Bedforms and their effect on sediment transport on ebb-tidal deltas

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Bedforms and their effect on sediment transport on ebb-tidal deltas

Bodemvormen en hun invloed op sedimenttransport op buitendelta's

(met een samenvatting in het Nederlands)

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“Gelukkig is degene die zijn kennis heeft vergaard door onderzoek.”
Euripides 910

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Samenvatting

De Waddenkust is een hoog-dynamisch ecosysteem, dat niet alleen belangrijk is voor ecologie, maar ook mogelijkheden biedt voor toerisme en daarmee de economie. De Waddenkust bestaat uit meerdere duidelijk herkenbare morfologische elementen, zoals de Waddeneilanden, getijdenbekkens en zeegaten, die de Waddenzee verbinden met de Noordzee. Aan de zeewaartse kant van een zeegat bevindt zich vaak een buitendelta, ook wel ebdelta genoemd, die wordt gekenmerkt door diepe geulen en een of meerdere ondiepe zandplaten. De buitendelta staat onder invloed van zowel golven als getij, en daarom verandert deze voortdurend van vorm.

Veel van de buitendelta's in het Nederlandse deel van de Waddenzee verliezen zand, terwijl het zandvolume in de Waddenzee juist toeneemt. Het zandverlies van de buitendelta's is onwenselijk, omdat de ondiepe delen van de buitendelta's belangrijke filters zijn voor golf-energie, waardoor de achterliggende bekkens beschermd worden tegen stormgolven. Bovendien hebben de getijdebekken in de Waddenzee steeds meer zand nodig om mee te kunnen stijgen met de versneld stijgende zeespiegel, waardoor er in de toekomst een stijging in erosie van de buitendelta's wordt verwacht. Daarom wordt er vanuit de kustbeherende instanties overwogen om de buitendelta's te suppleren, om zo het volume in de delta's terug te brengen en tegelijkertijd zand via natuurlijke processen de Waddenzee in te laten gaan. Voor het ontwerp en de voorspelling van het gedrag van zulke suppleties worden echter modellen gebruikt, die vaak niet zijn getest voor de golf- en stromingsgedomineerde condities op de buitendelta's. Daarom is het belangrijk om de patronen van zandtransport op buitendelta's beter te snappen, om zo een goede wetenschappelijke basis te leggen voor het gebruik van modellen in het kustbeheer.

Omdat buitendelta's zo dynamisch zijn, kunnen er veel soorten bodemvormen gevonden worden, op verschillende tijd- en ruimteschalen. Deze bodemvormen kunnen het zandtransport op verschillende manieren beïnvloeden; zowel direct, door te migreren en daarbij zand te verplaatsen, als indirect, door hun invloed op de structuur van de stroming van het water en daarmee op de hoeveelheid zand die wordt opgewoeld en getransporteerd. Het is nog niet goed bekend hoe dynamisch bodemvormen op buitendelta's zijn, en hoe hun aanwezigheid en kenmerken zijn gerelateerd aan de lokale golf-stroomgedomineerde hydrodynamica en zandtransport. In numerieke modellen van zeegatsystemen worden de verschillende bodemvormen ieder op een andere manier gerepresenteerd. Zaagtandbanken zijn groot genoeg om opgenomen te kunnen worden in de bathymetrie van het model. Megaribbels en kleinschalige ribbels zijn kleiner dan het rekenrooster van het model, dus hun effect op de hydrodynamica en het sedimenttransport wordt normaliter geparameteriseerd, maar het is niet bekend hoe accuraat deze parameterisaties zijn op buitendelta's. Daarom zijn de drie hoofddoelen van deze thesis het beschrijven van de kenmerken van de verschillende bodemvormen op buitendelta's, het vergelijken van

de kenmerken met bestaande voorspellers en het bestuderen van de gevoeligheid van sedimenttransportvoorspellingen voor de keuzes in de bodemvormvoorspellers.

In deze thesis worden verschillende datasets gebruikt: bestaande bathymetrische kaarten van meerdere Nederlandse en Duitse buitendelta's, hydrodynamische data en bodemvorm-data zoals gemeten tijdens de SEAWAD meetcampagne op de Amelandse buitendelta (Sediment supply At the Wadden sea ebb-tidal Deltas), en het numerieke model Delft3D. In deze thesis worden de bodemvormen beschouwd op drie ruimtelijke schalen. De grootste bodemvormen op buitendelta's zijn zaagtandbanken, die zich bevinden op het benedenstroomse (oostelijke) deel van de delta. Omdat ze zo groot zijn, zijn ze zichtbaar op bathymetrische kaarten, die zijn gebaseerd op singlebeam echosounding. Zulke kaarten zijn met enige regelmaat gemaakt tussen 1970 en 2015 voor de meeste zeegaten in het Nederlandse en Duitse deel van de Waddenzee. In Hoofdstuk 2 worden de kenmerken en het gedrag van de zaagtandbanken onderzocht bij 9 zeegaten. Gemiddeld zijn de zaagtandbanken 0.74 m hoog en 670 m lang, en ze migreren met 30 m per jaar, maar er zijn grote variaties rond deze gemiddelde waarden (hoogtes 0.5-2 m, lengtes 300-900 m, migratie 0-112 m per jaar). Zaagtandbanken transporteren veel sediment door te migreren, gebaseerd op hun zandvolume en migratiesnelheid, maar vergeleken met het totale transport langs de kust zijn deze hoeveelheden klein (gemiddeld 0.5-2%, met maxima van 15%). Hoge zaagtandbanken migreren sneller dan lagere. Dit onderscheidt zaagtandbanken van andere grote bodemvormen, die vaak langzamer migreren zodra ze hoger zijn.

Megaribbels en zandgoeven zijn bodemvormen met een veel kleinere schaal dan zaagtandbanken en hebben typisch lengtes van een paar meter tot 100 m. Vanwege een gebrek aan data zijn zandgolven in deze thesis niet bestudeerd. Megaribbels hebben een lengte kleiner dan 50 m, en zijn dus niet zichtbaar op de Vaklodingen van Rijkswaterstaat. Tijdens de SEAWAD meetcampagne werden megaribbel karakteristieken bepaald aan de hand van drie multibeam datasets: (1) vier gebieden dwars over het Amelandse zeegat zijn meerdere keren gemeten gedurende een getijcyclus en gedurende een springtij-doodtij cyclus, (2) jaarlijkse opnames van de oostelijke kant van de hoofdgeul 2005 en 2016, en (3) voor en na de SEAWAD meetcampagnes zijn er oppervlaktes rondom de meetinstrumenten ingemeten. De gemiddelde megaribbel in het Amelandse zeegat was 11.2 m lang (standaarddeviatie van 6.7 m) en 0.46 m hoog (standaarddeviatie van 0.17 m). De geulgemiddelde migratie van megaribbels gedurende een halve getijcyclus was 0.45 m in de richting van de ebstroming, maar over het gemeten transect varieerde de migratie tussen 2 m in de richting van de vloedstroming en 4 m in de richting van de ebstroming. Door de relatief hoge migratiesnelheden transporteren megaribbels ruim tien keer zoveel zand als de zaagtandbanken, maar doordat hun netto migratie gedurende een jaar onbekend is (dagelijkse waarden zijn hier opgeschaald naar jaarlijkse waarden), is ook het jaarlijkse zandtransport onzeker. Hoewel de bestudeerde gebieden werden gedomineerd door ofwel alleen stroming, ofwel door zowel golven en stroming, waren alle megaribbel hoogtes en vormen alleen afhankelijk van de stroming, en meer specifiek van de eb- of vloeddominantie. Wanneer ofwel eb ofwel vloed duidelijk sterker is, worden de megaribbels hoger en meer asymmetrisch. Deze observaties zijn hoogstwaarschijnlijk representatief voor andere zeegaten en stromings- of golf- en stromingsgedomineerde gebieden.

Kleinschalige ribbels zijn de kleinste bodemvormen die voorkomen in zeegatsystemen, en deze zijn bestudeerd op meerdere locaties op de buitendelta van het Amelandse zeegat. Tijdens de SEAWAD meetcampagne werden vijf frames met op ieder frame een 3D Sonar

gedurende vier tot zes weken op de delta geplaatst (Hoofdstuk 4). De Sonar maakte elk uur een scan van de bodem onder het frame, terwijl andere apparatuur op hetzelfde frame de lokale hydrodynamica mat. Gedurende de periode van de metingen kwamen er stromingsgedomineerde, golfgedomineerde en golf- en stromingsgedomineerde condities voor. Desondanks waren de kleinschalige ribbels altijd zeer driedimensionaal en was er weinig variatie in ribbelhoogte en -lengte in zowel ruimte als tijd. De ribbels waren gemiddeld 0.015 m hoog en 0.11 m lang, met hoogtes variërend tussen 0.005 en 0.025 m en lengtes tussen 0.06 en 0.18 m. Er konden geen migratiesnelheden worden bepaald. De ribbelvormen waren alleen gerelateerd aan de orbitaalsnelheid bij de bodem: met toenemende orbitaalsnelheden nam de ribbelhoogte af. Waarschijnlijk wordt de lage variatie in ribbelhoogte en -lengte verklaard door de relatief kleine korrelgrootte van het zand in het Amelandse zeegat van 0.211 mm. Korrelgroottes op de andere buitendelta's in de Nederlandse en Duitse Waddenzee zijn ook relatief klein, dus waarschijnlijk zijn de ribbelkarakteristieken en hun gedrag op de buitendelta van het Amelandse zeegat representatief voor deze andere buitendeltas.

In Hoofdstukken 3 en 4 zijn de gemeten hoogtes van de megaribbels en de kleinschalige ribbels vergeleken met enkele vaak gebruikte ruwheidsvoorspellers. Zowel de voorspelde megaribbelruwheid als de voorspelde ribbelruwheid bleek lager te zijn dan de gemeten hoogtes, en de voorspelde ruwheid fluctueerde veel meer door de tijd heen dan gemeten. Het effect van deze parameterisatie op hydrodynamica en sedimenttransportpatronen is getest in Hoofdstuk 5 met behulp van een gevoeligheidsanalyse voor het gebruik van ruwheidsparameters van een Delft3D simulatie van de zes weken van de SEAWAD meetcampagne op de Amelandse buitendelta. De keuze voor de ruwheidsparameterisatie bleek een aanzienlijk effect te hebben op de gemodelleerde stroomsnelheid en het suspensie-transport. Kleine variaties in de ruwheidsparameterisatie kunnen leiden tot veranderingen van 10–20% in stroomsnelheid en meer dan 100% in het sedimenttransport. Deze effecten zijn voornamelijk belangrijk op de ondiepe zandplaten en dieper op zee; de stroming en het zandtransport in de diepe geul worden nauwelijks beïnvloed. De veranderingen in de ruimtelijke patronen van sedimenttransport geven aan dat de transportgradiënten ook zullen veranderen, wat zal leiden tot veranderingen in morfologische ontwikkeling van de delta als geheel. De grootste verschillen in transportgradiënten werden gevonden wanneer een constante Chézy ruwheidswaarde werd gebruikt in plaats van de standaard ruwheid in Delft3D die afhangt van bodemvormen.

Deze thesis heeft nieuwe inzichten gegeven in de directe en indirecte effecten van bodemvormen op buitendelta's op sedimenttransport op verschillende schalen in ruimte en tijd, maar veel andere processen in numerieke modellen worden ook geparametriseerd. Deze andere processen, zoals golf-stroom interactie, korrelgrootte en ecologie, hebben allemaal effect op de bodemvormen en vice versa, wat weer effect kan hebben op de sedimenttransportpatronen over de buitendelta. Op de langere termijn worden klimaatverandering en veranderingen in de bathymetrie ook belangrijk. Toekomstig onderzoek zou zich moeten richten op de effecten van bodemvormen op de grootschalige morfodynamiek en het effect van korrelgrootte op bodemvormen en sedimenttransport. Tot slot moet nog worden bestudeerd waarom de onderschatting van bodemvormruwheid in modellen nog steeds leidt tot bruikbare voorspellingen van hydrodynamica. Beter begrip van de modellen leidt tot beter onderbouwde voorspellingen van het gedrag van buitendeltasuppleties, wat uiteindelijk zal leiden tot een veiligere kust.

Summary

Barrier coasts are highly dynamic environments providing many important ecosystem services, such as offering opportunities for tourism and hence stimulating the local economy, and hosting a wide variety of ecosystems. Barrier coasts consist of several distinct morphologic features, such as barrier islands, a back-barrier basin and tidal inlets that connect the basin to the open sea or ocean. At the seaward side of the tidal inlet an ebb-tidal delta is often found, a region characterized by one or more deep channels and shallow shoals. Ebb-tidal deltas are subject to both waves and tidal forcing and their morphology is highly dynamic.

A well-known example of a barrier coast is the Wadden Sea in northwestern Europe. Many of the Dutch ebb-tidal deltas are decreasing in volume, while the Wadden Sea is gaining sediment. The decrease in volume might have negative effects, because the shallow parts of the ebb-tidal deltas are efficient filters of wave energy, protecting the hinterland against storm waves. In the future, an increasing amount of sediment is needed for the Wadden Sea basins to keep up with accelerating sea level rise, which might lead to even more erosion of the ebb-tidal deltas. For coastal management purposes, it has been proposed to nourish the ebb-tidal deltas, in order to restore the delta volume and to supply sediment to the Wadden Sea. Models that are used for the design and prediction of the behaviour of such nourishments have often not been tested for the mixed energy conditions of the ebb-tidal deltas. Thus, our understanding of the sediment transport patterns on ebb-tidal deltas should be improved, which will provide a solid science base for using numerical models in coastal management strategies.

Ebb-tidal deltas host a large variety of bedforms, on various spatial and temporal scales. These bedforms can affect the sediment transport pathways in several ways; directly by migrating and thus literally transporting sediment, and indirectly, by altering the flow structure and the sediment stirring. In numerical models of tidal inlet dynamics, the various bedform types are each treated in a different way. Saw-tooth bars are large enough to be incorporated in the model bathymetry. Megaripples and small-scale ripples are smaller than the grid cell size, so their effects on hydrodynamics and sediment transport are often parametrized, but the performance of these parametrizations on ebb-tidal deltas is unknown. Overall, our knowledge should be improved on the characteristics of bedforms on ebb-tidal deltas in space and time and how these are related to and affect local wave-current dominated hydrodynamics and sediment transport patterns and magnitudes. Therefore, the central issues of this thesis are to determine the spatio-temporal dynamics of bedforms on ebb-tidal deltas on a variety of spatial scales, to compare these characteristics to existing bedform roughness predictors and to determine how predicted sediment transport in mixed-energy environments is affected by predicted bedform roughness.

The data sets used in this thesis are based on existing bathymetric data at multiple Dutch and German ebb tidal deltas, hydrodynamic and sea bed data gathered in the framework of the SEAWAD (Sediment supply At the Wadden sea ebb-tidal Deltas) measurement campaign

at the Ameland ebb tidal delta, and the numerical model Delft3D. In this thesis, three spatial scales are considered. The largest bedforms on ebb-tidal deltas are saw-tooth bars, which are found on the downstream part of the delta. As they are so large, they are visible on bathymetric maps, which are based on singlebeam echosounding. Such maps were available for most inlets of the Dutch and German Wadden Sea, with regular soundings between 1970 and 2015. In Chapter 2, saw-tooth bar characteristics and behaviour were studied for 9 tidal inlets. On average, the saw-tooth bars were 0.74 m high and 670 m long, and they migrated with 30 m/y, but large ranges in height (0.5-2 m), length (300-900 m) and migration speed (0-112 m/y) were found. By migrating, saw-tooth bars transport a considerable amount of sediment, estimated from their sand volume and their migration speed, but this amount is still relatively small (0.5–2% on average, with maxima of up to 15%) compared to the total longshore sediment transport. Higher saw-tooth bars migrated faster than lower bars, which distinguishes them from other large-scale bedforms, which often move slower if they are higher.

On the medium spatial scale, megaripples and sand waves can be found in tidal inlets and on ebb-tidal deltas. Because of a lack of data, sand waves were not further studied in this thesis. Since megaripples are smaller than 50 m in length, they are not resolved in the gridded and interpolated singlebeam-based bathymetry. Megaripple characteristics were quantified at the Ameland inlet and ebb-tidal delta from three multibeam echosounding data sets: (1) four areas across the tidal inlet were measured multiple times during a tidal cycle and during a spring-neap tidal cycle, (2) a part of the inlet channel was measured yearly between 2005 and 2016, and (3) before and after the SEAWAD campaign areas around the measurement frames were measured. The average megaripple in the Ameland inlet was 11.2 m long (standard deviation of 6.7 m) and 0.46 m high (standard deviation of 0.17 m). The average migration across the inlet channel during half a tidal cycle (from peak flood to peak ebb) was 0.45 m in the direction of the ebb, but across the inlet channel net migration values range between 2 m in the direction of the flood and 4 m in the direction of the ebb. Because of the relatively high migration speeds, megaripples transport more sand than the saw-tooth bars, but yearly migration rates are unknown (daily rates were extrapolated to yearly values) so megaripple-related transport is highly uncertain. Even though the studied areas were dominated either by currents only or by waves and currents together, all megaripple heights and shapes were only dependent on the current velocity, and more specifically the difference between peak ebb and peak flood currents. When either the ebb or the flood current is stronger, so the flow is more unidirectional, the megaripples become higher and more asymmetric. These observations are likely to be representative of many other tidal inlets and current- or wave-current-dominated environments as well.

Small-scale ripples are the smallest bedforms that can be found in tidal inlets, and they were studied on several locations on the Ameland ebb-tidal delta. During the SEAWAD field campaign, five frames equipped with a 3D profiling Sonar were placed on the delta for four to six weeks (Chapter 4). Apart from the Sonar, which made images of the bed once per hour, hydrodynamics were measured as well. Over the course of the studied period, the conditions were current-dominated, wave-dominated and mixed wave-current dominated conditions occurred. Despite this variability, the small-scale ripples were always highly three-dimensional, and the spatio-temporal variation in ripple heights and lengths was small as well. On average, the ripples were 0.015 m high and 0.11 m long, while the range in heights was between 0.005 and 0.025 m and the range in lengths was between 0.06 and 0.18 m. Mi-

gration could not be determined. The ripple dimensions depended only on the near-bed orbital velocity: with increasing orbital velocities, the ripple height decreased. It was hypothesized that the small variability in ripple dimensions was caused by the relatively small grain size in the Ameland inlet (about 0.21 mm). Grain sizes of the other ebb-tidal deltas in the Dutch and German Wadden Sea are relatively small as well, so it is expected that the ripple characteristics and dynamics on the Ameland ebb-tidal delta are representative of these other deltas.

In Chapters 3 and 4, the measured megaripple and small-scale ripple heights were compared to some common roughness predictors. It was found that both the predicted megaripple- and ripple-related roughness height are lower than the measured bedform height, which is related to the roughness, and that the predicted roughness heights show much more variation in time than was observed in the measurements. The effect of this parametrization on hydrodynamics and sediment transport patterns was tested in Chapter 5 by means of a sensitivity study in Delft3D for the six-week SEAWAD field period at the Ameland ebb-tidal delta. It was found that the choice of roughness parametrization has considerable effects on the modeled current velocity and the suspended load transport. While the current magnitude changes by 10–20% through changes in bedform roughness, the sediment transport magnitude changes by more than 100%. These effects are mainly important on the shallow shoals and offshore; the current and transport in the inlet channel are hardly affected. The changes in the spatial patterns of sediment transport indicate that the transport gradients will change as well. The largest differences with the default Delft3D bedform roughness were found when a constant Chézy value was used, which will also affect the long-term morphological evolution of the ebb-tidal delta.

This thesis has provided insight on the spatio-temporal variability of bedforms on ebb-tidal deltas at a variety of scales and on their indirect and direct effects on sediment transport, but many other processes are parametrized in numerical models as well. These processes, including wave-current interaction, grain size and ecology, all interact with the bedforms and vice versa, which might affect the sediment transport pathways across the ebb-tidal delta. On the longer term, climate change and bathymetric changes become important as well. Future research should address the effects of bedforms on morphodynamics, as well as the influence of grain size on bedform characteristics and sediment transport. Lastly, it should be studied why the underprediction of bedform roughness in models still leads to adequate hydrodynamic predictions. By doing so, the behaviour of nourishments on ebb-tidal deltas can be predicted with more confidence, which will eventually improve coastal safety.

Chapter 1 | Introduction

1.1 Barrier coasts, ebb-tidal deltas and coastal management

Nearly a third of all ice-free coasts are sandy (Luijendijk et al., 2018), and approximately 10% of the world's coastlines are barrier coasts (Stutz and Pilkey, 2011). Examples of sandy barrier coasts are the Wadden Sea in northwestern Europe, the Malpeque barrier chain in southeast Canada and most parts of the Atlantic coast of the United States of America (Figure 1.1). As ecosystems, barrier coasts provide important services to their societal and natural subsystems (Barbier et al., 2011; Moser et al., 2012). Regarding ecology, barrier coasts are an important stopover area for birds (Skagen et al., 1999), while salt gradients in the water provide a unique environment for both plants and animals (Pennings and Bertness, 2001; Więski et al., 2009). This is reflected in the declaration of the Wadden Sea as Unesco World Heritage site, and many other barrier coasts being declared as national parks. On the other hand, tourism and harbours impose economic values as well, which puts stress on the ecological system (Elias, 2006; Foster et al., 2009; Oost et al., 2012). Furthermore, human interventions have strong impact on the dynamics of tidal inlet systems, modifying the natural morphological evolution of barrier systems (Oost et al., 2012; Elias et al., 2012; Ridderinkhof et al., 2016b; Wang et al., 2018). Both the ecological and the societal subsystem of the barrier coasts are under pressure because of climate change, which results in rising sea levels, increasing storm frequency and intensity and increasing temperatures (Więski et al., 2009; Feagin et al., 2010; Barbier et al., 2011; Moser et al., 2012). In order to manage and preserve barrier coasts, a solid knowledge base is required about their morphological behaviour and ecological functioning.

In general, most barrier coasts consist of several to numerous tidal inlet systems, of which some morphologic features can be seen from space (Figure 1.1). A typical tidal inlet system with a dominant longshore drift direction, which is characteristic for the Dutch and East Frisian Wadden Islands, is schematized in Figure 1.2. It consists of barrier islands, which act as a barrier between the open sea and a shallow basin and are separated by tidal inlet channels. Further into the basin, the tidal channels decrease in size and bifurcate, which causes an alternation of channels and tidal flats. In the basin a flood-tidal delta may be present, while on the seaward side of the inlet channel, an ebb-tidal delta can be found (de Swart and Zimmerman, 2009).

The ebb-tidal delta is an important part of the tidal inlet system, and it consists of one or more shallow sand bodies called shoals, which are intersected by ebb-dominated or flood-dominated channels. These shoals can be several kilometers in size, containing sand volumes between $1 \cdot 10^5$ and $95 \cdot 10^6$ m³ (FitzGerald, 1988; Gaudio and Kana, 2001). Ebb-tidal deltas act as a filter for high storm waves, thus protecting the back-barrier basin and the island coasts from high energy (FitzGerald, 1988; Lenstra et al., 2019). On top of that, the large variety in depths and amounts of exposure to waves and currents provide habitats for many benthic species (Holzhauer et al., 2019). As shown in Figure 1.2, the ebb-tidal delta is affected by

Figure 1.1 Examples of two chains of barrier islands. A: barrier islands along the east coast of Georgia, USA, B: part of the Wadden Sea in northwestern Europe.

waves and both cross-shore and alongshore tidal currents. As a result, tidal inlet systems in general and ebb-tidal deltas in particular are morphologically highly dynamic.

This thesis will focus on the dynamics of the ebb-tidal deltas in the Wadden Sea, which is one of the largest chains of barrier islands in the world. It consists of 44 islands, along the coast of the Netherlands, Germany and Denmark. Many of the ebb-tidal deltas of the Wadden coast are losing volume, while the back-barrier basins are gaining volume (Elias et al., 2012). This is caused by human actions like the reduction of basin area (e.g. the closure of the Zuiderzee and Lauwerszee in the Netherlands) and the connection of the islands to the main land by dams (e.g. the islands of Sylt in Germany and Rømø in Denmark) (Elias et al., 2012). This causes the change of the tidal inlet systems towards a new equilibrium between the volume of the ebb-tidal deltas and the tidal prism. The sediment lost from the deltas is transported into the back barrier basins in order to achieve this equilibrium (Elias et al., 2012). Also, relative sea level rise causes extra accommodation space in the basin, which leads to sediment transport from the deltas to the basin (van der Spek, 1994; Wang et al., 2018). In the future, the amount of sediment import to compensate for both sea level rise and the adjustment to the reduction of the basin area processes should be added up in order to find the total sediment demand (Wang et al., 2018). Yet, even if the sea level rise is smaller than the critical rate for drowning, the ebb-tidal delta volumes are still limited, and their decreasing volumes also decrease their protective effect (Elias et al., 2012). This is a possible threat for their hinterland, so measures should be taken in order to prevent the ebb-tidal deltas from eroding.

The Dutch management authorities are required by law to maintain coastal safety in a sustainable way (Ministry of Transport, Public Works and Water Management, 1990), while also conserving the position of the coast line under sea level rise (Ministry of Transport, Public Works and Water Management, 2000) and enhancing spatial and environmental qualities (Tweede Kamer der Staten-Generaal, 2005). Yet, the dynamic nature of barrier coasts makes

Figure 1.2 Sketch of an idealized tidal inlet system with a dominant longshore drift direction, including the morphologic elements and the dominant hydrodynamic (blue arrows) and sediment transport processes (red arrows). Based on de Swart and Zimmerman (2009).

it difficult to perform lasting coastal management measures. In recent decades, coastal engineering has, where possible, stepped away from hard solutions like dikes, and has increasingly performed soft engineering works, like sand nourishments (Hamm et al., 2002). This approach is called Ecological Engineering, or Building with Nature, and it has the intention that the engineering measures work with nature instead of merely existing in it or working against it, thus causing less net disruption of the coastal ecosystem and developing valuable habitats (Mitsch and Jørgensen, 2003).

Sand nourishments are executed regularly, also on the beaches of the barrier islands (e.g. Grunnet and Ruessink, 2005; Gijssman et al., 2021). The decreasing ebb-tidal delta volumes cause a rising need to nourish the ebb-tidal deltas as well (Wang et al., 2018), to help both in restoring the delta volume and in preventing the back barrier basin from drowning by sea-level rise (Elias et al., 2020). In the latter case, sand that is added on the ebb-tidal delta should be transported into the basin by natural processes. However, it is not fully known how sand migrates in wave-current dominated environments with intermediate depths (Wang et al., 2012), and the fate of nourished sand is thus not a priori clear. Contrary to regular beach or shoreface nourishments in wave-dominated settings, sand on the ebb-tidal delta is affected by both waves and currents in a cross-shore and alongshore direction, as currents enter, leave and pass the tidal inlet. Thus, the behaviour of an ebb-tidal delta might be different from the behaviour of a shoreface nourishment, since they are both subject to different hydrodynamic processes. For example, Herrling and Winter (2018) found that sediment transport pathways can differ for different grain size fractions, with finer sediment bypassing the inlet from the updrift side towards the downdrift side, while the coarser sediment ($>250\ \mu\text{m}$) can recirculate over the shoals, and either move into the basin or migrate towards the coast of the downdrift island. Also, the addition of extra sand will affect the cyclic behaviour of the deltas and channels (Lenstra, 2020).

Numerical models like Delft3D are often used to predict the effect of coastal management measures like nourishments on morphologic behaviour (Wang et al., 2018). These models incorporate parameterizations of processes like wave-current interaction, bedform-related roughness and ecology. Yet, these parameterizations are often tested in wave-dominated or current-dominated conditions, and it is unclear if these predictions are valid in wave-current dominated areas with many depth variations like ebb-tidal deltas. Therefore, when the effect of an ebb-tidal delta nourishment is predicted using such a model, a considerable amount of uncertainty might be imposed on the results. Thus, in order to perform a nourishment with optimum behaviour for coastal protection, the underlying processes of sand transport on ebb-tidal deltas with mixed-energy conditions needs to be better understood (Elias et al., 2020).

