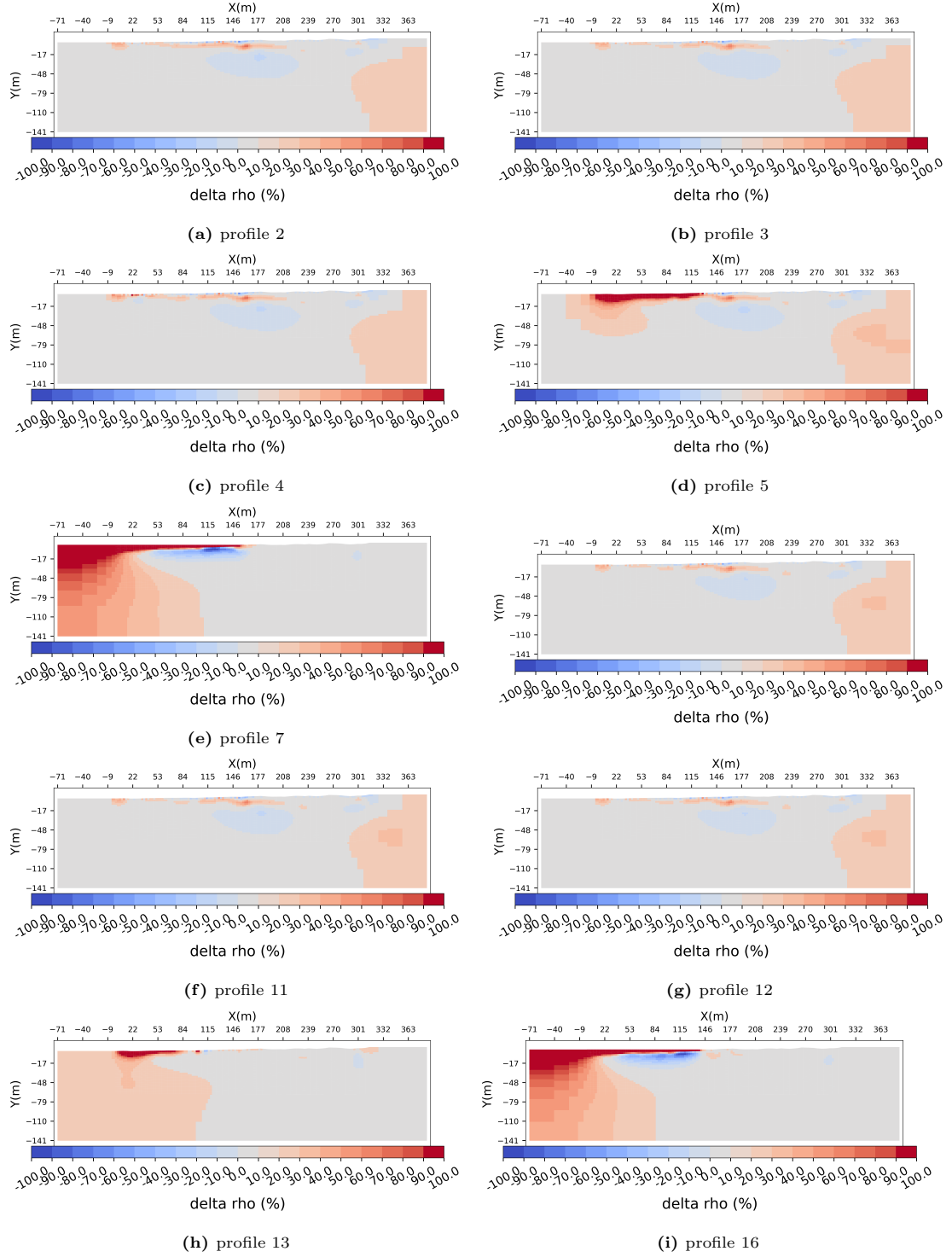


Figure F.3: Delta rho of the long profiles for the time constrained inversion, the difference is calculated with reference to the prior model.