1.2 SEAWAD

In order to design efficient and sustainable nourishments for the ebb-tidal deltas of the Wadden Sea, the research program SEAWAD (Sediment supply At the Wadden sea ebb-tidal Deltas) was set up. The main goals of this project were to identify the sediment pathways across a tidal inlet system and to develop system knowledge and tools to design mega nourishments on the ebb-tidal deltas of the Wadden Sea. The project was divided into four sub-projects:

1. Intra-wave sediment transport
This subproject aims to improve the prediction of wave-related sediment transport in the presence of complex tidal flows. The tidal currents through the inlet and parallel to the barrier coast can modify the wave field, causing current-induced refraction, shoaling and breaking. This can lead to significant changes in wave skewness and asymmetry, which are so far not accounted for in wave-averaged models.
2. Bedform dynamics on the ebb-tidal delta
This subproject aims to quantify the characteristics, dynamics and effects of bedforms on the ebb-tidal delta, on spatial scales of centimeters to kilometers and time scales of hours to years. Bedforms and their migration affect the hydrodynamics and sediment transport patterns, but their precise interaction in mixed-energy environments like ebb-tidal deltas is still unknown.
3. The importance of grain size on sediment transport pathways
This subproject aims to determine the sediment transport pathways for different grain size classes for natural and nourished ebb-tidal deltas. In an environment dominated by waves and currents, the sediment transport pathways are dependent on the grain size. In addition, the net sediment transport through the inlet is much smaller than the flood and ebb fluxes. This may lead to simultaneous import of fines and export of coarser grains, or vice versa (Herrling and Winter, 2014).
4. Eco-dynamic design of nourishments
This subproject studies how hydrodynamic and sediment parameters affect the response of benthic communities to nourishments. Ebb-tidal deltas are highly dynamic in terms of hydrodynamics and sediment transport, with large depth variations as well.

This causes a large variety of habitats for benthic communities. Identification of the relations between the physical conditions, sediment characteristics and the benthic species will help in reducing the impact of nourishments on the environment.

All research questions in the subprojects are answered by a combination of field data and modelling studies. The focus will be on the Wadden Sea ebb-tidal deltas in general, and on the Ameland ebb-tidal delta in particular. For this inlet, large historic and recent data sets are available. For example, the Dutch authorities have been measuring the bathymetry of this inlet since 1926. Also, this inlet is considered to be the most naturally behaving inlet of the Netherlands, without major ongoing human disturbance directly impacting the natural processes.

Field data were not yet available with high spatio-temporal resolution, so it was not yet possible to couple the hydrodynamics to bed states and sediment transport. Therefore, an extensive field campaign was set up and executed at the Ameland inlet in September and October 2017 (van Prooijen et al., 2020). Five stationary frames were placed in the inlet and near the western delta lobe, equipped with various instruments to measure the hydrodynamic and sediment and bed properties. Furthermore, grab samples were taken at 200 locations across the entire delta to measure bed composition and macrobenthos. Hydrodynamics were further measured by floating wave buoys and sailing measurements. Particle tracking was used to measure the pathways of the different grain size classes. Because all data sets were gathered simultaneously, it is now possible to study the interaction between hydrodynamics, sediment transport, morphology and benthic ecology in great detail. Furthermore, these field data can also be used to improve model parameterizations.

This thesis focuses on SEAWAD subproject 2, studying the direct and indirect effects of bedforms on sediment transport pathways and magnitudes on ebb-tidal deltas, and how well these are represented in numerical models. First, Section 1.3 will provide an overview of the existing knowledge on sediment transport pathways on ebb-tidal deltas and how they are affected by bedforms.

1.3 Sediment transport pathways and bedforms on ebb-tidal deltas

On the scale of the ebb-tidal delta, a lot of research has already been performed on the sediment transport pathways. The shoals typically show cyclic behaviour with time; in each cycle they grow, migrate, and then attach to the downdrift island (FitzGerald, 1996; Israel and Dunsbergen, 1999; Lenstra et al., 2019). This onshore movement occurs because landward flow is caused by waves and tides, and therefore it is dominant over seaward flow, which consists of only tides (Herrling and Winter, 2014; Lenstra et al., 2019; Ridderinkhof et al., 2016a). In the Wadden Sea area, ebb-tidal delta shoals migrate with a velocity of 67 to 351 m/y, and the cycle lengths are between 4 and 130 years (Ridderinkhof et al., 2016b). The cyclic migration of the shoals is a form of sediment bypassing in the direction of the wave-driven alongshore drift: the sediment in the shoals is transported across the inlet, from the updrift to the downdrift barrier island coast. The varying position of the shoals causes variations in tidal asymmetry, mean current and current magnitudes, which in turn affect the sediment transport pathways (Lenstra et al., 2019). When the inlet channel is directed updrift, the sediment transport across the shoal is high and takes place in the direction of the basin, while in other phases the mean sediment transport is smaller and directed towards

Figure 1.3 Sketch of a bedform, including some characteristics that can be used to describe it. Source: Motamedi et al. (2014)

the open sea (Lenstra et al., 2019). Furthermore, the transport magnitudes and directions are dependent on the grain size and weather conditions (Herrling and Winter, 2014). In calm weather, sediment is transported through the channels. On the other hand, during storms, fines bypass the tidal inlet while in suspension in wave-driven currents, whereas coarse sediment is transported towards the downstream coast across the shoals as bed load and in large bedforms (Herrling and Winter, 2014; Herrling and Winter, 2018).

The large variety of depths and wave- versus current dominance induces a wide variety of bedforms on ebb-tidal deltas, all with different spatial scales (Table 1.1). Often, various types of bedforms are found at the same location, with the smaller ones superimposed on the larger ones (Figure 1.5) (e.g. Bartholdy et al., 2005; Bellec et al., 2019; Jones and Traykovski, 2019). In this thesis, bedforms are defined as repetitive relief features which are generated by the interaction of water and a sandy bed. They can be characterized by several parameters, like height, wavelength or slip face angle (Figure 1.3). The wavelength of tidal and marine bedforms is usually taken crest-to-crest, because the crest is easiest to define. The trough is often more flat, so the 'actual' trough point can be difficult to determine. The bedform height is defined as the vertical distance between crest and trough.

The largest bedforms on ebb-tidal deltas are resolved in bathymetric images. Figure 1.4 shows that the shoals consist of several swash-bars (SWB in the figure). When waves break on the edges of the ebb-tidal delta, they rework the sand into swash bars (Hayes, 1980; Hine, 1975), which are 0.5-2.0 m high, 50-150 m long and 50 m wide, and migrate with velocities of 100-400 m/y (FitzGerald et al., 2011). The bar height increases in the onshore direction, which decreases the migration rate (FitzGerald, 1988). They merge together, forming large bar complexes and eventually even shoals (FitzGerald, 1982; Cheung et al., 2007). The volume of the main Ameland ebb-tidal delta shoal above 7 m below mean sea level is in the order of 50 million m³ (Cheung et al., 2007).

On the eastern, downdrift, part of the deltas in Figure 1.4, as well as on most other ebb-tidal deltas in the Dutch and German part of the Wadden Sea, shore-oblique saw-tooth bars are found (Reineck, 1963; Antia, 1994; Antia, 1996; Gordeau, 1999). Saw-tooth bar heights are in between 0.5 and 3 m, and their wave lengths are in between 400 and 1200 m (Reineck, 1963; Antia, 1994; Antia, 1996; Gordeau, 1999). Yet, these bars have only been studied for a restricted amount of years and only on a select number of tidal inlet systems of the Wadden Sea. It is unclear how fast saw-tooth bars migrate, how their characteristics evolve over multiple years, as well as how they are related to hydrodynamic forcing.

In the channels, current-dominated conditions create megaripples and sand waves (Figure 1.5). Sand waves, or large dunes, are formed by tidal currents, and are therefore not found in

Table 1.1 The bedform classification used in this thesis. Typical heights and lengths are given, as well as characteristic environments in which the bedforms are typically found, and available studies of the Ameland Inlet. In the penultimate column, w, c and wc denote wave-dominated, current-dominated, and wave-current dominated conditions, respectively. Question marks indicate that it is still unknown either where a bedform type occurs on an ebb-tidal delta, or if the environment is dominated by waves or currents.

Name	Height [m]	Length [m]	Environment [w/c/wc]	Study at the Ameland Inlet
Saw-tooth bar	0.5 - 3	400-1200	downstream ebb-delta [?]	Gordeau (1999)
Swash bar	0.5 - 2	50-100	ebb-delta shoal [w]	Cheung et al. (2007)
Sand wave/dune	1-10	10-800	channels [c]	-
Megaripple/dune	0.06-2	0.6-20	all [all]	Elias (2018) and van Rhijn (2019)
Wave-current ripple	0.01-0.1	0.1-1	? [wc]	-

wave-dominated environments (Hulscher, 1996). They are found in various tidal inlets, e.g. by Buijsman and Ridderinkhof (2008a) and Barnard et al. (2006). Therefore, it is expected that they will be present in the Ameland Inlet channel as well. Generally, sand waves have lengths in the order of hundreds of meters, heights in the order of meters and they migrate up to 100 m per year, thus causing direct sediment transport (Buijsman and Ridderinkhof, 2008a).

Megaripples have smaller heights and wavelengths than sand waves; their heights are 0.06 - 2 m and their lengths are in the order of 0.6 - 20 m (Idier et al., 2004). In some studies, megaripples are called dunes as well, following the classification of Ashley (1990) (e.g. van den Berg et al., 1995; Hoekstra et al., 2004; Bartholdy et al., 2005). They are found in various types of hydrodynamic environments, like intertidal shoals (Hoekstra et al., 2004), current-dominated channels (Elias, 2018) and the continental shelf (Passchier and Kleinhans, 2005). In the North Sea, it was found that their planform is three-dimensional in wave-dominated conditions, while their crests are two-dimensional (linear) in tidal current-dominated con-

Figure 1.4 Example of tidal inlets, including large-scale bedforms. Bedform abbreviations are: WDS: western delta shoal, EDS: eastern delta shoal, SWB: swash bar, SOB: shore oblique bar (saw-tooth bar), SCR: shoreface connected ridge. Source: Herrling and Winter (2014)

Figure 1.5 Example of megaripples, superimposed on larger sand waves in the North Sea. Source: Passchier and Kleinhans (2005)

ditions (Figure 1.5) (Passchier and Kleinhans, 2005). Rubin (2012) confirmed that megaripples with the most straight crests are found in reversing flows. Near the Ameland ebb-tidal delta, the megaripples were found to migrate in the direction of the dominant flow, both for wave-related and tidal flow, but actual migration rates are not yet known (van Rhijn, 2019). The roughness predictor of van Rijn (2007) predicts that they react instantly to the hydrodynamics, but other studies state that the adaptation time scale of megaripples is longer than typical hydrodynamic time scales, like a tidal period (Allen, 1974; Dalrymple et al., 1978). It is known that they can be washed out during storms (van Rhijn, 2019), but their dynamics and interaction with hydrodynamics on time scales of hours to years are still unclear.

Even smaller than the megaripples are small-scale ripples, which have been reported in the literature for all kinds of depths, from the continental shelf to the surf zone and intertidal areas (e.g. Li and Amos, 1999; Krämer and Winter, 2016; Meirelles et al., 2016; Wengrove et al., 2018; Wu and Parsons, 2019). However, most of these locations are either current- or wave-dominated, while on the ebb-tidal delta shoals, wave-current dominated conditions are expected. These conditions are expected to create small-scale wave-current ripples, but up to now such ripples have mostly been studied in flumes (e.g. Li and Amos, 1999; Andersen and Faraci, 2003). These ripples have heights in the order of centimeters and lengths up to 1 m, and in flumes they have been reported to become three-dimensional when the waves and currents are not parallel (Li and Amos, 1999). They are highly dynamic, and can change their appearance over a single tidal cycle (Wengrove et al., 2018). Because they are small, they react instantly to the local hydrodynamics. They will wash out when waves or currents exceed a certain threshold, but will also return quickly when the wave- or current velocity decreases again (Soulsby et al., 2012). It is not yet known how variable their dimensions are on the dynamic wave- and current dominated environment of an ebb-tidal delta, and therefore, the dynamics of the small-scale ripple-related roughness are unknown as well.

The effect of bedforms on sediment transport is different for each bedform scale. Large bedforms can affect the transport directly by migrating and therefore literally transporting sediment (e.g. Simons et al., 1965; Aberle et al., 2012; Kwohl et al., 2014). Indirectly, both larger and smaller bedforms affect the flow velocity and structure, which in turn affects the sediment transport as well. Apart from their height, which increases the friction in the flow, another parameter that affects the flow is the slip face angle (α in Figure 1.3). If this angle is larger than approximately 10° , flow separation takes place over the crest, which leads to extra vertical turbulence in the flow. This zone of turbulence is generated from velocity shear

between compressed flow with high velocity over the crest and decompressed, slower flow downstream of the bedform lee side (Best, 2005). Increasing turbulence increases the vertical mixing and sediment entrainment as well, thus increasing the amount of suspended load transport (van Rijn, 2007).

In numerical morphological models, bedforms that are smaller than the model grid size cannot be a part of the bathymetry. Therefore, bedforms with lengths smaller than approximately 50 m are parameterized by roughness predictors, which are an aggregated way of describing their effect on hydrodynamics and sediment transport. Often, this roughness is used as a calibration parameter to tune current velocities (e.g., (Elias and Hansen, 2013)). In these models, bedform roughness is usually a function of either bedform height (van Rijn, 2007) or a combination of height and wavelength (Soulsby and Whitehouse, 2005), which are in turn affected by the hydrodynamics. For example, when the small-scale ripple heights increase, their effect on the vertical flow structure is implicitly incorporated by an increase of roughness and vertical mixing. Therefore, it is not only important to know which bedforms are found on ebb-tidal deltas, their specific characteristics and interaction with hydrodynamics are very important as well for the prediction of their effect on sediment transport.

1.4 Aims and outline of this thesis

The overall aims of this thesis are (1) to determine the spatio-temporal dynamics of bedforms on ebb-tidal deltas on a variety of spatial scales, (2) compare these characteristics to existing bedform roughness predictors and (3) to determine how predicted sediment transport in mixed-energy environments is affected by predicted bedform roughness. This will improve our understanding of the sediment transport pathways on ebb-tidal deltas, and eventually, this will aid in providing a solid science base for coastal management strategies.

The three aims are divided into four research questions. The first three research questions concern the spatial distribution, temporal variability, growth and migration of bedforms on three spatial scales on ebb-tidal deltas. The bedform roughness and its implications for sediment transport will be studied for the medium and small-scale bedforms in research questions 2, 3 and 4. The research questions are:

1. What are the characteristics and dynamics of saw-tooth bars on the ebb-tidal deltas in the Wadden Sea and how do they affect sediment transport?

Saw-tooth bars are amongst the largest bedforms on ebb-tidal deltas, and they are only found on the downstream side of the delta. However, only a few years of data have been studied so far, of only a few selected deltas. Yet, they are found on many more deltas and many more time periods and their dynamics over time are still unknown. Chapter 2 presents an analysis of the characteristics and dynamics of the saw-tooth bars. They are analysed using single-beam data, which are available for all Dutch and German Wadden inlets and for many decades. Furthermore, an estimation is given of the amount of sediment that they transport through their migration.

2. What are the characteristics and dynamics of megaripples on the Ameland ebb-tidal delta, and how do these depend on time-varying hydrodynamic forcing?

On the medium spatial scale, with lengths smaller than 100 m, several bedform types are found, i.e. swash bars, dunes and megaripples. Swash bars have been studied ex-

tensively, so they are not considered here. Megaripples, on the other hand, have been given much less attention, especially in wave-current dominated areas like ebb-tidal deltas. It is unknown how dynamic they are, and on which temporal scales they react to the local hydrodynamics.

In Chapter 3 the megaripples in the Ameland Inlet are studied using multibeam data gathered between 2005 and 2017, including those of the SEAWAD field campaign. Because the time interval between consecutive measurements ranges between hours and years, the behaviour of megaripples can be studied on temporal scales of a single tidal cycle, a spring-neap tidal cycle and multiple years. Also, a first comparison is made between the measured megaripple characteristics and the roughness as predicted by van Rijn (2007).

3. What are the characteristics and dynamics of small-scale wave-current ripples on the Ameland ebb-tidal delta, and how do these depend on time-varying hydrodynamic forcing?

Small-scale ripples, with lengths and heights in the order of centimeters, are expected to occur on a large part of the delta, but since this is a mixed wave- and current-dominated environment, it is unclear what they look like and how they behave. Thus, relations between ripple characteristics and hydrodynamics need to be determined.

In Chapter 4 small-scale ripples are studied using the 3D SONAR images of the SEAWAD field campaign in 2017. Also, the measured ripple characteristics are compared to the roughness as predicted by some common roughness predictors, like van Rijn (2007) and Soulsby and Whitehouse (2005). The high temporal resolution of these images, and the inclusion of the hydrodynamic data of the same locations, makes it possible to discuss the ripples in a broader context, i.e. for all wave-current dominated areas with fine sand.

4. What are the effects of medium and small-scale bedforms on hydrodynamics and sediment transport on ebb-tidal deltas?

It is unknown how hydrodynamics, bedform-related roughness and sediment transport interact, and what level of uncertainty is imposed on the predicted sediment transport by using a roughness predictor. The parameterized bedforms in such a predictor affect the hydrodynamics, which can indirectly affect the sediment transport. Also, smaller bedforms are expected to react much faster to the hydrodynamic conditions than larger bedforms. For coastal management purposes, numerical models are often used to predict the sediment transport patterns, but it is unknown how sensitive they are to the bedform-related roughness.

This research question is answered in Chapter 5 by means of a model sensitivity analysis. The numerical model Delft3D was used to simulate hydrodynamics and sediment transport throughout the Ameland Inlet during the six weeks of the SEAWAD field campaign, using the van Rijn (2007) bedform roughness predictor. Several ripple- and megaripple-related roughness scenarios were run, also based on the answers to the second and third research question, which enabled us to study in which locations and which hydrodynamic circumstances the bedforms are of most influence on the sediment transport.

This thesis ends with Chapter 6, which gives a summary of the main conclusions of this thesis, discusses their implications and gives directions for future research.

Chapter 2 | Characteristics of saw-tooth bars on the ebb-tidal deltas of the Wadden Islands

Abstract

Saw-tooth bars are shore-oblique sand bars that are found along most ebb-tidal deltas of the Frisian Wadden Islands. Although they might significantly affect sediment transport pathways and volumes on the deltas, their general characteristics and dynamics are largely unknown. The main aim of this paper is to determine the typical saw-tooth bar heights, wave lengths, widths, orientations, migration rates and depths of occurrence. To this end, we analysed bathymetries from the Dutch and German Frisian Wadden Islands between 1970 and 2015. Bar heights range between 0.5 and 2 m, with an average of 0.74 m, and wave lengths range between 300 and 900 m, with an average of 670 m, and the bar crests have a down contour orientation of approximately 25° . The bars are between 800 and 2200 m wide. Saw-tooth bars are located at depths from 3 to 12 m, depending on the slope of the area. Migration speeds of up to 112 m/year were found, using a spatial correlation method. Bar height and migration speed are positively linearly correlated, as well as bar wave length and bar width, bar height and the orientation of the depth contours and migration speed and the orientation of the area. The derived characteristics are compared to those of other bar types to evaluate hypotheses regarding the formation mechanism of the bars, and a first estimation is given of the amount of sand that they transport by migrating.

Based on:

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2.1 Introduction

Barrier coasts make up some 10% of the world's coast line (Stutz and Pilkey, 2011). They are morphologically and ecologically highly dynamic, and understanding their dynamics is therefore vital for managing purposes (Wang et al., 2012). Typically, barrier coasts consist of a series of islands, intersected by tidal inlets, with flood-tidal deltas on their landward side and ebb-tidal deltas on their seaward side (Hayes, 1979; Hayes, 1980). Ebb-tidal deltas are relatively shallow and can protect the adjacent islands and their back-barrier basin against high storm wave energy. Their high sediment volumes, between 1 and $100 \cdot 10^6 \text{ m}^3$ (Gaudiano and Kana, 2001) can serve as a sediment source for adjacent islands and the back-barrier basin, as suggested for the Dutch Wadden Sea barrier coast by Elias et al. (2012). Nonetheless, sediment transport pathways on ebb-tidal deltas are still poorly understood (Wang et al., 2012).

The effects of sediment transport on ebb-tidal deltas can be visible by bars that migrate and attach to the coast or migrating sand waves (e.g. Bruun and Gerritsen, 1959). Large-scale bedforms can affect the amount of transported sediment in two ways: directly by migrating and therefore literally transporting sediment (e.g. Simons et al., 1965; Aberle et al., 2012), and also indirectly by altering the spatial flow structure (Kwoll et al., 2014). A typical example of an ebb-tidal delta showing different types of bedforms is given in Figure 2.1; the bedforms are described below.

On some ebb-tidal deltas, single bars are found, which are referred to as swash bars (Hayes, 1980). They are generated by waves breaking on the edges of the ebb-tidal delta, thereby reworking sand that was transported seaward through the ebb-channel (Hine, 1975). These swash bars are 0.5-2.0 m high, 50-150 m long and 50 m wide (FitzGerald et al., 2011). As visible in Figure 2.1, swash bars are oriented more or less perpendicular to the shoreline. They migrate with velocities of 100-400 m/y. The migration rate of the individual bars decreases in the onshore direction because of the increasingly larger bar height (FitzGerald, 1988). This leads to the formation of large bar complexes or shoals (FitzGerald, 1982). FitzGerald (1988) and Gaudiano and Kana (2001) found that volumes of these shoals range between $1 \cdot 10^5$ and $95 \cdot 10^6 \text{ m}^3$ for inlets in South Carolina, USA. Shoal migration velocities on the ebb-tidal deltas of the Wadden Sea range between 67 and 351 m/y (Ridderinkhof et al., 2016b). Shoals move onshore because landward flow by waves and tides is dominant over seaward flow consisting of only tides (Ridderinkhof et al., 2016a). They have a typical cycle of growth, migration and attachment to the downdrift island (FitzGerald, 1996; Israel and Dunsbergen, 1999). The periods between successive shoal attachments vary between 4 and 130 years (Ridderinkhof et al., 2016b).

In the Wadden Sea area, besides the bedforms mentioned above, saw-tooth bars (or shore-oblique sandbars or subtidal rhythmic morphology) are found on the downdrift side of the ebb-tidal delta (longshore drift is to the east). They have been given much less scientific attention and a general understanding of their characteristics is missing. We do have results from case studies that were based on an analysis of sometimes 1 to maximum 3 bathymetries of single ebb-tidal deltas, but a systematic overview based on an analysis of all available bathymetries along the Wadden Sea is missing. The first description of saw-tooth bars was given by Reineck (1963), who studied the shoreface of Norderney in 1961. He found that the downdrift side of saw-tooth bars is steeper than the upstream side and therefore named them saw-tooth bars. Yet, no other accounts of their asymmetry are known other than that of Gordeau (1999), who found that the crests and troughs are of approximately equal width.

Figure 2.1 Upper plot: Typical example of a Wadden Sea Inlet (Otzumer Balje Inlet, 2006) including the following bedforms: western/eastern ebb-tidal delta shoals (WDS/EDS), swash bars (SWB), and saw-tooth bars (STB). Red line indicates the transect location of the lower plot. Lower plot: cross-section along the transect in 2006 and 2007, showing some typical characteristics like height, wavelength and asymmetry of saw-tooth bars.

In general, saw-tooth bar heights are 0.5 – 3 m (Reineck, 1963; Gordeau, 1999; Brakenhoff et al., 2017). Wave lengths are in between 400 and 1200 m, with large differences between inlets (Reineck, 1963; Antia, 1994; Antia, 1996; Gordeau, 1999). Reineck (1963) and Gordeau (1999) showed that the cross-shore bar widths are larger than the wave lengths; widths range from 500 to 2000 m. Antia (1994) and Antia (1996) showed that saw-tooth bars originate in a water depth of 2-6 m. In contrast to the swash bars, saw-tooth bars are shore-oblique and current-transverse, but tend to become more shore-normal in the easterly (downdrift) direction (Reineck, 1963; Antia, 1994). Figures in Antia (1994) show that the saw-tooth bars originate on the part of the ebb-tidal delta that is downdrift of the dominant longshore drift direction, but it is still unknown how these bars are generated. Saw-tooth bars tend to migrate. Yet, Antia (1994) states that migration is rotational around nodal points, while

Gordeau (1999), Elias (2006) and Herrling and Winter (2014) found translational migration in the direction of the littoral sediment drift in the order of 100-200 m/y. Also, volumetric transport rates were estimated to be in the order of $10^5 \text{ m}^3/\text{y}$, assuming that the unknown constant factor is zero (Brakenhoff et al., 2017).

In short, all knowledge on saw-tooth bars is based on a few case studies, and a more systematic overview of their characteristics is missing. The main aim of this paper is to determine the characteristics of saw-tooth bars based on an analysis of bathymetries of all ebb-tidal deltas of the Wadden Sea region. In section 2.2, the available data and data analysis methods are described. Section 2.3 gives an overview of the various characteristics that were studied, as well as the relations between those characteristics. Furthermore, we will relate saw-tooth bar characteristics to the orientation and slope of the shoreface and the tidal range. In section 2.4 the characteristics as found in the present study will be compared to those found in literature, and they will be used to evaluate hypotheses regarding saw-tooth bar formation and to give an estimate on how much sand they transport. The conclusions are provided in section 2.5.

2.2 Methods

2.2.1 Study area

The study area is the Wadden Sea, which stretches from the north of the Netherlands to the west of Denmark. Because saw-tooth bars were not observed in Denmark, we focus on the area of the West and East Frisian Islands (Figure 2.2). This area can be classified as mixed energy tide-dominated to wave-dominated, depending on location. All inlets have well-developed ebb-tidal deltas (Elias et al., 2012). Mean significant wave height is 1.3 m,

Figure 2.2 The West and East Frisian Wadden Islands (border in yellow dots). Numbers denote the inlets with saw-tooth bars. 1=Texel Inlet, 2=Vlie Inlet, 3=Ameland Inlet, 4=Frisian Inlet, 5=Osterems, 6=Norderneyer Seegat, 7= Accumer Ee, 8= Otzumer Balje Inlet, 9= Harle Inlet (Source: NASA).

Table 2.1 Available bathymetries

Inlet	Years with bathymetry available
1. Texel Inlet	1926, 1948, 1971, 1981, 1986, 1991, 1994, 1997, 1999, 2001, 2006, 2009, 2012, 2015
2. Vlie Inlet	1971, 1975, 1981, 1988, 1992, 1995, 1998, 2000, 2002, 2004, 2007, 2010, 2013, 2016
3. Ameland Inlet	1975, 1989, 1993, 1996, 1999, 2002, 2005, 2006, 2007, 2008, 2010, 2011, 2012, 2014, 2016, 2017
4. Frisian Inlet	1927, 1949, 1958, 1970, 1975, 1979, 1982, 1987, 1991, 1994, 1997, 2000, 2002, 2005, 2009, 2012, 2015
5. Osterems	1985, 1986, 1989, 1990, 1995, 1996, 1998, 1999, 2000, 2001, 2002, 2003, 2004, 2005, 2006, 2007, 2008, 2009, 2010, 2011
6. Norderneyer Seegat	1986, 1987, 1989, 1990, 1992, 1993, 1995, 1996, 1998, 1999, 2001, 2002, 2004, 2005, 2006, 2007, 2008, 2009, 2010, 2011
7. Accumer Ee	1985, 1986, 1989, 1990, 1992, 1993, 1995, 1996, 1998, 1999, 2000, 2001, 2002, 2004, 2005, 2006, 2007, 2008, 2009
8. Otzumer Balje Inlet	1985, 1986, 1989, 1992, 1993, 1995, 1996, 2001, 2002, 2004, 2005, 2006, 2007, 2008, 2009, 2011, 2012
9. Harle Inlet	1992, 1993, 1995, 1996, 1997, 1998, 1999, 2000, 2001, 2002, 2003, 2004, 2005, 2006, 2007

but wave heights can become up to 6 m during storms. The tidal wave propagates from west to east, with the tidal range increasing from 1.4 m near the Texel Inlet to 3 m near the Harle Inlet (Elias et al., 2012; Balke et al., 2016).

2.2.2 Available data

Since 1990, the bathymetries of the Dutch ebb-tidal deltas have been measured every three to six years by Rijkswaterstaat (part of the Dutch Ministry of Infrastructure and the Environment), see Table 2.1. Since 2005, the Ameland Inlet and ebb-tidal delta have been measured almost every year. Before 1990, measurements occurred more sporadically. Until the 1960's, data were recorded manually and after that with a singlebeam echosounder on arrays perpendicular to the shore with 200 m spacing. They were linearly interpolated and stored on a 20×20 m grid with an estimated depth accuracy of 0.11 to 0.40 m (Perluka et al., 2006).

In Germany, bathymetric surveys were performed by various institutes between 1982 and 2012, and these were collected by the Bundesamt für Seeschifffahrt und Hydrographie (BSH). The bathymetries are stored in regular 50×50 m grids, and are composed of data that were interpolated in space and time (Valerius et al., 2013). The depths were originally measured at irregular spatial and temporal intervals, but in the data set the uncertainties are provided for all grid cells in space and time. Data were not used if the uncertainty on the depth was larger than 2 m (due to interpolation in space and time), or when they were interpolated in time. The average depth accuracy of the German survey data used here was 0.83 m.

The inlets where saw-tooth bars are found are indicated in Figure 2.2 and the years in which bathymetries are measured for those inlets are presented in Table 2.1. All x and y coordinates are given in UTM, and all depth values are given in m with respect to Mean Sea Level (MSL).

2.2.3 Data analysis

An example of a measured bathymetry (Ameland Inlet, 2005) is shown in Figure 2.3A. Although saw-tooth bars are already visible, to better see the bedforms a spatial running mean

with an area of 1000×1000 m was calculated around each grid point in each bathymetry (Figure 2.3B). Subsequently, this running mean was subtracted from the actual bathymetry (Figure 2.3A), revealing the bedforms as perturbations from the mean (Figure 2.3C). On such a perturbation map, the area where saw-tooth bars occur was indicated manually (further referred to as bar area). An example of a bar area is given in Figure 2.3C. Saw-tooth bars were classified as such if (1) they are located on the downdrift side of the ebb-tidal delta, (2) more than 1 bar is visible, (3) the bar width (i.e. shore-normal bar crest length) is larger than their wavelength, (4) their height is at least 0.5 m (larger than the measurement uncertainty) and (5) they are not oriented parallel to the downdrift shoreline (that would imply that they are swash bars or surfzone bars).

The bar width and depths of occurrence were determined manually for all available bathymetries. First, the two ends of each bar crest were selected in the perturbation plot (Figure 2.3C). After that, the distance between them was calculated and the accompanying depths in the mean bathymetry (Figure 2.3B) were determined.

Following Smith (1997), the bar height (crest to trough) was calculated by multiplying the standard deviation of the perturbations around the mean in the bar area by a constant, which depends on the shape of the bars (square, triangle or sine waves). In general, saw-tooth bars have the closest resemblance to sine waves, so the bar height was calculated by $2 \cdot \sqrt{2}\sigma$, with σ being the standard deviation within the bar area.

To verify that the bars are actually symmetric, the asymmetry (AS) was calculated. AS is given by the distance from trough to crest, divided by the distance from crest to crest. With this calculation, values are always in between 0 and 1, and the bar is fully symmetric if $AS=0.5$. If the bars are asymmetric, a value between 0.5 and 1 is to be expected, this would indicate a downdrift directed bar.

Bar wave length and crest orientation were deduced from a 2D Fast Fourier Transformation (FFT) of the bar area in the perturbation plot (Figure 2.3D). The FFT is used to calculate the bar wave length and orientation of the saw-tooth bars with respect to the north. For this, the spatial frequency k (inverse wavelength) was first calculated as $k = \sqrt{k_{x,max}^2 + k_{y,max}^2}$. In this equation, $k_{x,max}$ and $k_{y,max}$ are the x and y location of the maximum intensity in the FFT plot (Figure 3D). Wave length L was then determined as $L = 1/k$. Orientation θ is given by $\theta = \tan^{-1}(k_{y,max}/k_{x,max}) - 90$ (Figure 2.3C). However, since the orientation of the ebb-tidal deltas and coastlines of the various islands differs, these orientations cannot be compared to each other. Therefore, the orientation of the saw-tooth bar area was calculated as well. This was done by fitting a linear surface through the mean bathymetry of the bar area and subsequently determining the average bed slope and the orientation. The orientation of the area α was defined as the average angle between the depth contours of the linear surface and the North (Figure 2.3B).

The migration speeds of the saw-tooth bars were derived from spatial correlation of two perturbation maps of the bar area at subsequent moments in time. The method of Duffy and Hughes-Clarke (2005) was used. First, a search area was defined, which typically contained one saw-tooth bar. Therefore, in the present study the search area was chosen to be 600×600 m. Within the search area, a square matrix was created around each grid point, called the fit matrix ($f(x, y)$ at time T_1). The size of the fit matrix was 200×200 m. Next, a displacement matrix (r) was determined, consisting of the correlation between $f(x, y)$ and a displaced fit

Figure 2.3 Calculation of the various bar characteristics. A: the original bathymetry (Ameland Inlet, 2005), with 4 and 10-m depth contours in red. B: the mean bathymetry, with 4 and 10-m depth contours in red and the definition of the orientation of the depth contours in white. C: perturbation bathymetry, with the black angle indicating the orientation of the bars. D: FFT of the bar area. B is used for the calculation of the slope and orientation of the area α and the depth of bar occurrence. C is used for the calculation of bar height and width. D is used to calculate bar wave length and orientation θ .

matrix of a subsequent year T2 for each possible displacement within the search area ($g(x + \Delta x, y + \Delta y)$). The result is a matrix of correlation values $r_{k,l}$ on each location (x,y):

$$r_{k,l}(x, y) = \sum_{k=0}^{kmax} \sum_{l=0}^{lmax} f(x, y) g(x + \Delta x_k, y + \Delta y_l). \quad (2.1)$$

To prevent biases due to potential outliers, the correlation r was normalized by subtracting the mean from each dataset, and subsequently dividing by the standard deviation. From now on, the correlation was called R . Values in R lower than 0.3 were removed, because they indicate a low to insignificant correlation. In the resulting R matrix, the migration correlation was determined as the weighted centroid of all $R > R_{max}/\sqrt{2}$. The migration of the fit matrix was given as the distance between the centre of the fit matrix and the location of the migration correlation. The total migration of the saw-tooth bar field was determined by averaging all fit matrix migration values.

The calculation of migration speed was only possible when the two bathymetries were taken within 1 or 2 years from each other. When there is more time between two bathymetries, either the shape of the bars changes too much, or the bars migrate more than one wavelength. Therefore, the migration speed was not computed for the saw-tooth bars at the Texel, Vlie and Frisian Inlets.

2.3 Results

The relations between all studied variables, based on a linear regression, can be found in Tables 2.2 and 2.3. All variables will now be described separately.

2.3.1 Bar height and asymmetry

The height of the saw-tooth bars is between 0.5 and 2 m (Figure 2.4), although for the German inlets (Osterems until Harle Inlet) it is more commonly between 0.5 and 1 m. The largest bar height is 2.27 m, which was found for 2006 on the ebb-tidal delta of the Ameland Inlet. Both median and maximum bar heights decrease from Ameland to the east. Along each individual ebb-tidal delta, the bar heights are more or less constant in space.

On most ebb-tidal deltas, the bar heights were found to vary in time, but the time scales involved varied at and between locations. Overall, the largest time scales were found at the Dutch inlets. As an example, the bar height at the Ameland Inlet first increased from less than 0.5 m to 2.27 m in about 15 years, and then decreased again to less than 0.5 m in about

Table 2.2 Values of the linear coefficient of correlation (r) for all combinations of sandbar characteristics. Italicized values have $p < 0.01$.

	Bar height	Wave-length	Orientation	Width	Landward depth	Seaward depth	Asymmetry	Migration
Migration	0.47	0.19	-0.19	-0.03	-0.20	0.04	0.49	
Asymmetry	0.22	0.28	-0.10	0.09	0.12	-0.04		
Seaward depth	-0.17	-0.57	-0.19	-0.83	0.79			
Landward depth	-0.28	-0.47	-0.04	-0.58				
Width	0.21	0.45	0.21					
Orientation	-0.13	-0.22						
Wavelength	0.17							
Bar height								

Table 2.3 Values of the linear coefficient of correlation (r) between bar area and sandbar characteristics. Italicized values have $p < 0.01$.

	Bar height	Wave-length	Orientation	Width	Landward depth	Seaward depth	Asymmetry	Migration
Area slope	0.28	0.05	0.12	0.52	-0.44	-0.38	0	0.05
Area angle	0.27	-0.21	0.04	-0.34	0.20	0.35	0.06	0.46

10 years (Figure 2.4A). Other systems show similar behaviour. The peaks in bar height occur simultaneously at the Texel, Vlie and Frisian inlets, but no relation can be found with the peak bar height at the other inlets. For the German inlets, the absolute bar height variations in time are small, and at several moments no bars were found at all. This was the case in the Accumer Ee Inlet in 1987 and in the Harle Inlet in 1987, 1988, 2002, 2003 and 2006 (Figure 2.4C). After their disappearance, they can reappear within a few years. These observations will be further addressed in the discussion.

Bar heights are linearly correlated to the slope and orientation of the area. The correlation is negative between bar height and the slope of the area, meaning that the bar height decreases for steeper shore faces. The correlation between bar height and area orientation is positive, meaning that the bars are higher when the depth contours are deflecting more away from the north.

Figure 2.4 Absolute height of the saw-tooth bars for all inlets through time

The asymmetry of the saw-tooth bars is approximately constant in space and time. On average, the asymmetry is 0.53, which indicates that the assumption of symmetric bars was correct, and thus proving that saw-tooth bars are actually very symmetric. The standard deviation is 0.13, indicating that the spread is indeed low.

2.3.2 Bar wave length and orientation

Figure 2.5 shows the distribution of wave lengths per tidal inlet system. The values range between 313 and 909 m, with an overall average of 672 m. The smallest median wavelengths (400 m) are found for the westernmost inlet (the Texel inlet), whereas the largest median wave lengths (830 m) are found in the next (Vlie) inlet. From there on, the median wave length decreases in eastward direction to 500 m. This observation will be further addressed in section 2.4.

Figure 2.5 Occurrence of alongshore saw-tooth bar lengths. The red line in each box indicates the median, and the bottom and top edges of the box indicate the 25th and 75th percentiles, respectively. The whiskers extend to the most extreme data points not considered outliers, and the outliers (more than 1.5 times the interquartile range away from the top or bottom of the box) are plotted using the '+' symbol.

Analogous to the bar heights, the bar wave lengths also fluctuate on a multi-annual timescale. Time scales are similar to those of the bar heights, but peak heights and lengths do not occur simultaneously. Again, the timescales associated with wave lengths on the German ebb-tidal deltas are shorter, more in the order of years. On the other hand, the range in wavelengths, as given by the interquartile range, is larger for the German inlets (Figure 2.5).

Figure 2.6 presents the orientation of the bars with respect to the mean orientation of the bar area, showing that the bar orientation is similar for all inlets. On average, the angle

between the bar crests and the underlying depth contours is 66° , meaning that the bar crests have a slight down contour orientation of 25° . No changes with time were found.

No significant relation was found between the orientation of the bars and the bar height. The bar wave length, on the other hand, seems to decrease for increasing bar angles (Table 2.2). Wave length and bar orientation are not correlated to the slope and orientation of the area (Table 2.3).

2.3.3 Bar width

Figure 2.5 shows that the bar widths between 670 and 2500 m were found. The widths increase from west to east in the Dutch inlets, whereas in the German inlets widths decrease from west to east. The widths vary on a time scale similar to the lengths and heights, but peaks do not coincide. They are negatively correlated to the bed slope and orientation (Table 2.3), meaning that the bar width is smaller when the shoreface is steeper or more oblique. On top of that, bars with larger wave lengths also have larger widths (Table 2.2). The ratio width/wave length is in between 1.5 and 3.5 for all inlets, with the higher values for the Texel, Frisian and Osterems Inlet, and the lower values for the other inlets.

2.3.4 Depths of occurrence

Depths of the landward ends of the bars range between 3 and 8 m while depths of the most seaward end of the bars are in between 5 and 12 m (Figure 2.8). The largest interannual

Figure 2.6 Occurrence of the angle between the bar crests and the bar area. Symbols are explained in Figure 2.5.

variability in nearshore and seaward depths is found in the German inlets, which is visible in the amount of scatter in Figure 2.8. For example, the seaward depths of the Osterems ranges between 9.5 and 12 m, while those of the Ameland Inlet range between 7.5 and 9 m.

Bars that start deeper also end deeper and vice versa. Seaward and landward depth are positively correlated (Table 2.2). Besides the relation between the landward and seaward depth, these depths and the slope of the shoreface are also negatively correlated (Table 2.3). When the shoreface is steeper, the most landward end of the bars is found on shallower depths. The same holds for the seaward depth. The depths of occurrence also have a relation with the bar wave length; bars that occur on larger depths are longer. This is true for both landward and seaward depth (Table 2.2). Naturally, bars that are wider also extend to deeper areas. Yet, it was also found that wider bars also tend to start at larger depths.

2.3.5 Migration speed

The migration speed of the saw-tooth bars for all inlets varies between 2 and 112 m/y (Figure 2.9). On average, the bars migrate with 30 m/y, but it is visible that the migration speed highly depends on the location of the inlet along the Wadden Sea coast. At the Norderneyer Seegat and the Accumer Ee Inlet, migration speeds are very low, while at the same time the bars move quite fast at the Ameland and Harle Inlet. Apart from their relation with their position along the Wadden coast, bars with larger heights tend to migrate faster than lower

Figure 2.7 Bar crest length for all inlets. Symbols are explained in Figure 2.5.

bars (Table 2.2). Both characteristics fluctuate on similar multi-annual timescales. On top of that, migration is positively correlated to the orientation of the shoreface (Table 2.3).

2.4 Discussion

In this study, the characteristics of saw-tooth bars were analysed and combined into a complete and systematic overview. The found characteristics were within the range as found in previous studies, but the present study is the first in which spatial trends, temporal behaviour and relations between the bar characteristics and their environment could be determined. Yet, it is still unclear how their characteristics compare to other bar types found on continental shelves, by which mechanisms they are generated and how they affect sediment transport. These three questions will be discussed below.

2.4.1 Saw-tooth bar properties compared to other bar types

Antia (1994) classified saw-tooth bars as surf zone transverse bars, but the findings of this chapter indicate that they are not. Like saw-tooth bars, transverse bars are oriented perpendicular or oblique to the shore and are 0.3 – 2 m high (e.g. Ribas et al., 2015). However, they are typically found in wave-dominated surf zones (e.g. Wright and Short, 1984). Yet, the longest type of transverse bars, the Transverse Bar and Rip (TBR), has a wavelength of 75–750 m and a width in the order of 100 m. Other transverse bars (large-scale finger bars, LFB)

Figure 2.8 Relation between nearshore and offshore depth.

were found in the low-energy microtidal environment of Florida (USA) by Gelfenbaum and Brooks (2003). The main similarities between LFB and saw-tooth bars are their height, and the fact that both types of bars reach out of the surf zone. However, LFB have wavelengths of approximately 100 m and widths of up to 4 km. Saw-tooth bars, on the other hand, always have wavelengths of at least 300 m and widths in the order of 800-2200 m. While the maximum depth at which saw-tooth bars are found is 12 m, for LFB this is approximately 8 m. This, in combination with the high-energy macrotidal environment of the Wadden Sea, suggests that the processes acting on saw-tooth bars are probably different from those acting on LFB.

Another bar type with similarities to saw-tooth bars is the shoreface connected sand ridge (SCR) (Swift et al., 1978; de Swart and Calvete, 2003). They also occur on depths of 5-20 m. SCR are several meters high, and much larger than saw-tooth bars: widths are in the order of 10 km and wave lengths in the order of 5 km (Swift et al., 1978; de Swart and Calvete, 2003). Migration on the other hand is much smaller: only several meters per year, and only by storm-induced currents (de Swart and Calvete, 2003). Another difference is their orientation: while saw-tooth bars have a slight down-contour orientation (the angle between the bar crests and the depth contours is smaller than 90°), SCR are up-contour oriented (the angle between the bar crests and the depth contours is larger than 90°) (Swift et al., 1978).

Lastly, saw-tooth bars have a resemblance to sand waves (Németh et al., 2002; Besio et al., 2006; Blondeaux and Vittori, 2011). Saw-tooth bars and sand waves have similar wave length

Figure 2.9 Migration speed of saw-tooth bars. Symbols are explained in Figure 2.5.

scales, but sand waves occur on larger depths (~ 30 m) and their heights can be several meters (Németh et al., 2002). Because of the small currents at these large water depths, migration speeds of sand waves are in the order of only a few meters per year (Németh et al., 2002). In tidal channels, sand wave migration rates are 20–80 m/y, which is equal to the migration rates found in the present study (Buijsman and Ridderinkhof, 2008a). Time scales of formation and growth are more than 20 years for sand waves (Knaapen et al., 2006), while the present study found that saw-tooth bars can form 2–5 years. Sand waves are oriented more or less perpendicular to the tidal current, with a maximum counter clockwise deviation of 10° (Hulscher, 1996). Hulscher (1996) further found that sand wave height increases in offshore direction, so that would explain why saw-tooth bars are lower. Sand waves are created by an instability of a sandy bed interacting with the tidal current (Besio et al., 2006; Blondeaux and Vittori, 2011). Therefore, saw-tooth bars might be shallow water sand waves occurring on ebb-tidal delta lobes.

2.4.2 Hypotheses regarding saw-tooth bar formation

The results of this study are in agreement with Gordeau (1999) and Herrling and Winter (2014), in finding a translational migration of the bars in the direction of littoral sediment drift. However, it was also confirmed that the bars on several ebb-tidal deltas of the East Frisian islands show little to no migration, as found by Antia (1994). Antia (1994) suggested that the troughs of the saw-tooth bars were formed by rip currents resulting from edge waves. This hypothesis is an example of the so-called forcing template theory, which assumes a fixed flow pattern that would result in a fixed bedform pattern. Since then, the self-organisation theory was further developed, which has provided a better explanation for the generation and occurrence of nearshore bedforms (Coco and Murray, 2007). The main reason for this is the fact that the forcing template theory does not account for the feedback and interactions between hydrodynamics and morphodynamics. On top of that, oblique wave incidence angles (which are found on most ebb-tidal deltas) cause edge waves to be progressive, which in turn causes a non-stationary flow pattern that moves downdrift much faster than the bars (Ribas et al., 2015). Lastly, flow velocities of edge waves are not large enough to significantly change the morphology (Bryan and Bowen, 1998). In the light of these findings, it is more likely edge waves or rip currents – if present – are an effect of the bar morphology, rather than its cause. The formation of the bars is hypothesized to be a result of an instability mechanism caused by the interaction between the sandy bed and the hydrodynamics.

The main reason for this hypothesis is the cyclic behaviour of the bar height and their observed migration. The linear relations between bar height and migration speed, and between bar height and the orientation of the larger area including the shoal suggest that the cyclic behaviour of the shoals plays an important role in the development of saw-tooth bars. Since there are clear stages of saw-tooth bar growth and disappearance, the development of saw-tooth bars is hypothesised to be similar to that of sand waves: an instability mechanism related to tidal and wave-driven currents.

There are three types of hydrodynamics that can interact with the bed and thereby generate bedforms: tides, wind- and wave-driven currents. Wind-driven currents are known to create SCR. However, as described above SCR have much larger dimensions than saw-tooth bars and are orientated differently with respect to the depth contours. Also, the peaks in saw-tooth bar height do not occur simultaneously at the different deltas, implying that they are not related to storminess or yearly-averaged wave height, since these characteristics should

be more or less similar throughout the study area. Therefore, it is unlikely that saw-tooth bars are formed by wind-driven currents.

The decrease in median wavelength and bar height in the easterly direction could indicate a relation with the tide, since the tidal amplitude increases over the same distance. The smallest wavelengths occur under the largest tidal currents, which is also found for sand waves (van Santen et al., 2011). However, the increase of the tidal amplitude does not necessarily cause an increase in tidal current velocity, since the tide is a standing wave. On top of that, the tide cannot explain the cyclicity in the bar heights.

Since wave directions are relatively constant in the study area, the effect of wave-related currents is mainly dependent on their direction with respect to the bars. Because of the cyclic behaviour of the ebb-tidal deltas, the orientation of the depth contours changes through time. A positive linear relation was found between bar height and the orientation of the bar area. A larger orientation angle of the depth contours with respect to the north indicates that the shoal is protruding further into the sea. At this part in the cycle, the shoal creates a shadow zone in its downdrift part, where wave incidence angles are large and the waves propagate more or less parallel to the depth contours. Here, saw-tooth bars might develop. When the shoal has just attached to the coast, the depth contours and the direction of wave propagation are no longer aligned, causing the saw-tooth bars to disappear.

Also, the larger ebb-tidal deltas have the highest saw-tooth bars. Larger deltas have a larger shadow zone, so the bars have more space to develop. Out of the shadow zone, along the island coast, the bars gradually disappear. This is probably caused by the increasing angle between the direction of wave propagation and the depth contours at these locations. These observations related to the waves suggest that a high-angle wave instability might generate saw-tooth bars (Ashton and Murray, 2006). Although it is also possible that morphologic parameters of the shoreface are important as well (e.g. composition of the subsurface, types of sediment available, or slope of the shoreface) this could also explain why no saw-tooth bars are found on the N-W oriented Wadden Islands of Germany and Denmark. No relation could be found between the annual downdrift wave-driven sediment transport (as described by Ridderinkhof et al. (2016b)) and the presence and dimensions of the saw-tooth bars. Further research should be done on the interaction of the bars with the hydrodynamics, and their effect on sediment transport. Since observations are limited, this should be done by a modelling study. Such studies could reveal the underlying morphodynamic instability mechanism that causes the formation of these saw-tooth bars.

2.4.3 The effect of saw-tooth bars on sediment transport

By migrating, the saw-tooth bars transport sediment. When the height H and migration velocity U of the saw-tooth bars are known, the volumetric sediment transport per unit width per unit time (q_b in m^3/y) can be calculated using $q_b = 0.5 \times UH(1 - \phi) + C_1$ (Simons et al., 1965). In this equation, ϕ denotes the porosity of the sediment, which is generally taken to be 0.4. C_1 is a constant, representing the part of the bed load that is not transported by bedforms. For now, C_1 is assumed to be 0, in order to only calculate the bedform-related bed load transport. When q_b is multiplied with the bedform width (in m), the volumetric sediment transport in m^3/y is obtained.

The present study found average bar heights, widths and migration speeds of 0.74 m, 1442 m, and 30 m/y, respectively, resulting in an average volumetric sediment transport of 9.603

m^3/y . Using maximum values of $U = 112 \text{ m/y}$, $H = 2 \text{ m}$ and width = 2200 m leads to a maximum volumetric transport of $147.840 \text{ m}^3/\text{y}$.

2.5 Conclusions

Saw-tooth bars were studied using bathymetric maps of the ebb-tidal deltas of the Frisian Wadden Islands. Saw-tooth bars are down-current oriented sandbars at the downdrift part of ebb-tidal deltas. Contrary to what their name suggests, they are not highly asymmetric. The heights of saw-tooth bars are in between 0.5 and 2 m, and are varying through time with a multi-annual variability. Their wavelengths range between 300 and 900 m, with an average of 670 m, and crest lengths between 800 and 2200 m are found on the ebb-tidal deltas of the west- and east Frisian Wadden Islands. They occur between 3 and 12 m depth. Migration speeds vary between 2 and 112 m/y . Bar height and migration speed, wave length and width, and seaward and landward depth are positively correlated. Bar height and width both depend on the slope and orientation of the shoreface, and migration speeds only depend on the orientation of the shoreface. Not all inlets in the Wadden Sea contain saw-tooth bars (i.e. they are not found at the north Frisian and Danish Wadden Islands), and not all saw-tooth bars migrate. Based on the present results, it is hypothesised that saw-tooth bars are sand waves, generated by an instability mechanism forced by tidal and wave-driven currents. On average, migrating sawtooth bars transport almost 10.000 m^3 of sand per year.

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Chapter 3 | Characteristics of megaripples at the Ameland tidal inlet

Abstract

Megaripples are generally defined as bedforms with lengths between 0.6 and 20 m, and heights between 0.06 and 2 m. It is not yet known how they behave in areas with both waves and tidal currents, so their dimensions were studied in the wave-current dominated Ameland inlet. This was done using multibeam data from (1) one tidal cycle (2) a spring-neap tidal cycle and (3) yearly data. In the case of the Ameland inlet, the megaripples were between 2 and 28 m long and between 0.05 and 1 m high. Their heights and lengths do not change significantly over a tidal cycle or a spring-neap tidal cycle. Intriguingly, over longer periods of time, the variability in wave lengths is much larger (standard deviations of approximately 4 m), while the megaripple heights are more constant (standard deviation of approximately 0.1 m), indicating that the steepness varies in space and time. The megaripple dimensions were compared to the local hydrodynamics. Typical migration speeds range between 0 and 2 m/h, depending on the local current velocity. Megaripple shape and migration directions reverse some time after the tidal flow reversal and a net migration was observed in the direction of the tidal flow with up to 4 m in the period between maximum flood and maximum ebb. Therefore, measurements of megaripple asymmetry are only indicative of the transport direction at the time of the measurements, and not of the net migration direction. Both the megaripple migration and the steepness of their slipface slope are related to the current velocity. The steepness of the slip face is linearly related to the megaripple height, and the megaripple height and asymmetry are linearly related to the difference between ebb- and flood flow velocities. Furthermore, megaripple dimensions were compared to existing predictors. In Delft3D the van Rijn (2007) predictor is used, in which megaripple characteristics depend on flow velocity. Therefore, on a tidal time scale the predicted megaripple characteristics change significantly, while observations only show minor variations. Because megaripple-related roughness was shown in earlier studies to significantly affect the hydrodynamics and sediment transport, further research is needed to quantify the relation between bedform characteristics and external forcing in wave-current environment like a tidal inlet.

3.1 Introduction

Bedforms in coastal zones occur on a variety of spatial scales, from bars with a length in the order of kilometers to centimeter-scale ripples. Bedforms of different scales are often superimposed (e.g. Dalrymple et al., 1978; Bartholdy et al., 2005; Ernsten et al., 2005; Bellec et al., 2019; Jones and Traykovski, 2019). Megaripples are in the intermediate part of the scale, having lengths (λ) between 0.6 and 20 m and heights (η) between 0.06 and 2 m (Idier et al., 2004). In the literature, megaripples have also been called dunes (e.g. van den Berg et al., 1995; Hoekstra et al., 2004; Bartholdy et al., 2005). They occur in various environments, such as intertidal shoals (Hoekstra et al., 2004), current-dominated tidal channels (Elias, 2018), and the continental shelf (Passchier and Kleinans, 2005; Aldridge et al., 2015). Results of a modelling study show that the presence of megaripples has relatively large effects on the hydrodynamics and sediment transport (Chapter 5). Thus, it is important to understand their characteristics, as well as their relation with local sediment properties and flow and wave conditions. This will improve our understanding on sediment transport pathways.

In general, it is difficult to relate the megaripple height and length to the hydrodynamics, because the equilibrium/adaptation time scale of these bedforms is in the order of days, which is longer than typical hydrodynamic time scales, like a tidal period (Allen, 1974; Dalrymple et al., 1978). However, some studies suggest that the dimensions might be related to the water depth. For example, in the surf zone megaripples are 1.2-2.5 m in length and 0.02-0.1 m in height (Clarke and Werner, 2004; Larsen et al., 2015), while in a 6 m deep tidal inlet lengths of up to 25 m and heights of up to 0.7 m are found (Dalrymple et al., 1978). Based on similar studies in flumes, rivers and tidal channels (van Rijn, 1984), van Rijn (2007) created a predictor for the megaripple height that depends on water depth and the mobility number, which is a function of the grain size and the combined wave- and tidal current related velocity. The van Rijn (2007) predictor results in megaripple heights that are always in between 0.04 m and 0.01 times the water depth, with a maximum of 0.4 m, hence dismissing the fact that megaripples can in fact become higher than 0.4 m (Idier et al., 2004). This raises questions about the conditions under which the van Rijn (2007) predictor can be applied and whether it should be made more generally applicable.

In a tidal inlet channel, Bartholdy et al. (2005) found that megaripples did not scale with water depth, but that megaripple dimensions instead depended on the grain size only, or on grain size and current velocity combined. Using the relations observed by Bartholdy et al. (2005), Aldridge et al. (2015) predicted that megaripples should be present on all depths (up to 50 m) of the southern North Sea between the Netherlands and the United Kingdom, because tidal current velocities in this entire area are sufficiently strong.

While it is thus unclear how megaripple size depends on environmental factors, there is clear consensus that megaripple height (η) and wave length (λ) are positively correlated (e.g. Dalrymple et al., 1978; Flemming, 1988; Lefebvre et al., 2016). Dalrymple et al. (1978) found this relation for both two-dimensional and three-dimensional megaripples in the Bay of Fundy (Canada). Flemming (1988) used data from many environments including flumes, and showed that $\eta = 0.0677\lambda^{0.8098}$ ($r=0.98$). The data set did not only include megaripples (called dunes in that study), but also small-scale ripples and sand waves (or large dunes), which all abide to this parametrization. This implies that both ripples and megaripples are part of the same morphologic family. It was also found that the maximum bedform height is related to its wave length in a similar way. Megaripples rarely reach this maximum height,

which Flemming (1988) hypothesized to be caused by a lack of sediment availability. Despite this lack of sediment, the wave length is always reached (Flemming, 1988).