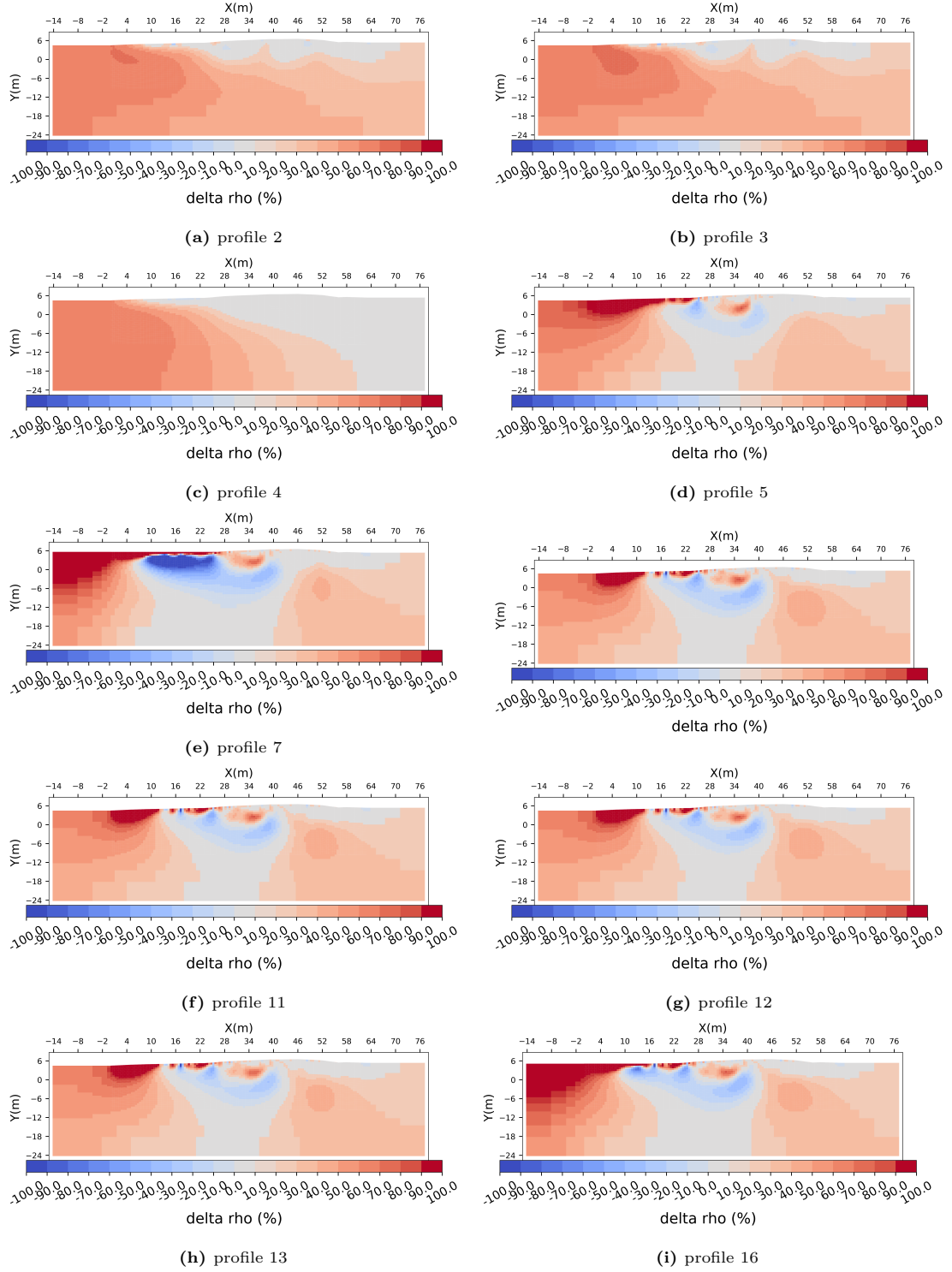


Figure F.4: Delta rho of the short profiles for the time constrained inversion, the difference is calculated with reference to the prior model.