Several clear relations have been found between currents and megaripple creation and migration. Clarke and Werner (2004) and Larsen et al. (2015) found a hysteresis effect in the presence of surf zone megaripples, i.e. megaripples become present several hours after the onset of the right hydrodynamic conditions (i.e. high enough current velocity). In deeper water (5–10 m) near an ebb-tidal delta, hence subjected to both waves and tidal currents, megaripples were observed to wash out during storms, and to reappear within weeks (van Rhijn, 2019). In relatively shallow wave-dominated areas, megaripples migrated in the direction of wind wave propagation, but in channels and other current-dominated areas they propagated in the direction of the current (van Rhijn, 2019).

By migrating, megaripples transport sediment. This transport occurs in the direction of the steep lee side, which is visible in the distribution of grain size fractions across each megaripple, with a mixture of grain sizes on the lee side (Koop et al., 2019). Furthermore, the presence of megaripples affects the roughness experienced by the flow, which indirectly affects the sediment transport. According to van Rijn (1989), hydraulic roughness depends on the bedform height, length and the slopes of the lee side of the bedform, while Lefebvre et al. (2016) states that the steepness of the lee side slope is most important in determining roughness. Both van den Berg et al. (1995) and van Rijn (2007) state that the megaripple-related roughness is in the order of half of their height. In a tidal estuary with water depth of 6 m, van den Berg et al. (1995) found megaripple-related hydraulic roughness changed significantly over a tidal cycle. The megaripple asymmetry changed with the tidal flow reversal, after a time lag of a few hours. Heights and lengths remained more or less constant throughout a tidal cycle, but the steepness of their slipface was highest during maximum flow velocity, and therefore roughness was highest at that time as well (van den Berg et al., 1995). Ernsten et al. (2006) and Jones and Traykovski (2018) also found megaripples that reversed with the tide.

Previous research on megaripples has mainly focused on tide-dominated or wave-dominated environments, while megaripples are also present in combined wave-current dominated systems, such as tidal inlet systems with their deep channels and shallower shoals. In this chapter, we focus on megaripple dynamics in space and time at the Ameland Inlet. The inlet and its accompanying ebb-tidal delta are located in the north of the Netherlands, between the Wadden Islands of Terschelling and Ameland (Figure 3.1). Since 2005, parts of the bathymetry of this inlet system have been measured regularly using Multibeam echosounding (Mbes). Between 2005 and 2016, the measurement purpose was to monitor coastal protection measures that were taken on the west and northwest part of Ameland and therefore, only parts of the main channel in the inlet throat were surveyed. In 2017, a large field campaign was conducted in the inlet, during which several parts across the channel were measured repeatedly over the course of a week (30 August until 07 September) (van Prooijen et al., 2020).

Studies on subsets of these data (Chapter 5, Elias (2018)) have focussed on megaripple height and length, but did not examine megaripple dynamics from short (tidal period) to long (decades) time scales, and how these dynamics depend on hydrodynamical factors such as water depth and flow velocity. As such, it is also not yet known how existing megaripple predictors are valid in wave-current dominated tidal inlet systems and ebb-tidal deltas. Therefore, the aims of this study are (1) to determine the spatio-temporal dynamics of megarip-

ples on the Ameland ebb-tidal delta, (2) to study how megaripple dimensions are related to hydrodynamic forcing and geometry (depth, bed slope), (3) to study how well megaripple dimensions are predicted by existing ripple predictors; 4) to study how megaripples directly affect sediment transport by migrating.

We use all available multibeam data of the Ameland Inlet, encompassing surveys over one tidal period, a spring-neap tidal cycle and over multiple years. The surveys cover water depths between 3 and 25 m. We will compare the quantified megaripple characteristics to the hydrodynamics simulated with the numerical model Delft3D. Furthermore, we will compare measured and predicted megaripple heights and provide hypotheses on the effect of megaripples on sediment transport.

3.2 Methods

3.2.1 Study site and field data sets

The Ameland Inlet is located in the north of the Netherlands. Currently, two channels are visible in this inlet, with one being dominant (Figure 3.1). The delta itself consists of an eastern and a western lobe, with minimum water depths of approximately 1 m relative to mean sea level. The dominant wind-generated waves approach from the northwest, having an average significant height H_s of 1.3 m, and an average spectral period T_{m01} of 5 s (Roskam, 1988). Maximum significant storm wave heights can exceed 6 m, with storm surges of over 2 m. The tidal wave propagates along the coast from west to east, with a tidal range of ≈ 2 m near the Ameland Inlet. Maximum shore-parallel tidal current velocities are 0.5 to 1.0 m/s (Elias et al., 2012), and the tidal prism is 430 Mm³ (Sha, 1989).

In the Texel Inlet, which is located approximately 70 km to the southwest of the Ameland Inlet, sand waves were found with wave lengths of 125–250 m, and heights of 2–3 m (Buijsman and Ridderinkhof, 2008b). It was expected that similar bedforms would also occur in the Ameland Inlet, since similar hydrodynamic conditions occur. To check for the presence of sand waves, a loess filter was applied to the bathymetric data that removed length scales below 20 m and above 150 m. The sand waves in the Ameland Inlet were approximately 100 m long and 1.4 m high and occurred only in the deepest part of the channel. Yet, this part was not measured that often, so data on dunes were too scarce for any further analysis. Therefore, the focus of this study was on the megaripples only.

Figure 3.1 gives an overview of the locations of the bathymetric measurements that are studied in this chapter. The bathymetric data can be divided into two data sets, which were both measured using Multibeam echosounding. First, the bathymetry of the channel adjacent to the island of Ameland was measured almost every year between 2005 and 2016 by Rijkswaterstaat (pink in Figure 3.1), in order to monitor coastal protection measures. These data were all gridded with a 1 m resolution. Second, in 2017, a few areas across the main ebb channel were measured more intensively over the course of one week (green in Figure 3.1). These data were all gridded with a 0.5 m resolution and the measured areas were all 300 m wide, but the entire width was not always measured. Minimum widths were 200 m. Three parts (A, B and C) were measured several times throughout this week, during which the tide changed from neap to spring (Figure 3.3). Area D was measured 9 times on a single day between maximum flood and maximum ebb velocity (Figure 3.3). The exact dates of all measurements are given in Table 3.1. Some parts of the delta and the inlet were also

measured with a multibeam echosounder before and after the 2017 campaign, which showed that on several locations megaripples were present, with heights of 0.1–0.2 m and lengths of 10–20 m (see Chapter 4). However, these data could not give additional insight into either the interaction between megaripples and hydrodynamics or into the long-term behaviour of megaripples, so this data set was not considered in this chapter.

In the same 2017 measurement campaign, four frames were placed on the ebb-tidal delta (Figure 3.2) from 29 August until 10 October 2017; their locations were chosen to represent the wide range of hydrodynamic conditions on the delta. Frames 1 and 5 were positioned at the delta lobe, at depths of 8.6 and 8.9 m, respectively, on sediment with $D_{50} = 225$ (F1) and $186 \mu\text{m}$ (F5). Frame 3 was placed in the channel, at 15.1 m depth and on sediment with $D_{50} = 216 \mu\text{m}$. Frame 4 was placed at the ebb-delta shoal, where the average water depth was 6.5 m and the D_{50} was $186 \mu\text{m}$. Over the entire delta the mean grain size was $211 \mu\text{m}$, with a standard deviation of $30 \mu\text{m}$ (Pearson et al., 2019). The frames measured water levels, wave heights, periods and directions, current speeds and directions, and bed-forms, using pressure sensors, Acoustic Doppler Velocity meters (ADV) and Sonars, respectively (van Prooijen et al., 2020). More details on the small-scale ripple measurements are given in Section 4.2.3. Discharges through the tidal inlet were estimated from ship-mounted Acoustic Doppler Current Profiler (ADCP) flow measurements, collecting data along inlet cross-sections during 6 tidal cycles. Further information on the field campaign, including the data collection, can be found in van Prooijen et al. (2020).

Figure 3.1 Bathymetry of 2017 with the surveyed areas of data sets 1 and 2. Magenta areas: soundings of 2005-2016 (data set 1), green rectangles A, B, C, D: soundings of 30 August 2017 - 7 September 2017 (data set 2).

3.2.2 Model Setup and Boundary Conditions

Regarding hydrodynamics, the depth-averaged current velocity was modelled for the period in which Mbes data set 2 was acquired, using a hydrodynamically validated version of the Delft3D Coastal Genesis II Terschelling-Ameland inlet (CG2TA) model, with Delft3D-FLOW version 3.59.01.48550, FLOW2D3D version 6.02.13.7545 and SWAN version 41.10. The model setup, forcing, calibration and validation are described in (Nederhoff et al., 2019a;

Table 3.1 Dates (dd-mm-yyyy) at which the measurements were done per data set

Data set 1	Data set 2
18-07-2005	30-08-2017 (ABC)
18-09-2006	31-08-2017 (ABC)
24-05-2007	02-09-2017 (AB)
25-05-2009	03-09-2017 (AC)
06-07-2010	04-09-2017 (ABC)
18-04-2011	06-09-2017 (ABC)
29-11-2011	07-09-2017 (D)
17-04-2012	(9 times)
18-11-2012	
12-02-2013	
02-05-2013	
03-04-2014	
11-02-2015	
22-05-2015	
20-08-2015	
19-01-2016	
03-03-2016	
04-07-2016	
06-12-2016	

Nederhoff et al., 2019b) and will be repeated briefly below. The model grid extended from the island of Vlieland to the island of Schiermonnikoog, thus encompassing the Vlie, Ameland and Frisian Inlets. The seaward boundary was located approximately 30 km offshore, while in the basin, the tidal divides south of Vlieland and Schiermonnikoog were taken as boundaries. The grid was obtained from Nederhoff et al. (2019a), with typical grid cell sizes in the region of interest of 50–150 m (Figure 3.2), assuring that wave processes were represented in enough detail and also that main bathymetric features such as shoals and channels were present. At the open boundaries water levels were prescribed, which were derived from a large-scale North Sea model (nesting) that was forced by tides, air pressure and wind. Furthermore, at the open North Sea boundaries time series of significant wave height were prescribed, based on the wave stations given in Figure 3.2. In the back barrier basin, a zero-gradient Neumann boundary was applied. The model bathymetry was based on the so-called ‘Vaklodingen’, the bathymetry that is measured and processed by Rijkswaterstaat (part of the Dutch Ministry of Infrastructure and Water Management) once every 1–3 years using a singlebeam echosounder (van Prooijen et al., 2020). The bathymetry of the Ameland inlet was measured in 2017, and it was combined with data from earlier years (2017–2008) to construct the bathymetry for the entire model grid. At each grid cell, the most recent bathymetric information was used.

The model ran in depth-averaged (2DH) mode and was forced by water levels, waves, wind and atmospheric pressure. The water levels that were forced on the model boundaries were derived from the Dutch Continental Shelf Model-Zuidelijke Noordzee (DCSMv4ZUNOV6) model, which encompasses the entire southern part of the North Sea and is driven by tides, winds and air pressure (Zijl et al., 2013). The wave characteristics at the boundaries were derived from wave spectra that were measured at two stations near the western and eastern boundaries (Figure 3.2), which were interpolated onto the northern boundary. The waves were then propagated through the model domain by the SWAN wave module. SWAN and FLOW communicated every 30 min (two-way coupling).

Figure 3.2 Left: extent of the model grids, including grid cell sizes (given by the square root of the grid cell area A) of the FLOW model. Right: zoom of the model bathymetry at the Ameland Inlet and locations of the measurement frames (F1–F5).

The model was calibrated on water levels measured throughout 2017 offshore of Terschelling and Ameland and with current speeds, discharges and wave characteristics as measured during the measurement campaign of August–October 2017 (calibration stations are given in Figure 3.2). The calibrated model was validated hydrodynamically using field data of 2008, 2011, November 2017 and January and March 2018, running the model without morphological updates. Measured depth-averaged current speeds derived from the ADCP data of frames 1, 4 and 5 were reproduced with a root-mean-square error (rmse) of 0.10–0.15 m/s (Table 3.2). These values are in the same order of magnitude as those of validated models used in previous modelling studies (Nederhoff et al., 2019a; Zijl et al., 2013). The differences between model predictions and measurements did not depend on the wave- or current conditions, with no variation in rmse between storm and calm conditions. The biases were low (mostly in the order of 0.05 m/s) because the differences in positive and negative values were approximately equal. With 0.34 m/s the rmse values were highest at frame 3, while the measured current velocities at this frame did not exceed 0.5 m/s. At the other frames, rmse values were lower, but considering that current velocities were up to 1 m/s (Figure 3.3), they were still rather high. This is due to slight errors in the predictions of the current direction. The timing of the peak ebb and flood velocities was equal in the model predictions and the measurements.

Table 3.2 Root-mean-square error [m/s] and bias [m/s] between measured and modelled current velocities (x- (east-) and y- (north-) component and magnitude) at the frame locations given in Figure 3.2.

Frame	Bias u_{cx}	Bias u_{cy}	Bias u_c	Rmse u_{cx}	Rmse u_{cy}	Rmse u_c
F1	0.05	−0.03	−0.03	0.14	0.11	0.12
F3	0.00	0.03	0.25	0.18	0.37	0.34
F4	0.08	0.02	0.01	0.14	0.11	0.13
F5	−0.01	0.04	0.06	0.13	0.10	0.14

Figure 3.3 Modelled north-south (blue) and east-west (red) current velocity in the center of area D, positive = towards the north (blue) or east (red), indicating the ebb direction. Multibeam sounding periods are given in grey. Top: dd/mm in 2017, soundings of areas A, B and C, bottom: hh:mm on September 7, 2017, soundings area D.

3.2.3 Analysis

Visually, a superposition of large and smaller bedforms can be distinguished in some of the images of data set 2 (Figure 3.4). The large bedforms, with length scales approximately 100 m are sand waves, and they will be further discussed in the Discussion section. The smaller bedforms are megaripples with length scales of approximately 10 m. Areas A-D were oriented across the channel and had an along-channel width of 200-300 m, they can thus be seen as cross-channel profiles. The yearly maps of data set 1 were divided into 15 smaller areas with along-channel widths similar to areas A-D (between 296 and 726 m) to ensure that the characteristics and orientation of the megaripples were constant in each area.

A two-dimensional loess filter was applied to the data (the smaller areas of data set 1 and areas A-D of data set 2) to remove wave lengths above 30 m. This filter scale was chosen based on visual inspection of the loess-filtered megaripple maps. Visual inspection further showed that megaripples had a clear 2D character and therefore transects were drawn perpendicular to the megaripple crests for each megaripple map (which is also perpendicular to the cross-channel profile, see Figure 3.5). The distance between the transects was approximately 5 m. The Mbes data grids were interpolated onto these transects, resulting in a step size along each transect of 0.7-0.75 m. To ensure that each transect was perpendicular to the megaripple crests, the transects were rotated with respect to the previous transect with $\pm 5^\circ$ with a step

Figure 3.4 Left: bathymetry of area D, middle: length scales smaller than 20 m, right: length scales between 20 and 150 m.

size of 1° . The orientation at which the wave length was smallest, was taken as the profile orientation. The wave length of the megaripples along each transect was determined with a wavelet analysis using the method of Grinsted et al. (2004), which was based on Torrence and Compo (1998). The transect-specific wave length was taken to be the wave length with the maximum power in the wavelet spectrum. The megaripple height along each transect was quantified as $2\sqrt{2}\sigma$ (Krämer and Winter, 2016), with σ being the standard deviation of the loess-filtered bed elevation z along the transect. Megaripple asymmetry was calculated as:

$$As = \frac{\langle \mathcal{H}(z)^3 \rangle}{\langle z^2 \rangle^{3/2}}, \quad (3.1)$$

where $\langle \rangle$ is the along-transect-averaging operator and $\mathcal{H}$ is the Hilbert transform. The value of As theoretically ranges between -2 and 2 (for saw-tooth waves with a perfectly vertical slip face), but for natural conditions a range of -1 and 1 is found (Ruessink et al., 2012), with $As = 0$ representing a perfect sine wave. The sign of As indicates the direction in which the slip face is pointing. In the present study, the transects were consistently chosen to run in the direction of the flood current, so positive asymmetry values give an indication of migration in the flood direction (towards the east or south in the studied area).

The slip face angles along each transect were calculated using the method of Koop et al. (2019) (Figure 3.5). First, the transect was smoothed with a moving average of 2 datapoints to remove excess slope values, and then the slope was calculated of each point along the transect (similar to taking the first order derivative of the bed level elevation). The transects were chosen in such a way that positive slopes are in the ebb direction and negative slopes are in the flood direction. Thus, when the average of the negative values were largest, the steep lee side was pointing in the flood direction (towards the east or south), and when the average of the positive values were larger, the slip face was pointing in the ebb direction (towards the north or west). A peak-finding algorithm located the actual slip faces, i.e. the point with the maximum slope at each individual bedform. The average of these peak slopes per transect was taken as the slip face slope (i.e. the average maximum leeside angle).

Even though the slip face angle and As both describe the shape of the bedform, they are complementary instead of redundant. The absolute value of the slip face angle is indicative of the megaripple-related roughness (Lefebvre and Winter, 2016), and its sign shows if it points in the ebb or flood direction. As supports this value, because it can show the possibly more gradual change in shape. Only one side of the bedform is steepest, so the sign of the slipface slope will change in an instant, but the As value might show more correlation with the local hydrodynamics.

Figure 3.5 Left: example of the transects drawn on the megaripple areas, in this case on area D. Five transects are shown, to give an indication of the slightly different orientation of consecutive transects. The bold line is the transect of which the cross-section is given on the right. Right: example of a bed elevation (z) profile along a transect through the megaripples in area D. The crosses indicate the slip face of each single megaripple. In this profile, $\eta = 0.65$ m, $\lambda = 9.8$ m, $As = 0.37$ and slipface angle = -14° .

Megaripple migration velocity and direction were derived from a two-dimensional cross-correlation analysis of areas A-D of data set 2. On the images of the measurements done once per year (data set 1), migration could not be determined, because the net migration was more than one wave length and the shapes of the megaripples changed over time. The cross-correlation method was described previously in Duffy and Hughes-Clarke (2005) and Chapter 2, so for consistency it was used here as well (instead of a one-dimensional analysis per transect). Now, a 30×30 m search area was taken, in which a square matrix of 13×13 m (termed the fit matrix) was defined around each grid point. The location in the search area where the highest correlation was found between the fit matrices of two consecutive megaripple images was taken to be the new location of the fit matrix. The distance between the original and the new location of the fit matrix was defined as the migration distance of the specific bar region. When this procedure was finished for the entire grid, the migration values were averaged per transect, using the same transects as were used for all other megaripple characteristics, so one value was available for every 5 m.

3.3 Results

3.3.1 Megaripple behaviour on the time-scale of a single tide

Figure 3.6 shows the characteristics of megaripples along a cross-channel cross-section of area D during the time span of flood to ebb. The megaripple height ranged between 0.3 and 0.8 m, with the smallest values in the deepest part of the channel. The height variation in time was smaller than in space. In more detail, the height increased by up to 0.1 m from flood to ebb around $x = 200$ and $x = 500$ m, while in the shallowest part of the channel the height decreased by up to 0.2 m when the maximum ebb velocity was approached. This means that the heights decreased by 33% at most. In contrast, the megaripple length was much less affected by the tide. The spatial pattern was similar to that of the height, with larger values at the edges of the channel and the smallest values in the deepest part.

In the outer bend of the channel ($x = 0\text{--}300\text{ m}$) the asymmetry was always positive and in the range of 0.1–0.5 (Figure 3.6), indicating that the megaripples were slightly asymmetric. In the inner bend ($x = 300\text{--}1100\text{ m}$), the asymmetry increased with increasing depth. During the ebb phase A_s increased to larger negative values, while it increased to larger positive values during flood. Maximum values were 0.85 and -0.89 at the shallowest part of the channel, which means that the megaripples became increasingly saw-tooth shaped. The slip face angle ranged between 8 and 20 degrees, with the highest angles at the shallowest part of the channel, where the A_s was highest as well. Figure 3.6D shows that the direction of the slip face (i.e. an indication of the migration direction) changed almost simultaneously along the profile during measurement D7/D8 (near the peak of the ebb phase), while the current reversed after measurement D5 (see also Figure 3.3). This suggests a hysteresis effect: the slip face direction did not change immediately during flow reversal, but with a time lag of approximately 3 hours. This hysteresis effect is also shown in Figure 3.7A: the current velocity increased from -1.5 m/s to 1.4 m/s. At the same time, A_s increased until the current velocity reached 1.2 m/s, and only decreased after that, instead of decreasing at the same time that the current direction changes. Near the peak values of u_c , A_s approached zero, indicating more sinusoid bedforms. On the other hand, when current velocities were in between -1 en 1 m/s, A_s was relatively constant, and nearly independent of u_c .

Similar to the slip face angles, the migration did not change immediately with the current reversal (Figure 3.7B). Instead, migration velocity decreased gradually until u_c reaches 1.2 m/s, and increased more suddenly only after that. When the current velocity was in between 0 and 1.2 m/s, the migration of megaripples was negligible. Above this threshold, the maximum migration increased, although Figure 3.7B shows quite some scatter and combinations of high current velocities and low migration speeds also occurred. Thus, for increasing flow velocity the range in possible migration velocities increased as well.

Figure 3.6E shows that the largest values and variabilities in migration distance were found in the inner bend (closest to Ameland). On the other hand, a low variability was found in asymmetry values in the outer channel bend. Table 3.3 shows the width-averaged megaripple migration in m/h during the consecutive soundings, as well as the mean direction in which the migration took place. During flood (D1–D4), the megaripples migrated towards the southeast (approximately 50°), with the absolute migration decreasing with decreasing flood velocities. At slack water (D5–D6), migration was negligible, so the directions are not to be trusted. During the ebb phase (D7–D9), migration increased with increasing velocities, and the megaripples traveled towards the northwest. The migration direction was approximately 130° , which is the exact 180° opposite of the migration direction during flood.

Averaged over the channel, the net migration between maximum flood and maximum ebb was 0.45 m in the direction of the ebb. In absolute numbers, the spatio-temporally averaged migration in the ebb direction was 0.86 m larger than the migration in the flood direction. This is mainly caused by the large net migration in the inner bend of the channel ($x = 650\text{--}1000\text{ m}$), where net migration values of more than 4 m were found during half a tidal cycle (Figure 3.6E). In the first 300 m of the profile, net migration was in the flood direction, with maximum values of up to 2 m. Then, between $x = 300$ and $x = 450\text{ m}$, migration values were smaller and in the ebb direction. Between $x = 450$ and $x = 650\text{ m}$, net migration was up to 4 m in the direction of the flood.

Figure 3.6F shows that most of the profile along area D was flood-dominated. Only between $x = 600$ and 800 m , the modeled ebb flow was stronger. The largest difference between

Figure 3.6 Cross-section along area D. (A) megaripple height, (B) megaripple wave length, (C) asymmetry, (D) slip face angle, with negative angles meaning that the bedform points in the flood direction, (E) migration distance and direction: colored arrows are vectors per sounding and black line gives the total migration distance over the studied period, positive values are in the ebb direction and negative values are in the flood direction, (F) modelled depth-averaged current velocity: colored lines are values per sounding and black line gives the net ebb-flood value and (G) bed elevation profile z . Colours from red to green show the time periods from flood to ebb. Migration directions are slightly distorted due to the different scales on the x- and y-axis; actual directions can be found in Table 3.3.

Figure 3.7 A: As versus the current velocity u_c . B: Migration velocity versus the current velocity u_c . The grey dots represent each transect of area D, the black line is the average per Mb.

Table 3.3 Migration velocity and direction through time for area D. Directions are with respect to the east: 0° is towards the east, 90° is towards the north, -90° is towards the south and 180 and -180° are towards the west.

number	time at 7 Sep	migration [m/h]	migration direction [$^\circ$]
D1	7:54 - 8:26		
D2	8:36 - 9:27	1.39	-57
D3	9:37 - 10:17	0.66	-49
D4	10:24 - 11:08	0.11	-45
D5	11:13 - 11:52	0.06	-88
D6	11:57 - 12:31	0.05	8
D7	12:37 - 13:14	0.89	125
D8	13:18 - 14:13	1.37	127
D9	14:15 - 14:55	1.23	130

the ebb and flood flow was 0.45 m/s in favour of the flood flow, which was found around $x = 250 \text{ m}$. This is also the location where the strongest megaripple migration in the flood direction were observed (Figure 3.6F).

On the whole, the observations throughout one tidal cycle as presented in Figure 3.6 show that megaripple height, length and steepness (η/λ) do not react instantly to changes in depth and flow velocity. The shape and migration direction of the bedforms reversed after flow reversal, again not instantly but with a time lag. Thus, the slip face steepness and asymmetry in any multibeam image are only representative of the tidal phase during which the sounding took place.

3.3.2 Megaripple behaviour during a spring-neap tidal cycle

Figures 3.8, 3.9 and 3.10 show the change in megaripple characteristics over a neap to spring tidal cycle for areas A, B and C, respectively. It is unknown at which exact time the Mb data were collected, so the flow velocities plotted in Figures 3.8 - 3.10, and referred to in this section, are the peak ebb and peak flood velocities of the tidal cycle previous to the Mb survey time window. In general, the megaripple heights ranged between 0.1 and 1 m , and

the wave lengths ranged between 6 and 20 m. The megaripple height variability over time was about 0.1 m in all areas, but this variability appears to be unrelated to the spring-neap tidal cycle. The same holds for the wave length, where the variation in time was in the order of 2 m. The heights and lengths appear to be largest on the channel slopes, especially in area C. The average absolute value of A_s of the entire data set 2 was 0.34, indicating that the megaripples were relatively sinusoid. However, the standard deviation of A_s was 0.37, so the variability in space and time was large. Average slipface angles in areas A-D were 8.6° , with a standard deviation of 3.6° , which supports the observation that the megaripples were not saw-tooth like.

In area A (Figure 3.8), a non-natural disturbance was found in the bed between $x = 30$ and 130 m (possibly a shipwreck), so the megaripple height and A_s should not be trusted here. The largest temporal variability in heights was found at the shallowest part of the profile (6 m water depth). Here, heights varied between 0.15 and 0.35 m. The maximum wave length variation was 3 m (between 6 and 9 m), near the deepest part of the profile in 20 m water depth. The megaripples were relatively symmetrical with near-zero A_s and slip face angles between 5 and 8° . The predicted ebb-tidal current velocity was generally 0.1 m/s larger than the flood-tidal flow velocity, indicating that this part of the channel was ebb-dominant.

In area B (Figure 3.9), the range in wave lengths was most distinctive. While the maximum wave length of areas A, C and D were 10, 12 and 15 m, respectively, wave lengths in area B reached up to 20 m. The minimum wave length was found at $x = 130$ m, and was 5 m, so the range in wave lengths was 15 m. The megaripples were more asymmetric than in area A, with A_s values of up to 1 and maximum slip face angles of more than 10° . Across the entire area, the flood flow was dominant, with maximum differences between ebb and flood current velocity of 0.65 m/s at $x = 1300$ m.

Area C can be divided into two parts: a shallow platform with water depths of approximately 5 m ($x=0-1000$ m in Figure 3.10) and a channel wall where the water depth increases from 3 to 12 m between $x = 1000$ and 1600 m. On the platform, megaripples were similar to those in area A; relatively low ($\eta = 0.1-0.3$ m) and short ($\lambda = 6.5-8.5$ m). On top of that, they were also visually more three-dimensional than in the channel. Slip face angles on the platform were low and in the order of 5° and A_s ranged between 0 and 1. In the channel part of area C, the megaripple heights were twice as high (η up to 0.75 m) and wave lengths were longer as well (λ in the order of 12 m). Also, slip face angles of up to 20° were found. In the entire area, the current was flood-dominant, with $0.1-0.4$ m/s. The maximum flood dominance was found at $x = 1200$ m, which was near the location of the maximum slip face angles.

Table 3.4 shows the width-averaged migration rates in m/d for the megaripples in areas A, B and C. This is indicative for the migration per direction, although the actual migration rates are relatively uncertain because the actual time difference between two Mbes soundings is unknown. Yet, it is visible that the migration in area B was almost perfectly bidirectional along the channel axis. In area C, except for the first measurement the directions were similar to those of area B, although both the magnitude and the direction were slightly smaller. In area A, migration was either towards the northwest or towards the southwest, with the largest values towards the northwest. This confirms the ebb-dominance of the channel.

Figure 3.8 Cross-section along area A. (A) megaripple height, (B) megaripple wave length, (C) asymmetry, (D) slip face angle, with negative angles meaning that the bedform points in the flood direction, (E) migration distance and direction as vectors per sounding, (F) modelled depth-averaged current velocity of the previous ebb (positive) and flood (negative): colored lines are values per sounding and black line gives the net ebb+flood value and (G) bed elevation profile z . Colours from red to green show the time periods from neap to spring tide.

Figure 3.9 Cross-section along area B. (A) megaripple height, (B) megaripple wave length, (C) asymmetry, (D) slip face angle, with negative angles meaning that the bedform points in the flood direction, (E) migration distance and direction as vectors per sounding, (F) modelled depth-averaged current velocity of the previous ebb (positive) and flood (negative): colored lines are values per sounding and black line gives the net ebb+flood value and (G) bed elevation profile z . Colours from red to green show the time periods from neap to spring tide.

Figure 3.10 Cross-section along area C. (A) megaripple height, (B) megaripple wave length, (C) asymmetry, (D) slip face angle, with negative angles meaning that the bedform points in the flood direction, (E) migration distance and direction as vectors per sounding, (F) modelled depth-averaged current velocity of the previous ebb (positive) and flood (negative): colored lines are values per sounding and black line gives the net ebb+flood value and (G) bed elevation profile z . Colours from red to green show the time periods from neap to spring tide.

Table 3.4 Migration velocity and direction through time for areas A, B and C. Directions are with respect to the east: 0° is towards the east, 90° is towards the north, -90° is towards the south and 180° and -180° are towards the west.

date (dd/m)	A		B		C	
	v [m/d]	dir [$^\circ$]	v [m/d]	dir [$^\circ$]	v [m/d]	dir [$^\circ$]
30/8 - 31/8	0.78	135	1.78	175	0.29	-156
31/8 - 01/9	0.14	-135	0.82	-20	0.54	-15
01/9 - 02/9	0.14	-135	0.82	-20	0.54	-15
02/9 - 03/9	1.48	131	0.83	-20	0.54	-15
03/9 - 04/9	1.53	136	0.83	-20	0.65	-13
04/9 - 05/9	0.49	114	1.51	-12	1.22	-6
05/9 - 06/9	0.49	114	1.51	-12	1.22	-6

3.3.3 Megaripple characteristics over the years

Since the previous sections established that the asymmetry and slip face direction are dependent on the moment when the measurement was done, the yearly surveys could not provide any more insight into these parameters. Instead, only the megaripple height and wave length were determined, and their evolution in space and over time was studied (Figure 3.11).

The wave lengths in data set 1 (the yearly data) were between 2 and 28 m, with an average of 12 m and a standard deviation of approximately 4 m; this range is much larger than that of data set 2, but the average is similar to the average of areas B and D. The average of both data sets together was 11.2 m, with a standard deviation of 6.7 m. The bedform heights of data set 1 were always in between 0.05 and 1.4 m (with a mean of 0.48 m and a standard deviation of 0.1 m), although they are mostly below 1 m. The average of both data sets was 0.46 m, with a standard deviation of 0.17 m.

Spatially, on the other hand, a larger variability was found. The smallest heights and lengths were found in the northernmost part of the survey area, and they both increased towards the back-barrier basin. Landward of area 5, which is the deepest part of the channel and the boundary between the basin and the sea, the variability in lengths remained more or less constant. Temporally, both the range in megaripple lengths and in heights were relatively constant, and no trend could be found.

It appears that the lengths in areas A-D were in the lower range of the observations in data set 1. The locations of areas 4 and 7 of data set 1 were closest to areas A and D of data set 2, so their characteristics could be compared. The heights in both cases were similar, they always varied between 0.2 and 0.75 m. The lengths in A and D, however, were in between 5 and 10 m, while in areas 4 and 7 they could become up to 20 m (Figure 3.11). Yet, the depth variation in time was also larger in area 7 (up to 6 m) and area 4 (up to 5 m), which could explain the larger range in lengths. The variation in both height and length seems to be random over time. This was also true for almost all other areas. Only in area 12, the most southern area inside the Wadden Sea basin, the heights and lengths increased over the years.

3.3.4 Relations between megaripple properties and hydrodynamics

All megaripple characteristics and the environmental conditions were compared to each other using linear regression. For data set 1 the characteristics that were considered were megaripple height and length, water depth and the slope of the seabed, and for data set 2 we compared megaripple height and length, A_s , slipface angle, migration, current velocity

Figure 3.11 Megaripple height (left) and wave length (right) for all transects through areas of data set 1, over space (upper plots) and in time for area 7 (lower plots). The red line in each box indicates the median, and the bottom and top edges of the box indicate the 25th and 75th percentiles, respectively. The whiskers extend to the most extreme data points not considered outliers, and the outliers (more than 1.5 times the interquartile range away from the top or bottom of the box) are plotted using the '+' symbol.

u_c , ebb- or flood dominance ($u_{c,flood} - u_{c,ebb}$), water depth and the slope of the seabed. Five significant relations were found.

First, the relation between megaripple height and length as found by Flemming (1988) was observed here as well (Figure 3.12). All data sets overlapped, although especially the range in wave lengths was twice as large for data set 1 than for data set 2. The slope of the relation was larger for data set 2 than for data set 1. The slope of the line for data set 1 was 0.012, which is 3 times smaller than the one found by Flemming (1988), and even approaching the horizontal with a relatively small correlation coefficient of 0.34 (Table 3.5). This slope was the same for all sub-areas of data set 1. For data set 2, the relation between η and λ was not significant in area A, with r 0.073 and $p > 0.05$. The other areas of data set 2 were all significant, with r values of 0.76 and higher and p values of 0. The slope of the linear regression of area B was similar to the mean slope found by Flemming (1988), while the slope of the linear regression of area C was more similar to the maximum slope found by Flemming (1988) and the slope of the linear regression in area D was in between these two (Table 3.5).

Second, a positive linear relation was found between the megaripple height and the slope of its lee side (Figure 3.13). This relation can be explained by the fact that the lengths of the megaripples were relatively less variable than the heights. In other words, when the length

Figure 3.12 Megaripple height plotted against wave length for all areas (locations shown in Figure 3.1). Lines show the mean and maximum values as found by Flemming (1988).

Table 3.5 Coefficients of linear relations between η and λ for all data sets. Slope gives the slope of the linear regression, r gives the correlation coefficient and p gives the p-value.

data set	slope	r	p
Flemming (1988) mean	0.034	-	-
Flemming (1988) max	0.093	-	-
Data set 1	0.012	0.34	0
Data set 2A	0.011	0.073	0.012
Data set 2B	0.037	0.84	0
Data set 2C	0.089	0.76	0
Data set 2D	0.05	0.76	0

Figure 3.13 Absolute value of the megaripples' leeside slope plotted against megaripple height for all areas (locations shown in Figure 3.1).

Table 3.6 Coefficients of linear relations between the slipface slope and η for all data sets. Slope gives the slope of the linear regression, r gives the correlation coefficient and p gives the p -value.

data set	slope	r	p
Data set 1	4.85	0.42	0
Data set 2A	5.11	0.48	0
Data set 2B	3.75	0.55	0
Data set 2C	24.05	0.95	0
Data set 2D	13.96	0.76	0

remains constant but the height increases, the bedform automatically becomes steeper. Curiously, the ratio of slope over height was different for different areas: it was similar for areas A and B, and for C and D. The correlation coefficient r was significantly higher as well for areas C and D. Data set 1 was most similar to A and B (Table 3.6). It is unclear why these ratios were different, as the difference appears unrelated to water depth, wave length, current velocity or velocity differences.

Finally, it was found that both the megaripple height and its asymmetry, and to a lesser extent the migration as well, were negatively linearly correlated to the ebb- or flood dominance of the system (all p values are 0) (Figure 3.14). In none of the studied areas the flood was clearly dominant. The largest megaripple heights were found when the ebb current was strongest, while the lowest heights were found when the ebb and flood current were equal. This correlation was strongest for areas B and C, with r values of -0.65 and -0.56, respec-

Figure 3.14 (A) Megaripple height, (B) asymmetry and (C) net migration versus ebb-flood dominance in areas A, B and C (colours).

tively. For area A, the r value was -0.15, indicating that the correlation is weak. Similarly, the megaripples became more asymmetric when either ebb or flood was dominant, with As near zero when the currents were about equal. This is because a more unidirectional flow can move the megaripple more significantly in a certain direction, thus making it more saw-tooth shaped. Again, this correlation was strongest for areas B and C (r of -0.43 and -0.48, respectively), and weak for area A ($r = -0.10$). The correlation between ebb- or flood dominance and migration was relatively weak for all three areas, with $r = -0.20$, -0.28 and -0.34 for areas A, B and C, respectively. Yet, it is still clear that for increasing ebb-dominance in the flow velocity, the range in possible migration distances increased as well, similar to what was observed for area D (Figure 3.7).

3.4 Discussion

3.4.1 Megaripple characteristics and dynamics

The results of this study have shown that the megaripple heights and lengths in the Ameland inlet are similar to those of previous studies. For example, Dalrymple et al. (1978) found the same range of lengths and heights in a tidal inlet, although water depth in their case was only 6 m. In a tidal inlet with water depth of 12 m, Bartholdy et al. (2005) found smaller megaripples, with maximum heights of 0.5 m and maximum lengths of 12 m. Elias (2018) studied megaripples from the Mbes surveys of the Ameland Inlet channel of 2012 and 2017 by calculating megaripple height, length, asymmetry and migration based on profiles taken through the measured sections. Mean heights (calculated as difference between trough and crest) ranged from 0.2 to 1 m, while lengths (trough to trough) ranged between 5 and 21 m. This is highly similar to the results of the present study, which indicates that both methods of calculating bedform height and length give the same results. Elias (2018) further found that bedform heights and lengths were more or less constant over time, but the asymmetry and migration can change under locally varying current magnitudes and directions. Moreover, Elias (2018) found only small bedforms in the deepest part of the inlet, where flow velocities are highest, which do not seem to migrate over one tidal cycle. The larger bedforms were

found at the edges of the channel, and they also migrated faster. The present study partly confirms this. Regarding the bedform heights, our observations are similar, but even in the deepest part of the tidal inlet a net migration of 1 - 2 m was found in the period between maximum flood and maximum ebb.

On the other hand, the steepness (η/λ) as measured in the present study was often smaller than found and predicted by Flemming (1988), especially in data set 1 and in area C of data set 2. Following the hypothesis of Flemming (1988), this might be caused by a lack of sediment in the system, which results in preserved lengths but smaller heights. In field conditions, a lack of sediment might be caused by a hard, non-erodible layer near the bed surface. In the Ameland Inlet, such a hard clay layer was indeed found at approximately 23 m water depth, which coincides with the bottom of the inlet channel (Forzoni et al., 2018). A second explanation for the relatively low heights is a lack of time for the formation of megaripples because the current directions change four times a day. This means that the hydrodynamic conditions never persist long enough for the megaripples to reach their maximum height. This hypothesis is supported by Figure 3.14, which showed that a stronger ebb- or flood dominance leads to higher megaripples. In this case, megaripple creation is only effective during the dominant tidal current. When the ebb and flood flow are equally strong, they counteract each other, resulting in smaller bedforms.

The megaripple characteristics did not change immediately with the hydrodynamics. A decrease in migration velocity was observed in area D of data set 2 during an increase in current velocity. Furthermore, As only changed when the current velocity exceeded 1.2 m/s. These observations show that the adaptation of the ripple shape and migration only starts when the current velocity exceeds a threshold. In the present study, this threshold is approximately 1.2 m/s. Over an ebb-flood cycle or a spring-neap tidal cycle, variations in height and length are in the order of 0.2 and 2 m, respectively. Over the years, large variations in especially the wave length are found; the standard deviation was 4 m, which is 33% of their mean value of 12 m. Yet, no systematic change over time is found. Spatially, the height and length are largest in the inlet throat and decrease both towards the seaward side and into the basin.

One of the main problems addressed in Section 3.1, was the fact that the behaviour of megaripples in wave-current dominated environments was unknown. Figure 3.15 shows the ratio between wave- and current-related bed shear stress as predicted by Delft3D for dataset 2 during the time of the Mbes measurements. This bed shear stress is a measure of wave- or current dominance and is calculated as follows according to Kleinhans and Grasmeijer (2006):

$$\theta_x = \frac{\tau_x}{(\rho_s - \rho_w)gD_{50}}. \quad (3.2)$$

In this equation, $\rho_s = 2650 \text{ kg/m}^3$ and $\rho_w = 1025 \text{ kg/m}^3$ are the sediment and water density, respectively, g is the gravitational acceleration and D_{50} is the median grain size of the sediment. τ is the bed shear stress magnitude, with the subscript x referring to either currents (c) or waves (w),

$$\tau_c = \frac{\rho_w g u_c^2}{18 \log\left(\frac{12h}{2.5D_{50}}\right)^2} \quad (3.3)$$

and

$$\tau_w = 0.5 \rho_w f_w u_w^2. \quad (3.4)$$

Figure 3.15 Modelled wave- versus current-related bed shear stress for areas A, B, C and D, averaged over time. Black lines indicate sheetflow and incipient motion. Red dashed lines show the boundaries between regions of wave or current dominance, according to Kleinhans (2005).

In these equations, h is the water depth, u_c is the modelled time- and depth-averaged current velocity and u_w is the amplitude of the modelled near-bed orbital velocity. The friction coefficient f_w is

$$f_w = \exp(-5.977 + 5.213(\frac{A_\delta}{2.5D_{50}})^{-0.194}), \quad (3.5)$$

in which A_δ is the near-bed orbital excursion according to linear theory.

$$A_\delta = \frac{u_w T_{m-10}}{2\pi}, \quad (3.6)$$

with T_{m-10} the modelled spectral wave period. Figure 3.15 shows that area D and most of area A were current dominated. These areas were located in the main inlet channel, so most parts were too deep for waves to reach the bed. Areas B and C were located in secondary channels, which were less deep and were clearly dominated by both waves and currents. Yet, the differences in megaripple characteristics or behaviour could not be related to the dominance of waves or currents.

3.4.2 Predictors

To test how well bedform predictors perform, we will now compare the measured megaripple characteristics to those predicted by van Rijn (2007). In Delft3D, the van Rijn (2007)

predictor for megaripple-related roughness is used (Equation 3.7). As stated in Section 3.1, the roughness according to this predictor is always in between 0.04 m and 0.01 times the water depth, with a maximum of 0.4 m. For median grain sizes larger than 93 μm , megaripple heights in between these maximum and minimum values are given by $2k_{s,mr}$. Here, $k_{s,mr}$ is the megaripple roughness height, calculated by:

$$k_{s,mr} = \begin{cases} 0.0002\psi_{wc}h & \text{if } \psi_{wc} \leq 50 \\ (0.011 - 0.00002\psi_{wc})h & \text{if } 50 < \psi_{wc} \leq 550 \\ 0.02 & \text{if } \psi_{wc} > 550, \end{cases} \quad (3.7)$$

with h the water depth in m, mobility parameter $\psi_{wc} = u_{wc}^2 / ((s - 1)gD_{50})$, with $(s - 1)$ the dimensionless submerged density of quartz (1.65), g the acceleration due to gravity (9.81 m s^{-2}), and D_{50} the grain size in m (in this case 216 μm). $u_{wc} = \sqrt{u_w^2 + u_c^2}$. Since the maximum $k_{s,mr}$ equals 0.2 m, which represents a megaripple height of 0.4 m, it can a priori be expected that the predicted roughness will often be too low, as the present study found megaripple heights of up to 1 m.

Figure 3.16 shows that the predicted roughness heights are not only lower than estimated from the measured heights, but also that the variability in time is much higher. This is probably caused by the conditions for which the predictor was created, which might have been exposed to more unidirectional currents. The present study confirmed that megaripples were lower when the currents were more bidirectional, so when ebb and flood are of similar strength (Figure 3.14). In Delft3D, the modelled megaripple roughness shows 4 cycles during one day related to the ebb and flood flows, while the measured megaripple heights and lengths in area D did not change as quickly, and did not respond directly to the current velocity. While the measured roughness gradually increased over time, between 0.16 and 0.19 m, the predicted roughness height increased from 0.02 m to 0.13 m, and then decreased to 0.02 m again. The megaripple roughness was used to calibrate the predicted water levels to measurements (see Chapter 5). The roughness that caused the best representation of hydrodynamics after calibration was a factor of 2 smaller than the outcome of Equation 3.7, but still varied in the same relative amount. It is common practice to use the roughness as a fitting parameter in model calibration. The fact that observed and calibrated roughness are so different, suggests that either the model is right for the wrong reasons, or that megaripple height is influencing roughness height in a different way. Instead of a factor 0.5 maybe roughness height is related to bedform height by a factor 0.25.

Additionally, the direction of the lee side angle with respect to the flow is important. In this study, we have shown that the megaripples still point in the flood direction during a large part of the ebb phase. Therefore, the flow will encounter a lower resistance during the beginning of the ebb phase (and probably also during the beginning of the flood phase), because the lee sides are still gentle from the previous tidal phase. This concept was described for sand waves by Lau (1988) and Lefebvre et al. (2013). For megaripples the effect will be lower, because the bedforms themselves are lower, but the concept is still valid. This implies that the roughness as experienced by the flow will vary at least twice during a tidal cycle, which might be the reason that the variable roughness predictor of van Rijn (2007) is adequate for practical purposes, even though the actual megaripple heights are not predicted very well. In other words, Equation 3.7 may represent the roughness relatively well, but the roughness is not equal to half the bedform height.

Figure 3.16 Megaripple-related roughness ($k_{s,mr} = \eta/2$) as measured in the present study, and as predicted by Bartholdy et al. (2005) (B05) and (van Rijn, 2007) (VR07) at the center of area D over time. Red: default Van Rijn (2007) prediction, yellow: (van Rijn, 2007) prediction which resulted in the most accurate water level prediction, with $k_{s,mr}$ divided by 2. Top: dd/mm in 2017, bottom: hh:mm on September 7, 2017, see also Figure 3-3.

There might be various other reasons that a numerical model that incorporates van Rijn (2007) still performs well even if the megaripple characteristics are predicted incorrectly. For example, the roughness at peak current velocities might be of much greater importance than the roughness during other parts of the tidal cycle, because most energy dissipation and sediment transport occurs during the highest flow velocities. When the roughness at this moment in time is predicted correctly, the associated sediment transport might still be correct, and the roughness variation over time is of less importance to the predicted sediment transport. Yet, in this case it would be expected that the spring-neap tidal cycle is less well represented in the sediment transport patterns. This should be considered in future research.

We also compared our observations to the predictor of Bartholdy et al. (2005), who used curve fitting to find the following relations between grain size and megaripple height η and length λ :

$$\eta = 0.17 D_{50,\varphi}^{1.68} \quad (3.8)$$

$$\lambda = 7.90 D_{50,\varphi}^{0.68}, \quad (3.9)$$

with $D_{50,\varphi}$ given in dimensionless φ units: $D_{50,\varphi} = -\log_2 D_{50}$, with D_{50} in mm. Based on Equations 3.8 and 3.9 and the measured grain size of $216 \mu\text{m}$, the expected megaripple height

and length are 0.64 m and 13.55 m, respectively. Especially the wave length is very close to the measured average wave length. The predicted height is relatively high, but still within the range of the measured values. These equations are not consistent with the substantial spatial and temporal variation in megaripple height and length, but they do give a good first estimation of the characteristics and therefore also of the roughness.

Similar to the prediction of van Rijn (2007), η , λ and the megaripple-related bedload transport are related to the mobility parameter ψ_c , using the friction velocity u_f instead of wave-current related velocity u_{wc} :

$$u_f = U/[8.5 + 2.5 \ln(4/[2.5D_{50} + 0.187D_{50,\varphi}^{1.68} \cdot (1 - \exp(-0.538D_{50,\varphi})))]], \quad (3.10)$$

with U the flow velocity 4 meters above the bed. Thus, while the predictor of van Rijn (2007) depends on both waves and currents, the Bartholdy et al. (2005) predictor depends only on tidal currents. Using ψ_c , Bartholdy et al. (2005) further stated that η and λ can also be calculated as

$$\eta = -0.24\psi_c^4 + 0.61\psi_c^2 + 0.10 \quad (3.11)$$

$$\lambda = 6.42\psi_c^6 - 15.81\psi_c^4 + 15.06\psi_c^2 + 6.22. \quad (3.12)$$

From Bartholdy et al. (2005), the following equation for the megaripple migration rate c in m/d can be deduced as well:

$$c = 79.74((0.4\psi_c^2 + 0.013)D_{50})^{3/2}/\eta. \quad (3.13)$$

Equations 3.11 - 3.13 predict megaripple heights between 0.1 and 0.3 m (roughness between 0.05 and 0.15 m, see Figure 3.16), wave lengths between 6.2 and 11.4 m and migration speeds between 0.12 and 3.4 m/d, using the predicted current velocities of 0.01 to 2 m/s with the measured grain size. Both the wave length and the height are in the same order of magnitude as the measurements, although the measured heights and lengths can both become twice as large as in these predictions. The predicted migration speed is in the right order of magnitude as well. On the other hand, similar to results of the Van Rijn (2007) predictor, the predicted characteristics show much more variation in time than the measurements, because the predictions depend on the current velocity (Figure 3.16). A spring-neap cycle effect is visible as well; the roughness height variation is much smaller during neap tide (0.01 m) than during spring tide (0.05 m).

It is important to consider that the Bartholdy et al. (2005) equations were developed for an area with grain sizes of 300-600 μm . Even though the present study area consists of sediment with grain size around 220 μm , the range in megaripple heights and lengths is exactly the same as found in their study.

It is clear that the van Rijn (2007) and the Bartholdy et al. (2005) predictors disagree with many of the observations in the present study. Future research should focus on a better way of predicting the megaripple roughness and its effect on sediment transport. The most important factor in this is the relation between the megaripples and the hydrodynamics. The present study has shown that megaripple characteristics depend on the asymmetry of the flow, instead of the instantaneous current velocity. This should be incorporated in future predictors.

3.4.3 Sediment transport

The present study has shown that the net migration of the megaripples during half a tidal cycle can become up to 4 m, so in absolute terms throughout a year they might move a considerable amount of sand. As stated in Chapter 2, the direct amount of volumetric sediment transport by bedforms per unit width per unit time (q_b in m^3/y) can be calculated using $q_b = 0.5 \times U\eta(1-\varphi) + C_1$ (Simons et al., 1965). In this equation, φ denotes the porosity of the sediment, which is generally taken to be 0.4. C_1 is a constant, representing the part of the bed load that is not transported by bedforms. For now, C_1 is assumed to be 0, so that only the bedform-related bed load transport is calculated.

Averaged over both data sets, η was 0.46 m (standard deviation of 0.17 m). The average migration over half a tidal cycle in area D was 0.45 m, so 0.45 m per 6 hours and 12 minutes. This equals 0.072 m/h and 635 m/y. These η and U values result in a volumetric sediment transport of 88 $\text{m}^3/\text{m}/\text{y}$. The maximum values in area D of $\eta = 0.8$ m and $U = 4$ m per 6 hours and 12 minutes (0.64 m/h) result in q_b of 2258 $\text{m}^3/\text{m}/\text{y}$.

Even though the differences in flow velocity between ebb and flood are only in the order of 0.1 m/s in most cases, this might significantly affect the suspended load sediment transport q_s , because q_s is in the order of velocity to the third power. The megaripples might affect the suspended load transport as well, for example by affecting the total bed roughness. Two of the parameters that could be of influence in the effective roughness are the bedform steepness (η/λ) or the slip face angle. Previous research based on numerical modelling by Lefebvre and Winter (2016) shows that a steep slip face can cause a flow separation zone behind the bedform crest, which has a significant effect on the bedform-related roughness and the suspended load transport. In their study, the minimum dimensions at which this flow separation occurred were a lee side angle of 19° and $\eta/h = 0.05$, with h the water depth. In the present study, megaripples with these characteristics were found as well (e.g. Figure 3.6, at $x=900-1000$ m), although lee side megaripple angles larger than 19° are rare in the study area and are only found in areas C and D (Figure 3.13). In area A, η/h ranges between 0.01 and 0.46, but in areas B, C and D, maximum values for η/h are 0.07, 0.11 and 0.10, respectively. Since the bed roughness is an important parameter in sediment transport predictors, more research should be done on this subject. Lastly, the megaripple roughness should be considered in combination with the roughness of the underlying bathymetry, because the 'real' slipface angle of the megaripples might be higher than 19° when the megaripples are located on the lee side of a sand wave.

3.5 Conclusions

Megaripples are observed from the deepest part of the main inlet channel to the shallower parts surrounding the secondary channels of the Ameland inlet, implying they are ubiquitous bedforms in the entire inlet system. Wave lengths range between 2 and 28 m, with an average of 11.2 m and a standard deviation of 6.7 m. Heights vary between 0.05 and 1 m, with an average of 0.46 m and a standard deviation of 0.17 m. On average, the megaripples were relatively sinusoid, with average A_s of 0.34 and average slipface slopes of 8.6° . Their dimensions are relatively constant over shorter periods of time (days to weeks), and the steepness (height divided by length) is also relatively constant in space. Over longer periods of time, i.e. years, especially the wave lengths change, but this could not be related to water depth or bed slope, and due to a lack of hydrodynamic data also no relation with hydrodynamic forcing could be

identified. Apart from the migration, megaripple characteristics could not be related to local instantaneous hydrodynamics. Instead, the height and asymmetry are linearly related to the ebb- or flood dominance over a tidal cycle; they increase when there is a stronger residual current. Megaripples migrate in the direction of the flow, which reverses during the tidal cycle. The bedform asymmetry and the migration velocity reverse only when the flow exceeds 1.2 m/s after reversal. This means that during the first part of the tidal cycle, the flow encounters less resistance, because the megaripples are still pointing in the direction of the previous tidal phase. The predictors that were studied are highly dependent on the instantaneous current velocity, and therefore predict much more variability in megaripple heights over time than measured. Furthermore, the predicted heights are a factor 2 lower than the measured heights. Megaripples directly transport sediment by migrating, in the channel average direct sediment transport was 88 m³/m/y. Indirectly, they affect the bed roughness, which mainly affects the suspended load transport.

Chapter 4 | Spatio-temporal characteristics of small-scale wave-current ripples on the Ameland ebb-tidal delta

Abstract

Ebb-tidal deltas are highly dynamic environments affected by both waves and currents that approach the coast under various angles. Among other bedforms of various scales, these hydrodynamics create small-scale bedforms (ripples), which increase the bed roughness and will therefore affect hydrodynamics and sediment transport. In morphodynamic models, sediment transport predictions depend on the roughness height, but the accuracy of these predictors has not been tested for field conditions with strongly mixed (wave-current dominated) forcing. In this study, small-scale bedforms were observed in the field with a 3D Profiling Sonar at five locations on the Ameland ebb-tidal delta, the Netherlands. Hydrodynamic conditions ranged from wave-dominated to current-dominated, but were mixed most of the time. Small-scale ripples were found on all studied parts of the delta, superimposed on megaripples. Even though a large range of hydrodynamic conditions was encountered, the spatio-temporal variations in small-scale ripple dimensions were relatively small (height $\eta \approx 0.015$ m, length $\lambda \approx 0.11$ m). Also, the ripples were always highly three-dimensional. These small dimensions are probably caused by the fact that the bed consists of relatively fine sediment. Five bedform height predictors were tested, but they all overestimated the ripple heights, partly because they were not created for small grain sizes. Furthermore, the predictors all have a strong dependence on wave- and current-related velocities, whereas the ripples heights measured here were only related to the near-bed orbital velocity. Therefore, ripple heights and lengths in wave-current dominated, fine grained coastal areas ($D_{50} < 0.25$ mm), may be best estimated by constant values rather than values dependent on the hydrodynamics. In the case of the Ameland ebb-tidal delta, these values were found to be $\eta = 0.015$ m and $\lambda = 0.11$ m.

Based on:

Laura Brakenhoff, Maarten Kleinhans, Gerben Ruessink and Maarten van der Vegt (2020), Spatio-temporal characteristics of small-scale wave-current ripples on the Ameland ebb-tidal delta *Earth Surface Processes and Landforms*, 45, pp 1248– 1261, DOI:10.1002/esp.4802.

4.1 Introduction

Ebb-tidal deltas are subtidal bodies of sand located seaward of tidal inlets, which makes them subject to both waves and cross-shore and longshore currents. Interaction between these hydrodynamics and the sandy bed creates bedforms on a variety of scales. The largest bedforms on ebb-tidal deltas are sandy shoals, which have been studied thoroughly by, for example, FitzGerald (1982) and Ridderinkhof et al. (2016b). In the Wadden Sea region, saw-tooth bars are often present on the downdrift side of ebb-tidal deltas, with heights up to 2 m and wavelengths (i.e. spacings) of about 700 m (Chapter 2). The characteristics of smaller bedforms on ebb-tidal deltas, like ripples and megaripples, are less well studied, despite their relevance for sediment transport.

Megaripples (or dunes) are defined as bedforms with height $\eta > 0.1$ m and wavelength $\lambda > 0.6$ m (Ashley, 1990), while small-scale ripples have η in the order of centimeters and $\lambda < 1$ m (Jackson, 1975). Previous studies have focused on megaripples and other similarly-scaled bedforms in the current-dominated channels and tidal inlets (e.g. Buijsman and Ridderinkhof, 2008a) or on the more wave-dominated intertidal shoals (Hoekstra et al., 2004). Passchier and Kleinhans (2005) observed three-dimensional megaripples on the Dutch continental shelf in 14–18 m water depth when the current- and wave-related Shields numbers were both above 0.1. Megaripples in the Ameland inlet were studied in Chapter 3.

Small-scale wave-current ripples have not been studied for the mixed forcing conditions of ebb-tidal deltas either. Studies do exist on wave-current ripples in shallow water along straight coasts or on nearshore bars (e.g. Meirelles et al., 2016; Wengrove et al., 2018; Wu and Parsons, 2019) and on continental shelves (e.g. Li and Amos, 1999; Krämer and Winter, 2016). In most cases, rather two-dimensional ripples were observed, but that might be caused by either the constant angle between waves and currents, or by the fact that, although waves and currents were both present, one of the two was still clearly dominant. In flume studies it was found that it is possible for ripples to become three-dimensional when waves and currents are perpendicular to each other (Andersen and Faraci, 2003). Li and Amos (1999) even found three-dimensional ripples when the angle between waves and currents was 30° . Yet, it is unclear how these flume studies relate to the dynamic conditions of ebb-tidal deltas.

Bedforms like ripples affect both hydrodynamics and sediment transport. They create additional bed roughness on top of the grain-related roughness, which affects the vertical mixing in the flow (Smith and McLean, 1977; Nielsen, 1979; Lefebvre et al., 2013; Lefebvre et al., 2014). Models that are used to estimate sediment transport, like Delft3D, TELEMAC and MIKE21, are depending on roughness predictors to calculate sediment transport (Tonnon et al., 2007; Elias and Hansen, 2013; Herrling and Winter, 2014; Lenstra et al., 2019; McCann et al., 2011; Deigaard et al., 1999). This means that it is important to correctly predict the presence and height of the bedforms. The bedform roughness height predictor of van Rijn (2007) is incorporated in the frequently used model Delft3D, Tanaka and Dang (1996) is incorporated in TELEMAC and MIKE21 uses Nielsen (1979).

For a predictor to correctly represent the mixed hydrodynamic situation of ebb-tidal deltas, it should incorporate both waves and currents. Predictors that meet this condition are, for example, Tanaka and Shuto (1984), Tanaka and Dang (1996), Khelifa and Ouellet (2000) and van Rijn (2007). However, Tanaka and Shuto (1984) and Tanaka and Dang (1996) were created for the specific condition in which waves and currents are opposing each other (180° difference in direction). Khelifa and Ouellet (2000) do account for all possible angles, but the

predictor was only tested for laboratory conditions. Furthermore, both Khelifa and Ouellet (2000) and van Rijn (2007) calculate the equilibrium height of the bedforms. Because of the highly variable hydrodynamic conditions on ebb-tidal deltas, it can be questioned whether this equilibrium height will ever be reached.

Various recent studies show that it is important to incorporate time-dependent aspects in the calculation of bedform height, including the previous bed state (e.g. Davis et al. (2004), Nelson and Voulgaris (2014), Soulsby et al. (2012), and Traykovski (2007)). Soulsby et al. (2012) is the only predictor that incorporates both waves and currents and the time-variance. Wengrove et al. (2018) tested the Soulsby et al. (2012) predictor in shallow water (1–3 m) and found an r^2 of 0.55 for predicted ripple heights. Soulsby et al. (2012) tested their predictor in the field in 7-m deep continental shelf conditions, and found that 88% of the predicted ripple heights are within a factor 2 of the measured ripple heights. This is higher than all predictors tested by Camenen (2009), although he studied only wave ripples and did not test the predictors mentioned above. So far, none of the predictive formulas for bedform geometry have been tested for the wave-current dominated conditions on ebb-tidal deltas.

The main goal of this study is to better understand the characteristics of small-scale wave-current dominated ripples. Therefore, their height, length and three-dimensionality were measured on an ebb-tidal delta (Ameland, The Netherlands), during a large range of hydrodynamic conditions. In this paper, the ripple characteristics are analysed and related to the hydrodynamic forcing. The measured ripple heights are also compared to those estimated with predictors used in, for example, models like Delft3D, TELEMAC and MIKE21.

4.2 Data and methods

4.2.1 Study site

The ebb-tidal delta of the Ameland Inlet, which is located in the Dutch part of the Wadden Sea, was chosen as study site, as it is regarded as the least disturbed inlet in the Netherlands (Elias et al., 2012). The Ameland inlet alternates between a two-channel and a single channel configuration (Israel and Dunsbergen, 1999). Currently two channels are present, but one is clearly dominant (Figure 4.1). The delta itself consists of two lobes, one in the east and one in the west, with minimum water depths of approximately 1 m relative to mean sea level. It is subjected to wind-generated waves approaching mainly from the northwest (Figure 4.1), with an average significant height H_s of 1.3 m, and an average spectral period T_{m01} of 5 s (Roskam, 1988). Maximum significant storm wave heights can be more than 6 m, with storm surges of over 2 m. The tidal wave propagates along the coast from west to east, and near Ameland the tidal range is ≈ 2 m. Maximum shore-parallel velocities are 0.5 to 1.0 m/s (Elias et al., 2012), and the tidal prism is 430 Mm³ (Sha, 1989). Across the entire ebb-tidal delta, the mean grain size (D₅₀) is 0.211 mm, with a standard deviation of 0.03 mm and a mud content (< 0.063 mm) of 1% by volume (Pearson et al., 2019).

4.2.2 Campaign

Small-scale ripples and hydrodynamics were studied with measurement frames offshore (frame 1), in the inlet throat (frame 3) and on three parts of the ebb delta lobe (frames 2, 4 and 5) (Figure 4.1). A sketch and a photo of one of the frames are given in Figure 4.2. The water depths at which the frames were deployed, can be found in Table 4.1. Frames 2–5 were deployed between 29 August and 10 October 2017, and frame 1 was deployed between

Figure 4.1 Bathymetry of the Ameland ebb-tidal delta of 2017 (measured by Rijkswaterstaat), including the locations of multibeam measurements (red squares) and the measurement frames (black dots). A wave rose is plotted in the upper right, showing the wave climate between 1996 and 2019 at an offshore wavebuoy located approximately 40 km northeast of the study site.

8 November and 11 December 2017. To obtain an impression of the large-scale megaripples that could not be captured by the frames, areas of 500×500 m around frames 2-5 were mapped with a multibeam echo-sounder before and after the campaign. The exact mapping dates are given in Table 4.1. As visible in Figure 4.1, frame 3 was placed outside of the multibeam area, but the multibeam data of this area are assumed to be representative of the frame location as well. Before the field campaign, megaripples were present at all measured sites (Figure 4.3). Visually, near frames 2 and 3, they were quite regular, with heights of 0.3 - 0.4 m and lengths in the order of 10 m. Frames 4 and 5 were located in an area with more irregular megaripples, which had smaller heights (≈ 0.1 m) and larger wavelengths (≈ 20 m). After the field campaign, the megaripples at frames 2, 4 and 5 had largely disappeared; at frame 2 the height decreased to 0.05 m. Near frame 3, however, the megaripple heights and lengths were the same before and after the field campaign. The megaripple dimensions are in the same range as those found in Chapter 3.

4.2.3 Equipment

All frames were equipped with a Marine Electronics type 2001 3D profiling Sonar, mounted at approximately 0.9 m above the bed to obtain regular maps of small-scale bedforms beneath the frame. It was set to scan the bed once per hour, rotating around its own axis, which took approximately 15 minutes. Each scan consisted of 200 swaths of the bed in a 360° circle,

Figure 4.2 Sketch (top) and photo (bottom) of one of the five measurement frames.

thus all swaths were 0.9° apart. The maximum deviation of the Sonar beam was 75° from the vertical. A single swath contained 166 vertical backscatter profiles, so the profiles were spaced 0.9° apart. Each profile consisted of 842 samples between 0 and 3.7845 m from the Sonar head, so the vertical resolution (in the direction of the beam) was approximately 4.5 mm. The horizontal resolution for each swath was smallest right below the Sonar head (0.016 m) and increased to 0.22 m at the edges of the swath (>3.5 m away from the centre of the image). To prevent resolution issues in studying small-scale ripples, only the inner part of the scanned sea bed, a circle with a radius of 1 m was used here, which gave a maximum horizontal spacing between measurements of 0.03 m.

Each frame also contained two Acoustic Doppler Velocity meters (ADV) to measure wave- and current related velocities and directions. Their measurement volumes were located approximately 0.35 and 0.65 m above the bed. Frames 4 and 5 contained an extra ADV

Figure 4.3 The bed state around frames 2, 4 and 5 before (A, C, E) and after (B, D, F) the field campaign. The measurement frames are indicated with black dots. All plots are 100×100 m. The vertical lines in plots B, D and F are measurement artefacts of the multibeam swaths, which could not be corrected.

at 1 m above the bed. The ADVs at frames 1-4 (Nortek Vector) sampled continuously at 16 Hz, whereas the ADVs at frame 5 (Sontek Hydra) sampled continuously at 10 Hz. On frames 1 - 4 a pressure transducer was incorporated in each ADV, and on frame 5 a separate pressure transducer was used, sampling continuously with a measurement frequency of 4 Hz. With

these transducers, water depth, wave height and wave period were measured. The ADVs at frame 3 were subject to technical issues and most hydrodynamic data are missing for this frame. As visible in Figure 4.2, the frames were set up with the batteries on top, to disturb the flow near the bed and around the instruments as little as possible. Even though the space in between the frames' legs was maximized, it is unknown to what extent the flow was still disturbed.

Grain sizes near the frames were determined through box core samples (Table 4.1). The sediment was found to be relatively fine, with grain sizes between 0.185 and 0.226 mm. More information about the field campaign and measurement equipment can be found in van Prooijen et al. (2020).

Table 4.1 Average water depths and grain sizes at the measurement sites, and the dates of Frame deployment and multibeam echosounding (MBES).

Frame	Depth [m]	D50 [mm]	Frame deployment [dd-mm - dd-mm in 2017]	MBES survey dates [dd-mm in 2017]
1	20.4	0.226	08-11 - 11-12	-
2	8.6	0.225	30-08 - 09-10	29-08, 18-10
3	15.1	0.216	30-08 - 10-10	29-08, 15-11
4	8.9	0.186	29-08 - 09-10	29-08, 24-10
5	6.5	0.186	29-08 - 10-10	29-08, 24-10

4.2.4 Data processing and analysis

The sonar images were processed using the bed detection method described by Ruessink et al. (2015). First, the bed was detected by a 2^{nd} order polynomial smoothing of 11 data points around the point with maximum backscatter in each profile. Subsequently, a second-order scale-controlled filter (Plant et al., 2002) was used to interpolate the detected bed points onto a regular 2×2 m grid with cells of 0.01×0.01 m. The adopted filter scale ensured that 70% or more of the variance of bed features with a length scale smaller than 0.05 m is removed. The mean distance to the Sonar head was also removed from the bed elevation values, so that ripples are visible as perturbations. Positive values are the crests, negative values are the troughs. It was found that the frames' legs cause considerable scour; holes developed with a diameter of approximately 1 m and depth of 0.1 m. To remove the scour holes the perturbation maps were detrended with a third-order surface fit.

The orientation of the sonar is known with respect to the orientation of the other sensors. Given their heading, the sonar perturbation maps were rotated to the N-E-S-W scheme. After that, the quality of each image was assessed based on a visual inspection. Images in which the bed level clearly changed during the sonar rotation, and in which the bed could not be found because of a high amount of sediment in suspension, were removed from the data set. From the remaining images, the ripple height η_m was calculated by $2\sqrt{2}\sigma$ (Krämer and Winter, 2016), with σ being the standard deviation of the image. Then, each image was rotated in a full circle with steps of 5° and for each rotation the bedform three-dimensionality (Tb) was calculated using the auto-correlation of the bed elevation data in the x and y-direction according to Núñez-González et al. (2014). Tb is a dimensionless parameter, ranging from 0 (indicating a purely 2D bedform) to 1 (indicating a purely 3D bedform). The rotation angle that resulted in the lowest value of Tb was taken to be the orientation of the ripples, although for highly three-dimensional bedforms this orientation is arbitrary. Using this orientation angle, ripple lengths (λ) were calculated by a wavelet, using the method of Grinsted et al.

(2004), which was based on Torrence and Compo (1998). To obtain the dominant ripple length, the wavelength with the maximum power was calculated for each grid row, and then the median of all these wavelengths was taken to be the 'dominant' ripple length.

Current speeds in east-west (u) and north-south (v) direction were computed from the ADV data as 30-minute average values. Then, total speed U was given by $\sqrt{u^2 + v^2}$. The depth-averaged current velocity u_c was calculated from U using a logarithmic velocity profile on the lowest two available ADVs. This was validated by Nayak et al. (2015), who found that, even though the combination of waves and currents affects the roughness layer, the log layer is not much affected. The roughness layer is approximately 2 times the roughness element height (Florens et al., 2013), which is in our case a few centimetres. Our ADVs were mounted at 0.35 and 0.65 m above the bed, so that is well outside the roughness layer. Current directions were calculated using the Pythagorean theorem on u and v . The mean wave angle was also calculated from the ADV data, by a principal component analysis of the instantaneous high-frequency ($0.05 < f < 1$ Hz) velocities per 30 minutes.

The pressure sensors were used to calculate water depth h , significant wave height H_s and spectral wave period T_{m-10} . First, the data were divided into blocks of 30 minutes and linear wave theory was used to convert pressure to sea surface elevation. h was obtained by averaging the water depths in each block. H_s was calculated as $4\sqrt{m_0}$, where m_0 is the zero-order spectral moment of the high frequencies ($0.05 < f < 1$ Hz) in the sea surface elevation. The spectral wave period T_{m-10} was computed using the ratio of the moments m_{-1}/m_0 . For consistency with the methods of numerical models, the amplitude of near-bed orbital motion u_w was calculated according to linear theory with $\pi H_s / T_{m-10} \sinh(kh)$, in which k was approximated using the method of Guo (2002) with T_{m-10} and h . Wave- and current related Shields parameters were calculated according to Kleinhans and Grasmeijer (2006):

$$\theta_x = \frac{\tau_x}{(\rho_s - \rho_w)gD_{50}}. \quad (4.1)$$

In this equation, $\rho_s = 2650 \text{ kg/m}^3$ and $\rho_w = 1025 \text{ kg/m}^3$ are the sediment and water density, respectively, g is the gravitational acceleration and D_{50} is the median grain size of the sediment. τ is the bed shear stress magnitude, with the subscript x referring to either currents (c) or waves (w),

$$\tau_c = \frac{\rho_w g u_c^2}{18 \log\left(\frac{12h}{2.5D_{50}}\right)^2} \quad (4.2)$$

and

$$\tau_w = 0.5 \rho_w f_w u_w^2. \quad (4.3)$$

In these equations, h is the water depth and u_c is the time- and depth-averaged current velocity. The friction coefficient f_w is

$$f_w = \exp(-5.977 + 5.213 \left(\frac{A_\delta}{2.5D_{50}}\right)^{-0.194}), \quad (4.4)$$

in which A_δ is the near-bed orbital excursion according to linear theory.

$$A_\delta = \frac{u_w T_{m-10}}{2\pi}. \quad (4.5)$$

4.2.5 Ripple height predictors

Since predictors are commonly used to assess ripple characteristics and bed roughness, which are in turn used to calculate sediment transport, it is important to know how well they perform. According to Van Rijn and Havinga (1995), the bed roughness of small-scale ripples in natural conditions is equal to the ripple height. Therefore, some common ripple height predictors are chosen here. The predictor of van Rijn (2007) (called VR07 hereafter) is incorporated in the Delft3D model. Its inputs are the near-bed peak orbital velocity u_w , the depth-averaged current velocity u_c and the median grain size D_{50} . Given the Ameland D_{50} , the VR07 predicted ripple height η_p is:

$$\eta_p = \begin{cases} 150 \cdot D_{50} & \text{if } \psi \leq 50 \\ (182.5 - 0.652 \cdot \psi) \cdot D_{50} & \text{if } 50 < \psi \leq 250 \\ 20 \cdot D_{50} & \text{if } \psi > 250, \end{cases} \quad (4.6)$$

in which total wave-current velocity $u_{wc} = \sqrt{u_w^2 + u_c^2}$ and mobility parameter $\psi = \frac{u_{wc}}{(s-1)gD_{50}}$, with s the ratio of sediment density divided by water density and g the gravitational acceleration.

In TELEMAC, the Wiberg and Harris (1994) ripple model is used to calculate the wave-related ripple height in the wave-current dominated regime:

$$W_p = -0.095 \left(\ln\left(\frac{D_o}{\eta_p}\right) \right)^2 + 0.442 \ln\left(\frac{D_o}{\eta_p}\right) - 2.28 \quad (4.7)$$

$$\eta_p = \alpha \cdot \begin{cases} 0.17 \cdot 0.62 D_o & \text{if orbital ripples, } D_o/D_{50} < 2000 \\ 535 D_{50} \cdot \exp(W_p) & \text{if anorbital ripples, } D_o/D_{50} > 5000 \end{cases} \quad (4.8)$$

in which near-bed orbital diameter $D_o = H_s / \sinh(kh)$. Since η_p appears on both sides of Equation 4.8, this equation has to be solved iteratively. The correction factor α represents the effect of the current, according to Tanaka and Dang (1996) (TD96):

$$\alpha = 1 + 0.81 \cdot (\tanh(0.3 \cdot S_*^{2/3}))^{2.5} \left(\frac{u_c}{u_w} \right)^{1.9}, \quad (4.9)$$

in which dimensionless grain size S_* is given by

$$S_* = D_{50} \frac{\sqrt{(s-1)gD_{50}}}{4\nu}, \quad (4.10)$$

with ν being the kinematic viscosity of the fluid ($1.36 \cdot 10^{-6} \text{ m}^2/\text{s}$).

In the MIKE21 model, an adapted version of Nielsen (1979) (Ni79) is used to calculate ripple height. The inputs are near-bed orbital velocity u_w , amplitude of wave orbital motion A , water depth h , depth-averaged current velocity u_c and median grain size D_{50} . The ripple height is given by

$$\eta_p = h_r \cdot R_f \cdot \frac{u_{fc}}{u_{fw}}. \quad (4.11)$$

Here, h_r is the ripple height in pure wave motion, given by $h_r = 21\psi_w^{-1.85}u_w$, in which ψ_w is the wave-related mobility parameter, $\psi_w = u_w^2/((s-1)gD_{50})$.

R_f is a reduction factor,

$$R_f = \begin{cases} 1 & \text{for } \frac{u_{fc}}{u_{fw}} \leq 0.1 \\ 0.5 + 0.5 \cos \left(\left(\frac{u_{fc}}{u_{fw}} - 0.1 \right) \cdot \frac{\pi}{1.9} \right) & \text{for } 0.1 < \frac{u_{fc}}{u_{fw}} < 2, \\ 0 & \text{for } \frac{u_{fc}}{u_{fw}} \geq 2 \end{cases} \quad (4.12)$$

in which u_{fc} and u_{fw} are given by

$$u_{fc} = \frac{u_c}{6 + 2.5 \ln \left(\frac{h}{k_r + 2.5 D_{50}} \right)} \quad (4.13)$$

and

$$u_{fw} = \sqrt{0.5 f_w u_w^2}, \quad (4.14)$$

with ripple roughness $k_r = (16h_r^2)/\lambda$, in which ripple wavelength $\lambda = h_r / (0.342 - 0.34(0.5f_w\psi_w)^{0.25})$. The friction factor f_w is calculated by $f_w = \exp(5.213 \left(\frac{2.5D_{50}}{u_w} \right)^{0.194} - 5.997)$.

Khelifa and Ouellet (2000) (KhOoo) was additionally chosen here, because it is similar to VRo7 in the sense that it uses a combined wave-current velocity, but also considers the angle between waves and currents (β). In this case, u_{wc} is

$$u_{wc} = \left(\frac{u_w}{\pi} \right)^2 + u_c^2 + 2 \frac{u_w}{\pi} u_c \cos \beta. \quad (4.15)$$

The ripple height in KhOoo is given by:

$$\eta_p = 0.5 \sqrt{T u_{wc}} (0.32 + 0.017 \ln(1 + \psi)^2 - 0.142 \ln(1 + \psi)). \quad (4.16)$$

The four predictors above are equilibrium height predictors, whereas Soulsby et al. (2012) (So12) incorporates the time-evolution aspect of the ripple height to account for the non-equilibrium time evolution of the ripple characteristics caused by time-varying hydrodynamics. To do so, So12 uses the same inputs as VRo7, but adds wave height H , water depth, wave period and a time step. Instead of determining the combined wave-current velocity, the predictor determines for each time step whether waves or currents are dominant, using the Shields parameters. The ripples are wave-dominated if $\theta_w \geq \max(0.42\theta_c^{0.47}, 0.08)$, with θ_w and θ_c given by Equation 4.1. In this case, as long as θ_w exceeds the critical bed shear stress θ_{cr} , the equilibrium ripple length and height are based on $\Delta = A/D_{50}$ and the rate of change is based on ψ_w . If the ripples are current-dominated, equilibrium ripple length and height are based on D_{50} and the rate of change is based on θ_c and θ_{cr} . The actual ripple height is then calculated with a Runge-Kutta integration of the considered time step. For wave-generated ripples, the characteristic timescale is equal to one wave period (in accordance with Doucette and O'Donoghue (2006)), whereas for current-generated ripples this timescale depends on η , λ and D_{50} . When the wave- or current related bed shear stress is below θ_{cr} , the ripples do not change and remain at the pre-existing value.

Even though all of the predictors mentioned above are used to predict bedform heights in various environmental conditions, most of them were created for specific settings. VRo7 was based on field and flume studies with a large range of D_{50} , and was tested for coastal

applications using field data with D_{50} of 0.300 mm. Although TD96 was developed specifically for lab conditions, TELEMAC also depends on Wiberg and Harris (1994), which was based on both field and lab conditions. Both predictors were made for grain sizes ranging from 0.1 to more than 0.65 mm. Ni79 was based on a flume study with D_{50} ranging from 0.08 to 0.36 mm. KhOoo was created based on a flume study, in which a large range of regular wave conditions were tested, with waves and currents travelling either perpendicularly or obliquely. Finally, the wave-related part of So12 was based on lab studies with $D_{50} = 0.44$ mm and regular, asymmetric waves. The current-related part was based on lab studies with $D_{50} = 0.095, 0.231, 0.237, 0.238,$ and 0.510 mm. The predictor was tested against field and lab data, which were all coarser than 0.25 mm.

4.3 Results

4.3.1 Observed hydrodynamics

To give an overview of the hydrodynamic conditions during the field campaign, Figure 4.4D-G shows the significant wave height, near-bed current speed, water depth and the angle between waves and currents at frame 4 through time. The tidal range varied from 1 m during neap tide to 2 m during spring tide. The angle between waves and currents varied on the timescale of the tides, all values between 0 and 180° were measured. During fair weather conditions significant wave heights were typically smaller than 1 m and maximum current speeds were 0.5 m/s. A storm with winds from the west occurred around September 13 (the maximum wave height was reached on September 13 at 13:30 hours), with wave heights up to 3.5 m and near-bed current speeds of up to 1 m/s in both the u (East) and v (North) direction at frame 4.

Although Frame 1 was deployed during a different period, conditions were similar to those of frames 2-5. Wave heights were often in between 1 and 2 m, with a maximum of 3.1 m and a minimum of 0.2 m. Current speeds in u and v direction were around 0.4 and 0.1 m/s, respectively, so they are a little smaller than those of frames 2-5.

Figure 4.5 shows the wave- and current-related bed shear stresses at all frames, both for the entire field campaign and for the storm and fair weather period that are indicated in Figure 4.4. Conditions are wave-dominated if $\theta_w > \theta_c$ and current-dominated if $\theta_c > \theta_w$. Wave-only and current-only conditions occur when $\theta_w > 0.2\theta_{w,cr}$ and $\theta_c > 0.2\theta_{w,cr}$, respectively (Kleinhans, 2005). Figure 4.5A illustrates that conditions were mixed during most of the campaign at most of the frames. Even frame 1, which was located offshore of the delta at 20 m depth and measured during a different period of time, shows wave-current dominated conditions. Frames 2 and 5, which were placed on the delta lobes at 7 and 6 m water depth, show the largest spread, and also contain more wave-dominated conditions. Frame 3, which was located in the inlet, is the most distinct, with current-dominated conditions only.

During the storm, both wave- and current shear stresses increased through time at frames 4 and 5, until the mobility exceeded the sheetflow criterion ($\theta_w + \theta_c > 1$ (Kleinhans, 2005)). After that, waves remained important for a longer period of time, while the current impact was already decreasing. At frame 2, the shear stresses were around the threshold of motion, but clearly induced by both waves and currents. During calm weather, frames 2, 4 and 5 were at or just below the threshold of motion until increasing current velocities (flood) caused an increase in current-related shear stress. During ebb, the current velocities decreased again,

Figure 4.4 Time series of ripple heights η (A), lengths λ (B) and three-dimensionality Tb (C) for frames 2 - 5 and significant wave heights (D), current speeds (E), water depth (F) and angle between current vector and angle of wave propagation β (G) at frame 4 during the field campaign. Tb ranges between 0 and 1, with 1 indicating fully three-dimensional bedforms and 0 meaning fully two-dimensional bedforms. A representative calm weather tidal cycle (5 September) and a storm tidal cycle (13 September) are indicated between black lines.

while the wave-related velocities remained constant, so the mobility decreased to near or below the threshold of motion again. The mobility at Frame 3 was above this threshold for 84% of the time.

Figure 4.5 A: Measured wave- versus current-related Shields stress for all frames. Black lines indicate sheet-flow and incipient motion. Red dashed lines show the boundaries between regions of wave or current dominance, according to Kleinhans (2005). B: Solid lines indicating the calm periods and dash-dotted lines indicating the storm periods that were introduced in Figure 4.4. The time series start at the coloured dots and end at the black dots.

4.3.2 Observed ripple characteristics

As can be seen in Figure 4.4A-C, the variations in ripple dimensions were relatively small in space and time. The patterns in η_m and λ through time were similar for frames 2, 4 and 5, with peaks in heights and lengths occurring simultaneously. Frame 3, which was located in the inlet, showed different patterns, although the range in η_m , λ and Tb were similar to those of the other frames. The average ripple height and length among all frames were 0.015 and 0.11 m, respectively (Figure 4.6). Despite the large ranges in θ_w , θ_c , H_s , u , v , h , and β (Figure 4.5 and 4.4d-g), 79% of all observed ripples were in between 0.01 and 0.02 m high and 83% were between 0.08 and 0.13 m long. On a timescale smaller than that of the tides, variability is found in η_m and λ , but most of the time these variations are smaller than the measurement accuracy. Around 10 September, clear peaks in height and length can be observed at Frame 2, but during this period wave heights were low and current speeds were average.

On average, the ratio η/λ was 0.15. Most of the observed ripples were around the maximum equilibrium steepness of $\eta/\lambda = 0.16$, which was found by e.g. Wiberg and Harris (1994) and Camenen (2009) (Figure 4.6). Yet, it is intriguing that this steepness was found for 2D orbital vortex ripples, but the ones we observed were neither 2D nor orbital. Given the small adaptation time scales we do expect the ripples to be in equilibrium for most of the time. All ripples were highly three-dimensional; the Tb values ranged between 0.94 and 1, with the mean and median of the entire dataset being 0.994 and 0.996, respectively (Figure 4.4C). Frames 4 and 5 show the highest Tb values (between 0.98 and 1), whereas those at frames 2 and 3 are a little lower (between 0.94 and 1). Visually, ripples appeared to be

Figure 4.6 2D histogram of ripple heights and lengths for all frames. The black line is the maximum equilibrium ripple steepness $\eta/\lambda = 0.16$. The average height and length are given by the red cross.

relatively linear sometimes, e.g. after prolonged times of low waves. An example can be found in Figure 4.7C, which is representative of most of the images at frame 3, which was in a current-dominated regime (Figure 4.5). Yet, the Tb value is still 0.984 here, because the standard deviations parallel and perpendicular to the ripple crests are both high. The ripples in Figure 4.7B have a Tb value of 0.986 and are representative for frames 1, 2, 4 and 5, which were in a wave-current dominated regime (Figure 4.5). In all cases, the ripple field is changing frequently, although this does not lead to noteworthy variability in ripple length and height. Figure 4.7 shows that ripples migrate in some cases (Figure 4.7A vs. C), whereas their shape and orientation can also change significantly within an hour (Figure 4.7C vs. E). In this second case, individual ripples cannot be tracked through time.

Because the range in ripple heights, lengths and three-dimensionality is small, it is not surprising that relations between bedform characteristics and hydrodynamics (H_s , u_w , u_c , β , ψ , etc.) are generally absent. No statistically significant relation could be found between any of these hydrodynamic parameters and Tb and λ of the observed ripples; however, η_m was found to depend on the peak near-bed orbital velocity u_w ($r^2 = 0.38$ for a 2nd order polynomial fit) (Figure 4.8). There is a lot of scatter around the best fit line, but it is evident

Figure 4.7 Examples of ripples found on the Ameland ebb-tidal delta. The ripples occurred at Frame 3, 6 Sep 2017, 03:55 (A), 04:55 (C) and 05:55 (E) and at Frame 4, 24 Sep 2017, 18:00 (B), 19:00 (D) and 20:00 (F).

that the ripple height decreases when the orbital velocity increases. No such relation was found for the mean current velocity u_c , or for the ratio θ_c/θ_w . The ripples were not affected by the changes in megaripple characteristics either.

Figure 4.8 Relations between measured ripple characteristics and hydrodynamics for all 5 frame locations. A: Ripple height η versus wave orbital velocity u_w . The line is the second order polynomial fit. B: Ripple height η versus current velocity u_c . C: Ripple height η versus ratio θ_c/θ_w . D: Ripple length λ versus ratio θ_c/θ_w .

4.3.3 Bedform height predictors

The measured ripple heights (η_m) are now compared to those predicted (η_p) by the various predictors given before (Figure 4.9). As expected, all predictors are sensitive to either wave- or current velocity, or both. Apart from Ni79, all predictors mainly overestimate the ripple height. Also, it was found that the ratio η_p/η_m for all predictors was similar for frames 2-5, while η_p/η_m was higher at frame 1. The So12 and VRo7 predictors show similar results; they overpredicted ripple height in 96% and 97% of the cases, respectively, and the ratio η_p/η_m is only sensitive to the wave orbital velocity. Especially with intermediate orbital velocities (≈ 0.4 m/s) η_p is too high. For orbital velocities smaller than 0.2 m/s or around 0.8 m/s, η_p gets closer to η_m . Above 0.8 m/s, sheetflow conditions are predicted, in which the η_p decreases more than η_m . Overall, So12 works slightly better than VRo7: 74% of the η_p are within a factor 2 (P2) of η_m , versus 60% for VRo7.

The performance of KhOoo is related to the current velocity, and overpredicts most of the values. The P2 value is 10%, which is lower than all other predictors. The best predictions are found at very high ($u_c > 0.5$ m/s) or very low ($u_c < 0.1$ m/s) current velocities. Ni79 has the largest range in predicted ripple heights, and is both under- and overpredicting measured values. P2 is 23%. For TD96, P2 is a little higher, 33%, but it always overestimates η_p when $u_c > 0.3$ m/s.

Figure 4.9 Ratio between predicted and measured ripple height, as a function of wave orbital velocity u_w (left column) and current velocity u_c (right column). Grey dots are data from frame 1, black dots are data from frames 2 - 5. These were plotted together because they were highly similar. The black line is the line of perfect agreement, and the dotted lines are ratios of 0.5 and 2. For clarity, y-axes were clipped at 0.1 and 10, therefore results outside this range are not shown. For TD96, the iterative equations could not be solved when $u_w > 0.2$ m/s.

4.4 Discussion

The ripple heights and lengths as found in the present study are small, yet they are always 3-4 times larger than the maximum footprint of the sonar. Scour occurred around the frames'

legs, but no different characteristics were seen near the legs compared to directly under the 3D Profiling Sonar. Ripple heights, lengths and three-dimensionality are relatively constant, despite the large range in hydrodynamic conditions. The hydrodynamic conditions observed here are compared to those of earlier studies in Figure 4.10. When this figure is compared to Figure 4.5A, it is visible that the present study is unique in the large range of hydrodynamic regimes that was studied; only in the present study, all types of regimes were found. Because of this large extent of data, some overlap can be found with the Shields parameters that were found in other studies as well. The study of Dolphin and Vincent (2009) is most comparable in terms of Shields numbers and grain size -their field site had a D_{50} of 0.220 mm-, but they did not report actual bedform heights. However, they did find both two- and three-dimensional ripples of which most (52%) were longer than 0.35 m. Thus, their ripples were longer and more two-dimensional than those in the present study. Yet, this difference is probably not caused by the forcing conditions, but by resolution issues: the images of Dolphin and Vincent (2009) had an accuracy of ± 0.02 m, so if small-scale ripples were present, they would not have been visible in their study.

Figure 4.10 Shields diagram showing the range of Shields numbers found in previous studies.

Wengrove et al. (2018)'s data are also comparable to our study in terms of Shields numbers, but they also found ripples that were more two-dimensional than in the present study. Also, the range in ripple heights and lengths was much larger than in the present study; Wengrove et al. (2018) found lengths of 0.14 to 2 m and heights of up to 0.3 m. Both the different dimensions and three-dimensionality might be caused by the relatively fine sand on the Ameland ebb-tidal delta: the maximum grain size near the measurement frames was 0.226 mm. The study of Wengrove et al. (2018) was done in sand with 0.35 mm grain size. O'Donoghue et al. (2006) found that ripples become two-dimensional when the grain size is

larger than 0.30 mm. They suggest that in sand with $D_{50} \leq 0.22$ mm, the ripples can only become two-dimensional when the mobility (ψ) is very low. They explain this by the sand's susceptibility to turbulence: coarse grains will mainly react to the 'main' flow, which is 2D, while finer sediment reacts to the small-scale 3D turbulence in the flow. This implies that the bedform dimensions will not decrease further when D_{50} decreases below 0.22 mm, because they are only dependent on these small-scale hydrodynamics. This also explains why the ripples found in fine sediment are smaller than those in coarse sediment, which was also observed by O'Donoghue et al. (2006).

This hypothesis, linking fine-grained sand to small-scale 3D turbulence, is further confirmed by the other studies presented in Figure 4.10. With sand having $D_{50} = 0.255$ mm, the study of Bolaños et al. (2012) is more comparable to our study. They also found a smaller range in ripple dimensions than Wengrove et al. (2018): their height was generally between 0.02 and 0.04 m, with peak values of up to 0.17 m, and their wavelength was 0.2–0.3 m, with peak values of 1.0 m. However, only 2D profile measurements were done, so there is no information about three-dimensionality. The study of Krämer and Winter (2016) was done in a region with a small grain size of 0.105 mm, but also in an environment with low mobility (35 m water depth). Ripples were found to be three-dimensional, similar to those found in this study (Figure 4.7a), with heights of 0.015 – 0.02 m and a mean length of 0.215 m. This is further evidence that the ripple heights are probably independent of location in fine-grained sediment.

In our field study, ripple heights were only found to be related to orbital velocity, whereas most predictors are dependent on current velocity as well. Just like the three-dimensionality, the weak relation with the hydrodynamics and the lack of variation in the bedform heights and lengths in the studied area might be caused by the small grain size. O'Donoghue et al. (2006) confirms that, because 3D ripples are smaller than 2D ripples, their dimensions are overpredicted by models. This is also found in the present study: most ripple heights are overpredicted by more than a factor 2 by all 5 tested predictors. So12 and VR07 perform best, which is probably because they were tested for a wide range of field and flume conditions. On the other hand, they were only tested for cases with grain sizes larger than 0.25 mm. This could explain why they overpredict ripple heights in an environment with smaller grains. KhO00 incorporates the angle between waves and currents, but this parameter was found to have no effect on the ripple dimensions in the present study. Furthermore, this predictor was created based on a flume study, which, even though a large range of conditions were tested, was probably not representative of the field. This also holds for Ni79 and TD96.

The incorporation of the time-evolution aspect of bedform height in So12 was expected to give this predictor a major advantage, but it was not a substantial improvement over VR07. Therefore, the higher P2 for So12 can only be ascribed to its formulation in general. This is because the adaptation timescales are too small to affect the equilibrium heights. Since the sonar measurements were done once per hour, the time lag effect is not visible in the results. Furthermore, most of the ripples were around the equilibrium steepness. According to Doucette and O'Donoghue (2006) and Soulsby et al. (2012), in wave-dominated situations the time scale is equal to a wave period, so 5–10 s, and in current-dominated situations it is computed from So12 to be roughly 7 minutes. These small timescales imply that they are not relevant on highly dynamic locations with only very small ripples such as ebb-tidal deltas. In short, the predictors are highly sensitive to both wave- and current-related velocities, while the measured ripples were found to depend only on wave orbital velocity. This implies that

the prediction of ripple heights in fine sediment could be improved by using predictors that are only dependent on waves. Also, because most ripple heights and lengths are within a factor 2 of the average, a constant value could be used as well, in this case $\eta = 0.015$ m and $\lambda = 0.11$ m. By doing so, 97% of predictions are within a factor 2 of the measurements (P2), which is higher than all predictors that were studied. Further research is necessary to find out whether this improves the sediment transport calculations as well.

4.5 Conclusions

Our data of the Ameland ebb-tidal delta show small wave-current ripples that are highly three-dimensional, and are on average 0.015 m high and 0.11 m long, despite strong variations in hydrodynamical conditions. The weak to non-existing dependence of ripple characteristics on the hydrodynamics might be caused by the small grain size. The five tested ripple height predictors all overpredicted the measured ripple height in most cases. This is because they were not created for small grain sizes, and are highly dependent on current velocity, which was shown here not to be related to ripple height. Therefore, the use of constant ripple height and length values is more accurate than any of the predictors here. This could potentially also improve the prediction of sediment transport in wave-current dominated environments with grain sizes smaller than 0.25 mm.

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Chapter 5 | From ripples to large-scale sand transport: the effects of bedform-related roughness on hydrodynamics and sediment transport patterns in Delft3D

Abstract

Bedform-related roughness affects both water movement and sediment transport, so it is important that it is represented correctly in numerical morphodynamic models. The main objective of the present study is to quantify for the first time the importance of ripple- and megaripple-related roughness for modelled hydrodynamics and sediment transport on the wave- and tide-dominated Ameland ebb-tidal delta in the north of the Netherlands. To do so, a sensitivity analysis was performed, in which several types of bedform-related roughness predictors were evaluated using a Delft3D model. Also, modelled ripple roughness was compared to data of ripple heights observed in a six-week field campaign on the Ameland ebb-tidal delta. The present study improves our understanding of how choices in model set-up influence model results. By comparing the results of the model scenarios, it was found that the ripple and megaripple-related roughness affect the depth-averaged current velocity, mainly over the shallow areas of the delta. The small-scale ripples are also important for the suspended load sediment transport, both indirectly through the affected flow and directly. While the current magnitude changes by 10–20% through changes in bedform roughness, the sediment transport magnitude changes by more than 100%.

Based on:

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5.1 Introduction

To contribute to answering scientific and practical questions, numerical morphodynamic models are often used to predict the hydrodynamics, sediment transport processes and morphological developments of coastal systems. In such models, many of the processes are parameterised, based on various assumptions. One of the parameterised variables is the bedform-related hydraulic roughness, which has a direct influence on the magnitude of friction between the bed and the flowing water (Lefebvre et al., 2011b). At a flat seabed, roughness is caused by the individual grains (skin friction), but in many situations bedforms are present, inducing additional drag at the bed (form drag, or bedform-related roughness) (van Rijn, 1982). The total effective hydraulic roughness consists of both the grain roughness and the bedform-related roughness. Although it is difficult to measure this roughness directly, it affects the magnitude and vertical structure of both the steady flow and the tidal and wave-driven oscillatory currents, and consequently also the magnitude of the sediment transport (van Rijn, 2007). Therefore, it is important that a parameterisation is chosen in morphodynamic models that correctly represents environmental conditions.

There are many ways to describe or parameterise bedform roughness in models, ranging from constants in space and time to predictions of the actual bedforms present. Bedform roughness predictions are often a function of bedform dimensions, either height and length (Van der A et al., 2013) or only bedform height (van Rijn, 2007), which are predicted based on wave- and/or current velocities and median grain size (O'Donoghue et al., 2006; van Rijn, 2007). In the case of a spatio-temporally constant parameters, a Chézy value (e.g., (Dastgheib et al., 2008; Lenstra et al., 2019; Iglesias et al., 2019)), Manning's n value (e.g., (Herrling and Winter, 2014)) or a constant roughness based on measured ripple heights (e.g., (Zavala-Garay et al., 2015; Ralston et al., 2019)) can be used. In numerical models, the hydrodynamic roughness is often used as a calibration parameter to minimize mismatch in observed and predicted water levels, current velocities or even morphological predictions (e.g., (Elias and Hansen, 2013)). However, in most coastal systems, bedforms occur on various scales, so bedform-related roughness can have various dimensions in both space and time. For example, in tidal inlet systems, dunes (with lengths in the order of 100 m and heights in the order of meters), megaripples (lengths in the order of 10 m and heights in the order of 0.1 m) and small-scale ripples (with lengths in the order of 0.1 m and heights in the order of centimeters) can be found. Especially the small-scale ripples and megaripples respond quickly to the local hydrodynamics (Wengrove et al., 2018; Lefebvre et al., 2011a), which implies that ripple and megaripple-related roughness might vary on the time scale of minutes to hours and the spatial scale of meters. To accurately represent this behaviour, roughness predictors were created, so that the parameterised roughness can also vary in space and time (e.g., (van Rijn, 2007; Ganju and Sherwood, 2010)). Most predictors are well-tested on flume data and on field data of wave-dominated coasts (van Rijn, 2007; Nelson et al., 2013). For environments with both waves and currents, such as ebb-tidal deltas, less field data are available. Yet, roughness may be different here from wave-only or current-only conditions, especially because wave and current stresses can work in different directions. Therefore, when bedform-related roughness predictors are used, which naturally contain uncertainty, uncertainty is imposed on the current and sediment transport predictions as well. However, the magnitude of this uncertainty is still unknown.

This chapter builds on Chapters 3 and 4, where ripple and megaripple characteristics were studied on the wave-current dominated Ameland ebb-tidal delta in the north of the

Netherlands, which were also compared to some common ripple roughness height predictors. While multiple tested models (including van Rijn (2007)) predicted small-scale ripple roughness heights to vary significantly on the timescale of the tides, in Chapter 4 it was found that observed ripple characteristics varied much less in time, and only slightly depended on wave orbital velocity. Even though a large range of hydrodynamic conditions was encountered, the ripple height was nearly always between 0.01 and 0.03 m and ripple length was nearly always between 0.08 and 0.13 m. For the small grain sizes (i.e., $D_{50} = 211 \mu\text{m}$) at the study site, in Chapter 4 ripple dimensions were hypothesized to be in response to small-scale three-dimensional turbulence in the water column and not to the ambient flow. Moreover, in Chapter 4 it was found that ripple height was overestimated by the predictor developed by van Rijn (2007), which can be used in various morphodynamic models and is the default parametrization in the often applied Delft3D model. It was suggested that a constant ripple-related roughness might be used in modelling of hydrodynamics and sediment transport, but the actual effects were not tested.

The behaviour of megaripples through time was studied in Chapter 3. It was found that they were present all across the inlet channels, and co-occurred with the small-scale ripples. Not only are megaripple heights and lengths larger than those of the small-scale ripples, megaripple-related roughness is larger as well (Idier et al., 2004). This implies that their effect on the current and sediment transport will be larger as well. Davies and Robins (2017) found that for small current velocities ($<0.2 \text{ m/s}$) the predictor developed by van Rijn (2007) predicts, however, that the contribution of small-scale ripples to the total roughness exceeds that of the megaripples. Similar to the small-scale ripples, the predicted megaripple roughness height was also smaller than the measured roughness height. In the case of megaripples, measured and predicted roughness height differ by a factor 2.

In short, the use of bedform predictors to calculate roughness might affect both predicted hydrodynamics and sediment transport, but this impact has not been quantified yet. The main aim of this paper is to examine and quantify the importance of the ripple and megaripple-related bedform roughness for the calculation of hydrodynamics and sediment transport in the numerical morphodynamic model Delft3D. This model is used, because it is widely used and comparable to other morphodynamic models, such as MIKE21, TELEMAC, and ROMS. Although the way that the roughness predictors are incorporated is slightly different for each model, they can all make use of the van Rijn (2007) roughness predictor. First, it will be explained how roughness, hydrodynamics and sediment transport are computed and interact in Delft3D. Then, several predictors of bedform-related roughness are applied in a high-resolution Delft3D model of the Ameland Inlet. The modelled bedform-related roughness will be compared to the one parameterised from measured ripple characteristics, and the hydrodynamics and sediment transport patterns of the various scenarios are compared. This comparison is done for fair weather and storm conditions separately, because it is anticipated that roughness is different for the associated different wave and current conditions. The Ameland ebb-tidal delta is chosen as study site, because it is a wave- and tide-dominated coastal system. This hypothetically causes a large spatial variability in roughness, which makes it a suitable location to test various roughness parameterisations. Furthermore, many measurements, including those of megaripples (Chapter 3) and small-scale ripples (Chapter 4), are available as a basis for a sensitivity analysis. Also, distinct calm weather and storm periods were observed.

5.2 Materials and Methods

5.2.1 Interaction of Roughness, Hydrodynamics and Sediment Transport in Delft3D

Delft3D is a numerical morphodynamic model system that incorporates all parts of the morphodynamic feedback loop: hydrodynamics, sediment transport and morphologic change (Lesser et al., 2004). Both in Delft3D and similar models, roughness elements such as grains, ripples, and megaripples are smaller than the model grid (i.e., sub-grid), so they are not explicitly solved on that scale and hence need to be parameterised. When a single grain size fraction is used, grain-related roughness k_g does not vary in space or time, and is given by $k_g = D_{90} = 1.5D_{50}$, with D_{50} and D_{90} the median and 90th percentile of the grain size, respectively. The bedform-related roughness, on the other hand, interacts with the hydrodynamics and affects the sediment transport both directly and indirectly. A schematic overview of all processes in Delft3D that are influenced by the ripple-related roughness is given in Figure 5.1. There are two ways to impose roughness in Delft3D: either via a direct calculation of each roughness scale (ripples, megaripples and dunes), or via a prescribed roughness value or friction coefficient.

In the direct calculation of bedform roughness k_s , the separate parts $k_{s,r}$ (ripple-related roughness), $k_{s,mr}$ (megaripple-related roughness) and $k_{s,d}$ (dune-related roughness) are calculated from van Rijn (2007). The predicted ripple roughness $k_{s,r}$ is:

$$k_{s,r} = \alpha_r \begin{cases} 150D_{50} & \text{if } \psi \leq 50 \\ (182.5 - 0.652\psi)D_{50} & \text{if } 50 < \psi \leq 250 \\ 20D_{50} & \text{if } \psi > 250, \end{cases} \quad (5.1)$$

in which $\psi = u_{wc}^2 / [(s - 1)gD_{50}]$ is a mobility parameter, with s the ratio of sediment density divided by water density and g the gravitational acceleration. The total combined wave-driven and tide-induced velocity $u_{wc} = \sqrt{u_w^2 + u_c^2}$, with u_w the magnitude of the wave-induced orbital motion and u_c the magnitude of the depth-averaged current velocity. α_r is a ripple roughness calibration factor. As given in Equation (5.1), the ripple-related roughness is related to grain size and wave- and current velocity. Both ripple and megaripple-related roughness are explicitly calculated, so it depends on the flow conditions of the previous time step. Since bedforms do not always immediately react to the hydrodynamics, a relaxation option is included as well, which is calculated implicitly:

$$k_{s,r}(t) = \beta k_{s,r}(t - dt) + (1 - \beta)k_{s,r}(t), \quad (5.2)$$

with β being the relaxation factor: $\beta = e^{-dt/T}$, in which dt is the hydrodynamic time step in minutes and T the relaxation time scale in minutes.

The megaripple-related roughness is given by:

$$k_{s,mr} = \alpha_{mr} \begin{cases} 0.0002\psi h & \text{if } \psi \leq 50 \\ (0.011 - 0.00002\psi)h & \text{if } 50 < \psi \leq 550 \\ 0.02 & \text{if } \psi > 550, \end{cases} \quad (5.3)$$

with h the water depth. A relaxation option for megaripples is available in the model as well, but it was not used in the present study. Since dunes are only applicable in rivers (van Rijn et al., 2004), the prediction of dunes was turned off in the model and the formulas are not given here.

When all separate roughness scales are computed, the total bedform roughness k_s is calculated as

$$k_s = \min(\sqrt{k_{s,r}^2 + k_{s,mr}^2}, \frac{h}{2}). \quad (5.4)$$

Although this squared summation is not a physically exact representation of the actual roughness, van Rijn (2007) states that this is the most practical approach for engineering purposes. Using k_s , the Chézy friction coefficient C is calculated as

$$C = 18 \log_{10} \frac{12h}{k_s}. \quad (5.5)$$

As stated before, the Chézy value can also be given as input directly. The third possibility is calculating the Chézy value by

$$C = \sqrt[6]{h}/n, \quad (5.6)$$

with n being Manning's roughness coefficient. According to Marriott and Jayaratne (2010), Manning's n can be calculated by

$$n = \frac{1}{26} k_s^{1/6}, \quad (5.7)$$

In each of the cases with a spatio-temporally constant Chézy value, the bedforms will not affect the flow anymore, but $k_{s,r}$ is still calculated and affects the sediment transport (see Equation (5.8)). In each case, the Chézy coefficient is an input for the bed shear stress calculation, which in turn affects the flow.

To solve wave propagation the phase-averaged spectral wave model SWAN was used. SWAN solves the wave action balance equation in order to compute the wave energy spectrum throughout the model grid (Booij et al., 1999; Ris et al., 1999). Delft3D and SWAN are coupled, which means that the effects of waves on hydrodynamics and vice versa are included, so waves can feel and interact with the flow, by which they can affect $k_{s,r}$ and $k_{s,mr}$. $k_{s,r}$ and $k_{s,mr}$ can only affect the waves indirectly through the affected flow (5.1). Although the roughness is not directly coupled to SWAN, it does use a semi-empirical energy decay rate dependent on bed roughness typical for sandy beds and JONSWAP waves of $0.038 \text{ m}^2 \text{ s}^{-3}$ (Nederhoff et al., 2019a).

The suspended load is computed by solving the depth-averaged (2DH) advection-diffusion equation (Galappatti and Vreugdenhil, 1985). The depth-averaged equilibrium concentration, which is used to calculate the exchange of sediment between the bed and the water column, is solved using the expressions of Van Rijn (2007). This approach includes the computation of the velocity profile including wave-current interaction effects described by Van Rijn (1993) and a vertical concentration profile using a 1DV advection-diffusion equation. The expression for vertical sediment mixing contains contributions due to waves and currents, both being affected by the ripple-related roughness. To solve for the vertical sediment concentration profile, the near-bed reference concentration c_a is needed as well, which is calculated by

$$c_a = 0.015 \left(\frac{D_{50}}{a} \right) \frac{\left(\frac{\tau'_{b,cw} - \tau_{b,cr}}{\tau_{b,cr}} \right)^{1.5}}{D_*^{0.3}}, \quad (5.8)$$

in which $\tau_{b,cr}$ is the critical bed shear stress and $\vec{\tau}'_{b,cw}$ is the grain-related bed-shear stress due to both currents and waves, computed using the grain-related roughness k_g . D_* is the non-dimensional grain size and reference height $a = 0.5k_{s,r}$. Below a , the sediment concentration is equal to c_a (Lesser, 2009).

The depth-integrated suspended load transport is computed by multiplying the depth-averaged velocity $\vec{U}$ with the depth-averaged sediment concentration c and water depth h :

$$\vec{q}_s = \vec{U}ch. \quad (5.9)$$

The instantaneous bedload transport is computed by Van Rijn (2007):

$$\vec{q}_b = 0.5\rho_s D_{50} D_*^{-0.3} \left(\frac{\vec{\tau}'_{b,cw}}{\rho_w} \right)^{0.5} \left(\frac{\vec{\tau}'_{b,cw} - \tau_{b,cr}}{\tau_{b,cr}} \right), \quad (5.10)$$

in which ρ_s and ρ_w are the density of sediment and water, respectively. For further details on the sediment transport formulae, the reader is referred to Van Rijn (2007).

In short, the ripple roughness determines the reference height a and the vertical mixing of suspended sediment and therefore directly affects the suspended load transport. The mega ripple roughness does not directly affect the sediment transport; however, together with the ripple roughness it is a part of the total roughness k_s , and it can affect the vertical mixing and thereby the sediment concentration in the water. This means that in the model, both small-scale ripples and megaripples indirectly affect the transport through their effect on bed shear stress, the Chézy friction coefficient, vertical mixing and the flow (Figure 5.1). The grain-related roughness affects the bedload transport and the near-bed reference concentration c_a .

Figure 5.1 Schematic overview of interaction and feedbacks related to ripples and megaripples in Delft3D. The numbers refer to the scenarios explained in Section 5.2.2. Yellow arrow: this effect is removed by choosing a constant Chézy value C (scenarios 4–5, note that sediment transport in this case still depends on $k_{s,r}$). Red arrow: this effect is removed by choosing a fixed ripple roughness (scenarios 7–9).

5.2.2 Simulations

The study site is the Ameland Inlet, which is located in the north of the Netherlands (in north-western Europe) and is part of the barrier island chain called the Wadden Islands that borders

the coasts of the Netherlands, Germany and Denmark. The same version of the hydrodynamically validated Delft3D Coastal Genesis II Terschelling-Ameland inlet (CG2TA) model was used as in Chapter 3, Section 3.2.2. The CG2TA model was run for the six weeks covering the period of the field campaign of 30 August–8 October 2017. Nine roughness scenarios were run, in order to assess the sensitivity of the model to both ripple and megaripple roughness. In the reference scenario, the bedform predictor by van Rijn (2007) was used, which directly predicts the ripple and megaripple roughness. The model was calibrated to match hydrodynamic observations best using ripple and megaripple-related roughness calibration factors of 0.5 (Nederhoff et al., 2019a) (α_r and α_{mr} in Equations (5.1) and (5.3), respectively), which is why they were set to 0.5 in the present study as well. The relaxation time was set to 0, so the ripples and megaripples respond to the hydrodynamics immediately. Similar to the present study, no morphologic updates were used in calibration and validation of the model.

Given the structure of Delft3D (Figure 5.1), we will focus on the effects of small-scale ripples on hydrodynamics and sediment transport. Megaripples are incorporated differently in the model, which makes it difficult to study their effect on sediment transport in a similar way as the small-scale ripples. The scenarios are described below and are summarized in Table 5.1. Model output was stored at frame locations every ten minutes, and every 60 min at all other grid points. In all cases, output was given of ripple roughness (same as Equation (5.1)), depth-averaged current velocity in east and north component, wave height and suspended, bed load and total sediment transport magnitude in east and north component on all grid points. Following Nederhoff et al. (2019b), a single grain size fraction with $D_{50} = 200 \mu\text{m}$ was used in the model. This is similar to the measured D_{50} near Ameland of $211 \mu\text{m}$, which had a standard deviation of only $30 \mu\text{m}$ (Pearson et al., 2019). No mud fractions were used, as the mud content on the delta was less than 1% by volume (Pearson et al., 2019). The roughness scenarios are:

1. The Van Rijn (2007) roughness predictor with α_r and $\alpha_{mr} = 0.5$
This is the calibrated model which was referred to in Section 3.2.2.
2. The Van Rijn (2007) roughness predictor with $\alpha_r = 1$ and $\alpha_{mr} = 0.5$
van Rijn (2007) stated that the small-scale ripple-related roughness is equal to the actual ripple height, and this assumption is tested in this scenario. This is the same as scenario 1, but with the predicted ripple roughness twice as high as in scenario 1. This value was also used in Chapter 4 to compare observed and predicted small-scale ripples, although in that study the predictions were based on measured hydrodynamics.
3. The Van Rijn (2007) roughness predictor with $\alpha_r = 0.5$ and $\alpha_{mr} = 1$
This is the same as scenario 1, but with the predicted megaripple roughness twice as high as in scenario 1. In this way, the importance of megaripples for currents and sediment transport can be studied.
4. Fixed Chézy coefficient for hydrodynamics
In this scenario, a constant Chézy value of 57.2 was adopted. This was calculated manually with Equation (5.5), using the mean water depth of the model excluding the basin, which was 12.7 m. Measured mega ripples at the Ameland ebb-tidal delta were 0.1 m high and ripples were 0.015 m high, so following Equation (5.4), the mean (measured) k_s was 0.1011 m. As stated before, in this scenario the Chézy value was used

in the prediction of hydrodynamics, but for sediment transport the Van Rijn (2007) roughness with α_r and $\alpha_{mr} = 0.5$ was still used. In other words, only the coupling between bedforms and the hydrodynamics was removed.

5. Fixed Manning's roughness

In this scenario, using $k_s = 0.1011$ m, Manning's n was determined to be $0.0263 \text{ m}^{1/6}$ (Equation (5.7)). This is within the range of 0.014 to 0.028 as was used by Nederhoff et al. (2019a). Since Equation (5.6) depends on water depth h , the Chézy constant in this scenario varied over space and time. Similar to scenario 4, the Van Rijn roughness predictor with α_r and $\alpha_{mr} = 0.5$ was still used for sediment transport but not for hydrodynamics.

6. Relaxation 180 min

Scenario 6 is similar to scenario 1, only the small-scale ripples were given a relaxation time in adapting to the hydrodynamics. This relaxation time T was set to 180 min based on an analysis of the temporal behaviour of the ripples in relation to the hydrodynamics. The measured ripples were very small, so it would be expected that they would respond instantly to the hydrodynamics. However, the measured ripple heights showed small variations only on the time scale of the tide. Therefore, the added relaxation time will probably cause a better resemblance of the measured ripples.

7. No ripples; constant ripple roughness 0.0002 m ($=D_{50}$)

For scenarios 7–9, a constant and spatially uniform ripple roughness was defined. As visible in Figure 5.1, $k_{s,r}$ still affected hydrodynamics and sediment transport in these scenarios, the only difference was that the hydrodynamics could not affect $k_{s,r}$ anymore. In scenario 7, $k_{s,r}$ was set to 0.0002 m . This is equal to the median grain size, to simulate the effect of a sandy bed without small-scale ripples. $k_{s,mr}$ was still the same as in scenario 1, because it does not directly affect the sediment transport.

8. Constant ripple roughness 0.015 m

In scenario 8, the ripple roughness was 0.015 m , which is similar to the measured ripple heights (Chapter 4). $k_{s,mr}$ was still the same as in scenario 1.

9. Constant ripple roughness 0.03 m

In scenario 9, ripple roughness was set to 0.03 m to see the effect of a doubling of the constant roughness, similar to scenario 2. $k_{s,mr}$ was still the same as in scenario 1.

5.2.3 Analysis

First, predicted ripple roughness was compared to measurements at the frame locations, to verify that the modelled ripple roughness is realistic in all scenarios. (Measured and modelled megaripples were compared in Chapter 3, Figure 3.16.) As described in Chapter 4, the ripples were measured by a 3D Profiling Sonar (Marine Electronics type 2001) on each frame. Ripple heights were calculated with $2\sqrt{2}\sigma$, where σ is the standard deviation of the detrended bed elevation. Ripple wavelengths were calculated using a wavelet analysis. The ripple dimensions were found to be highly similar at the four frame locations: all ripples were between 0.01 and 0.03 m high, with an average of 0.015 m , and all lengths ranged between 0.08 and 0.20 m , with an average of 0.115 m . Despite a large variation in tidal, wind and

Table 5.1 Overview of model scenarios, with $k_{s,r}$ the ripple roughness height and α_r and α_{mr}

the ripple and megaripple calibration factors, respectively, with which the predicted ripple and megaripple roughness heights are multiplied.

Number	Short Name	$k_{s,r}$ Predicted By	α_r [-]	α_{mr} [-]	Does $k_{s,r}$ Affect Hydrodynamics?	Do Hydrodynamics Affect $k_{s,r}$?	Relaxation [min]
1	base	Van Rijn 2007	0.5	0.5	Y	Y	0
2	highripples	Van Rijn 2007	1	0.5	Y	Y	0
3	megaripples	Van Rijn 2007	0.5	1	Y	Y	0
4	chezzy	Van Rijn 2007	0.5	0.5	N	Y	0
5	Manning	Van Rijn 2007	0.5	0.5	N	Y	0
6	relax180	Van Rijn 2007	0.5	0.5	Y	Y	180
7	no ripples	0.0002 m(= D_{50})	1	0.5	Y	N	0
8	constant	0.015 m	1	0.5	Y	N	0
9	constant high	0.03 m	1	0.5	Y	N	0

wave conditions, the ripple characteristics did not vary notably over time. A more detailed description of the analysis of the ripple characteristics is given in Chapter 4.

After this, the model results of scenarios 2–9 were compared to the default scenario in space and time. All model processes were considered, following the structure of Figure 5.1 (from roughness and Chézy factors to hydrodynamics and sediment transport). For each of these parameters, the mean absolute relative difference (MARD) on each location in the model was calculated. The MARD is defined as

$$MARD(x, y) = \frac{1}{n} \sum_{i=1}^n \Delta X_{rel,i}(x, y) \times 100\%, \quad (5.11)$$

with $\Delta X_{rel,i} = \left| \frac{X_i - X_1}{X_1} \right|$, in which X is a parameter from the scenarios 2 to 9 and X_1 belongs to the default scenario. x and y are all grid locations in the model domain and n denotes several output time steps. For vectors such as current velocity and sediment transport, the MARD is calculated for the magnitude. Direction differences are calculated in absolute degrees. The sediment transport is calculated in a time-integrated way, so that $n = 1$ and values are obtained in kg/m.

Two time periods were considered in this study: (1) a 36-h period of calm weather and (2) a 36-h period of storm conditions (Figure 5.2). These periods were chosen in such a way that water levels at the beginning and end of each period were the same, which is one of the ways to define a tidal period or multiples thereof (Duran-Matute and Gerkema, 2015). A storm with such high waves coming from the west is relatively rare in this area (see wave rose in Figure 4.1), but studying the measured storm conditions allows for comparison of the modelled roughness with measured ripple heights. On top of that, various environments were studied, varying from the current-dominated tidal channel to the wave-current dominated ebb-delta shoals. This facilitated in determining the sensitivity of hydrodynamics and sediment transport to ripple-related bedform roughness in various conditions. Because the model was not validated for sediment transport, we focused only on relative changes and absolute transport values are not discussed. The sensitivity in the tidal basins was not considered either, because neither hydrodynamics nor sediment transport were validated here, and the potentially high mud concentrations could result in different processes than considered by the present model.

Figure 5.2 Hydrodynamics as measured at frame 5. (A): significant wave height. (B): east (u_{cx}) and north (u_{cy}) component of the depth-averaged current velocity. (C): water depth. Black lines indicate the representative storm and calm period described throughout this paper.

5.3 Results

5.3.1 Bedform Roughness

Ripple roughness predicted in scenarios 1-6 is compared to the heights measured in the field campaign in both space and time (Figure 5.3). During the storm, ripple measurements at the frames were largely absent, because the sonar could not detect the bed due to the large amount of suspended sediment near the bed. Figure 5.3 shows that all model scenarios predict that ripples are washed out during the storm: ripple roughness decreases to 2 mm, which is equal to the $0.5 \cdot 20D_{50}$ (Equation (5.1)). Only in the channel (at frame 3) ripples were not washed out in either the measurements or the predictions. The measured and modelled ripple roughness at this location were similar, but the model values varied more with time.

In calm weather, the measured ripple heights were similar at all frames (Figure 5.3). Still, the variability in time was highest in the main channel (frame 3). This is the only location

where the measured ripple heights showed a clear tidal signal. Modelled ripple roughness values were all highly variable, showing a tidal signal (Figure 5.3). However, this modelled tidal signal was much stronger than measured and the timing was different. No dependency on the wave-related orbital velocity can be observed in either the modelled or the measured ripple roughness (Figure 5.3), probably because the current is more dominant in the mobility parameter ψ (Equation (5.1)). Only the ripples in scenario 6 (180-min relaxation time) look similar to the measured values in terms of semi-diurnal variability.

Overall, all modelled $k_{s,r}$ values were in the range of 0.002 to 0.03 m, which is a realistic range for these small-scale ripples. On average, modelled $k_{s,r}$ was somewhat lower than in the measurements in all scenarios except the scenario with the high ripples (scenario 2).

Figure 5.3 also illustrate that modelled ripple roughness was virtually the same in scenarios 1, 3, 4 and 5. This is because in all these scenarios α_r is the same, $k_{s,r}$ is affected by the hydrodynamics and no relaxation is included. Especially during calm weather, $k_{s,r}$ in scenarios 1–5 varied between two values. Scenario 2 (high ripples) peaked at a different ripple roughness and had a larger range of roughness values, but the range itself was also constant among the frames. The minimum and maximum values in these simulations are caused by the low and high cut-off value for ripple roughness in the Van Rijn predictor (at 20 and 150 times the median grain size, respectively, see Equation (5.1)). The simulation with the 180-min relaxation time is the only one that did not alternate between two cut-off values. This describes the variability of ripple roughness best, for this situation, as the measured ripple heights were not alternating between two values either, but rather changed more gradually with time. Yet, the difference between measured and modelled ripple roughness, given by the rmse and bias in Table 5.2, is smallest in the scenario with constant $k_{s,r}$ of 0.015 m. The bias and rmse are largest in the scenarios with constant high ripples and without ripples. All other scenarios are similar to the base scenario in terms of rmse and bias.

Table 5.2 Root-mean-square error and bias between the modelled and measured ripple roughness (meas-mod), averaged over all frames.

Number	Short Name	Rmse [m]	Bias [m]
1	base	0.0057	0.0034
2	highripples	0.0115	−0.0092
3	megaripples	0.0054	0.0029
4	chezy	0.0055	0.0032
5	Manning	0.0051	0.0026
6	relax180	0.0052	0.0034
7	no ripples	0.0161	0.0156
8	constant	0.0041	0.0008
9	constant high	0.0148	−0.0142

The total roughness for all scenarios at all frames is given in Figure 5.4. The patterns are similar to those of the small-scale ripples in Figure 5.3. Again, the ripples and megaripples were washed out in most scenarios for large parts of the storm. Only in the channel (at frame 3) roughness was not affected by the storm; the patterns at this frame were similar during calm weather and storm conditions. During calm weather, all scenarios at all frames show that the total roughness depends on the tidal signal, with the largest k_s during the peaks of u_c . Similar to the small-scale ripples, no relation with orbital motion or water depth was found.

Figure 5.3 Top four rows: measured and modelled (scenarios 1–6) ripple roughness $k_{s,r}$ through time at all frame locations (F1, F3, F4 and F5) during the storm (left) and the calm (right) period as denoted in Figure 5.2. Bottom row: near-bed orbital velocity u_w and depth-averaged current velocity u_c at frame 1 during the storm (left) and the calm (right) period. (As the ripple roughness was constant in scenarios 7–9, they are not plotted in this figure.)

The scenarios were again highly similar, with only two scenarios standing out. First, in the scenario with the doubled megaripple roughness (scenario 3), k_s was twice as high as in all others, at all frames. This suggests that the megaripples are more important to k_s than the small-scale ripples. However, k_s of the scenario with constant high ripples (scenario 9) exceeded even the k_s in the scenario with high megaripples during the storm at frames 1, 4 and

5. This is caused by the fact that the ripples in scenario 9 were not washed out, because they were fixed at 0.03 m. In all other scenarios, k_s ranged between 0.01 and 0.07 m. Because the small-scale ripples were generally in between 0 and 0.015 m high, a first estimation would be that small-scale ripples constitute up to 20% of the total roughness.

Figure 5.4 Top four rows: modelled total bedform roughness k_s through time at all frame locations (F1, F3, F4 and F5) during the storm (left) and the calm (right) period as denoted in Figure 5.2. Bottom row: modelled near-bed orbital velocity and depth-averaged current velocity magnitude at frame 1 during the storm (left) and the calm (right) period.

However, many more ripple roughness values and percentages occur throughout the model domain. For example, Figure 5.5A,B shows that the ripples disappeared during the storm at the margin of the ebb-tidal delta, where frames 1, 4 and 5 are located. On other

locations, for example offshore and on the sandbar that is attached to the Ameland coast, the ripples did not disappear at all. During calm weather, on the other hand, ripple roughness values at the frames were higher than in the deepest part of the channel (west of frame 3).

Figure 5.5 Modelled ripple roughness (A,B), total roughness (C,D) and ripple roughness as a percentage of total roughness (E,F) averaged over time during the storm (left) and the calm (right) period as denoted in Figure 5.2 in scenario 1.

Both during calm weather and during the storm, the total bedform-related roughness averaged over time in the default scenario varied from 0.002 m on the shallow shoals to 0.15 m offshore (Figure 5.5C,D). Because $k_{s,r}$ cannot exceed 0.015 m, all values higher than that can be attributed to megaripples. As given in Equations (5.1) and (5.3) and Figures 5.3 and 5.4 both ripple and megaripple roughness depend on the flow velocity u_c , which explains the different $k_{s,tot}$ values during storm and calm weather in the default scenario.

The importance of ripples is given by the average over time of ripple roughness as a percentage of total roughness (Figure 5.5E,F). For both time periods, the lowest importance of ripples (<10%) was found in the deep tidal inlet channels. Offshore, the ripples made up for

20–40% of the total roughness. The importance of ripples increases with decreasing depths and especially during calm weather: on the shoals ripples constituted to 60–90% of the total roughness.

During the storm, the ripples were much less important to the total roughness when averaged over the entire domain excluding the basin; while the spatially average percentage was 40% for the calm period in the base scenario, it was 22% for the storm (Table 5.3). Because the total modelling period consisted of more calm periods than storms, the percentage of the total period is close to that of the calm period: 35% in the case of the default scenario. The same is found for all other scenarios. Especially the percentages of scenarios 4, 5 and 6 are similar to those of the default scenario. Naturally, the importance of ripples in scenario 7 (with ripple roughness of only 0.0002 m) approaches 0. In scenarios 8 and 9, the ripple roughness was more important than in the default scenario, with 44 and 63% in the calm period and 40 and 59% during the storm. This means that the importance of ripples increased much more during the storm when a constant ripple roughness was used. This is because the ripples are washed away in the base scenario during storm, while they are forced to exist in scenarios 8 and 9.

Table 5.3 Ripple-related roughness as a percentage of the total bedform-related roughness, averaged over the model domain excluding the basins, and averaged over time for the calm weather, storm and total modelling period.

Scenario	Calm	Storm	Total
1	39.8	22.1	35.1
2	56.8	37.9	52.2
3	25.8	12.4	21.9
4	40.8	22.3	35.8
5	41.5	22.6	36.2
6	40.3	23.6	35.6
7	1.7	1.3	1.7
8	44.9	40.3	43.8
9	62.9	58.8	62.2

Based on these model observations, it can be expected that changes in the small-scale ripple roughness will mainly affect the shoals during calm weather. Changing the megaripple predictor settings will affect the offshore area and the channels, mainly during storms. For example, k_s increased with 80% on the shoals during calm weather when $k_{s,r}$ was doubled, while a much smaller change (less than 10%) was found in the same area during the storm.

The Chézy coefficient is a function of $k_{s,tot}$ and water depth h in all scenarios except 4 and 5. As h varies more than k_s in the model domain (but does not differ between the simulations) and Equation (5.5) includes a logarithm, differences between the scenarios are dampened out on the shallow ebb-tidal delta. Therefore, differences between modelled Chézy values are relatively small between the different scenarios. The largest differences in Chézy value with respect to the reference scenario are found in scenario 5 (with a constant Manning's n) on the shallow shoals (up to 30% during both calm weather and the storm).

5.3.2 Hydrodynamics

Among various hydrodynamic parameters examined, the current magnitude u_c is the most affected by the choice of the roughness parameterisation. Effects on water level and wave

height are minimal, because the SWAN wave model does not directly respond to the ripple-related roughness, and are therefore not further discussed. During the storm, the flow was strong across the entire ebb-tidal delta, while it was weaker and more concentrated in the main channel during calm conditions (Figure 5.6).

Figure 5.6 Time-averaged value of depth-averaged current velocity magnitude (A,B), time-integrated bed-load transport (C,D) and time-integrated suspended load transport (E,F) during the storm (A,C,E) and the calm weather period (B,D,F) on a select number of model grid points.

In general, a higher roughness leads to a lower current velocity. When the roughness was changed, the flow direction generally remained the same. Throughout the model domain, direction differences between all scenarios and the base scenario were mostly below 5° , with maximum values of 30° on the western shoal in some scenarios. Thus, directions are not affected in such a way that the current is reversed. Only the shape of the tidal ellipse is slightly altered by (temporal) changes in bedform roughness. The largest differences with the base scenario in both current velocity direction and magnitude are found when a constant Chézy or Manning roughness value was prescribed, or when the megaripple roughness was doubled. During calm weather, in these cases relative differences in current velocity exceeded 100%, but this only occurred at locations where the current velocity in the base run was near 0 (Figure 5.7). On the delta, the current was affected by approximately 50%. During the storm, choosing a constant Manning value resulted in up to 70% relative difference in current velocity on the delta, while the offshore areas were affected by approximately 20% (Figure 5.8). Choosing a constant Chézy value resulted in 30% relative difference in the entire model domain, with maximum values near the western shoal of 70%. This is similar to the ef-

fects of prescribing a doubled megaripple roughness. The use of a constant ripple roughness (scenarios 7–9) mainly affected the current at the shoals. The relative differences with the base scenario were larger for larger ripple roughness, and were 40% at the most. A doubled varying ripple roughness (scenario 2) resulted in a similar pattern, but the changes the current velocity in this case were maximum 10%. Since the effect of doubling the megaripples was much larger, it can be stated that small-scale ripples roughness is not really important for the current. Scenario 6, with a 180-min relaxation time, has even less effect; the maximum relative differences with the base scenario are 5%, but mostly they are below 1%.

Figure 5.7 Mean depth-averaged current velocity u_c (m/s) of the base scenario during calm weather (1), and mean absolute relative difference in u_c of all other scenarios with the base scenario during calm weather (2–9).

5.3.3 Sediment Transport

The bed load and suspended load transport are both considered by time-integration over the calm and storm period. The amount of suspended load transport was approximately ten to twenty times larger than the bed load transport, but apart from that the patterns in the base scenario were similar (Figure 5.6). Most transport takes place during the storm, at the seaward extent of the ebb-tidal delta shoals and at the inlet channel. During calm weather, less transport takes place, and it is mostly confined to the channel.

Similar to the current velocities, transport directions are not significantly affected by the roughness scenario. Direction differences are a little bit larger in sediment transport than in current velocity, with the largest differences occurring on the shallow parts of the delta.

Figure 5.8 Mean depth-averaged current velocity u_c (m/s) of the base scenario during the storm (1), and mean absolute relative difference in u_c of all other scenarios with the base scenario during the storm (2–9).

In the deeper parts of the model domain, differences with the base scenario are in the order of 10 degrees. Especially for suspended load transport, differences with the base scenario are found of up to 180 degrees, which means that the sediment transport direction is reversed. This happens only in areas in which the directions in the base run also show large variations on small areas. The large differences with the base run are thus caused by the transport patterns that remains the same, but slightly shifts in space.

The patterns in relative difference in time-integrated sediment transport magnitude between the base scenario and the other scenarios are also similar to those of u_c . During calm weather, for both bedload and suspended load transport, the relative differences with the base scenario were generally below 10%. Yet, values of up to 200% were found at the shallow areas, where the time-integrated amount of sediment transport in the base run approached 0 (Figures 5.9 and 5.10).

During storm, the MARD-patterns of the bedload and the suspended load magnitude are not similar. The time-integrated bedload transport is affected most when a constant Chézy value (scenario 4), Manning coefficient (scenario 5) or a constant high ripple roughness (scenario 9) is prescribed (Figure 5.11). In these cases, relative differences with the base scenario of more than 100% are found, but they occur only on the locations where the transport magnitude in the base run was very small. In the scenarios with high megaripple roughness (scenario 2), a constant Chézy value (scenario 4), very small ripple roughness (scenario 7)

Figure 5.9 Time-integrated suspended load transport q_s during calm weather (1), and absolute relative difference in q_s of all other scenarios with the base scenario during calm weather (2–9).

and constant high ripple roughness (scenario 9), the time-integrated bedload transport magnitude is also affected offshore by some 20–30%.

The highest relative differences with the base run are found for suspended load during the storm. Here, relative differences of more than 100% occur in all scenarios (Figure 5.12). The effects of changing roughness are not confined to the shoals anymore as during calm weather, but are found throughout the model domain. Even when a relaxed ripple roughness (scenario 6) was used, relative differences of up to 70% were found, even though the current was hardly affected. This implies that the change in sediment transport is caused by a change in sediment concentration (see Equation (5.9)). Apart from the scenarios with the constant Chézy value (scenario 4) or the doubled megaripple roughness (scenario 3), the eastern shoals are most affected. At this location the ripples in the base scenario made up to 60% of the total roughness during the storm (Figure 5.5). The constant Chézy value was chosen based on an average depth, so offshore and in the deeper parts of the channel, where the water depth is larger than the average, the Chézy value was too high. In all other simulations, the change in transport in the channel is near 0. When the megaripple roughness was doubled, no clear effect was found on the shoals and in the channel, and a larger effect was seen offshore, in the area where megaripples are dominant. As described before, the suspended load does not directly depend on $k_{s,mr}$ (Figure 5.1), implying that the change in transport in this scenario was caused by the changes in hydrodynamics. Lastly, it should

Figure 5.10 Time-integrated bedload transport q_b during calm weather (1), and absolute relative difference in q_b of all other scenarios with the base scenario during calm weather (2–9).

be noted that low relative differences were found for all scenarios at the northern part of the delta, where the absolute values of the suspended load transport magnitude in the base run were highest.

5.4 Discussion

The aim of this paper was to assess the effect of the roughness induced by small-scale ripples and megaripples on hydrodynamics and sediment transport in Delft3D. First of all, the modelled ripples were compared to measurements. Measured ripples were found to have relatively stable dimensions. Ripple heights were only weakly related to orbital velocity (which is a function of wave height and water depth), while no relation was found at all between ripple dimensions and current velocity (Chapter 4). Therefore, it was suggested in Chapter 4 to use a constant ripple roughness in modelling. In the present study, the effects of a constant ripple roughness (scenarios 7, 8 and 9) were compared to those of the default Van Rijn roughness predictor (scenarios 2, 3 and 6) and to situations with roughness incorporated into constant Chézy or Manning values (scenarios 4 and 5).

Figure 5.3 showed that the ripple-related roughness simulated by using a 180-min relaxation time was closest to the measured roughness. Yet, this does not mean that this scenario

Figure 5.11 Time-integrated bedload transport q_b during the storm (1), and absolute relative difference in q_b of all other scenarios with the base scenario during the storm (2–9).

also gives the best resemblance of the true current and sediment transport. The current in this scenario changes with only 2% with respect to the base scenario. However, no measures of sediment transport are available of the studied location and time period, so it is unclear if the base scenario or the relaxation scenario performs best.

Based on the importance of small-scale ripples and megaripples for the total roughness, it was expected that changes in the small-scale ripple roughness would mainly affect the shoals during calm weather, while megaripples will affect the offshore area and the channels, mainly during storms. This is also confirmed by the model results. During calm weather, both the current and the time-integrated sediment transport are mainly affected on the shallow parts of the delta. During storm, the offshore area is affected as well. In the deep channel in between the islands, the current depends on the chosen roughness scenario, but the transport does not. Since sediment transport q is a function of current velocity u and sediment concentration C , the fact that transport does not change means that higher ripple roughness also causes a higher depth-integrated sediment concentration. It is expected that these different effects of small-scale ripples and megaripple roughness will also be found on other ebb-tidal deltas, since current and depth distributions in space will be similar (i.e., alternating shoals and channels). Yet, differences in grain size distribution might also cause different ripple dimensions and behaviour (Chapter 4), so one should be careful in extrapolating the results to other deltaic systems.

Figure 5.12 Time-integrated suspended load transport q_s during the storm (1), and absolute relative difference in q_s of all other scenarios with the base scenario during the storm (2–9).

The transport in the channel is only affected when a constant Chézy value is chosen. This is because this Chézy value is constant throughout the model domain, so it is not dependent on depth. Therefore, in areas deeper than the average depth, the prescribed roughness is higher than when a depth-dependent C is chosen. This is also the case on shallow areas during the storm when constant roughness values are prescribed. Both in the field data and in the model, mobility parameter ψ often exceeds 250 during the storm, which results in the small-scale ripples being washed away. Therefore, when the ripples are forced to be constant in the model, they are too high, resulting in large errors in current velocity and sediment transport. Therefore, contrary to what was suggested in Chapter 4, a constant ripple roughness should not be implemented in morphologic studies, especially when shallow (<10 m) areas and/or storm periods are modelled. Davies and Robins (2017) drew a similar conclusion when studying bedforms in an estuary with TELEMAC. Even though their study concerned megaripples and dunes predicted by van Rijn (2007), they also found that a spatio-temporally varying k_s resulted in the most accurate prediction of local hydrodynamics. This gives the use of the Van Rijn roughness predictor an advantage over choosing a constant ripple roughness or a constant Chézy value.

The Delft3D model, similar to other numerical morphodynamic models such as MIKE21, TELEMAC and ROMS, is highly complicated and involves numerous feedbacks (see Figure 5.1). While this resembles reality better, this also makes it difficult to say whether changes

in sediment transport are mainly caused directly by changes in roughness or indirectly by changes in hydrodynamics. Figure 5.1 shows that both causes are possible. In the scenarios with doubled small-scale ripple roughness, relaxed ripple roughness, and a constant very small ripple roughness, the current velocities during the storm in most parts of the domain change with 2% or less. Yet, during storm, the suspended sediment transport in these scenarios changes by 60–70%. This implies that the direct effect of bedform-related roughness is very important: it increases the depth-averaged sediment concentration and therefore also q_s .

In short, the largest effect of roughness is found in the depth-averaged current velocity and the suspended load transport. Changes in transport magnitude appear larger than changes in current magnitude. Yet, it is unknown which of the scenarios shows the most realistic amount of sediment transport, and although the differences with the base scenario are higher, the scenarios with the constant ripple roughness do show similar patterns in MARD-values as the Van Rijn scenario with high ripples. It therefore might be valid to use a spatio-temporally constant roughness value in exploratory studies, when the 'true' behaviour of the system is unknown, (e.g., determining causes for spatial patterns or morphologic behaviour). Therefore, it is especially important to put effort into choosing a roughness predictor when transport or morphologic calculations are made, e.g., in hindcasting studies. If only hydrodynamic simulations are done, the errors are expected to remain mostly below 10%, and they will be located at the shallow areas. Only if a constant Manning's n is prescribed, errors can increase further, in our case the differences with the base scenario were up to 40%. Moreover, even with small changes in ripple-related roughness $k_{s,r}$, the sediment transport magnitude still changes. This means that it is important to carefully consider which ripple roughness predictor is used and that coastal models that simulate sediment transport or morphological change should incorporate spatio-temporally varying ripple roughness predictors. Also, a model user should always keep in mind that over time, small changes in the predicted ripple roughness could lead to large changes in the predicted total sediment transport and thus morphology.

The present study is representative of the sediment transport predictor of van Rijn (2007). This predictor can also be incorporated into other models, such as TELEMAC. It is hypothesised that the results would be similar if a predictor was used that also calculates bedload and suspended load transport separately. Since the present study showed that currents are affected by up to 60% when the roughness scenario is changed, most bedload predictors will be affected by changes in roughness. This is because in all bedload predictors, and even some total load predictors such as the one from Engelund-Hansen (Engelund and Hansen, 1967), the transport depends on the depth-averaged current velocity to a power of 3–5. The calculation of suspended load varies more between various predictors, because it mainly depends on the calculation of the reference height. Therefore, it is advised to perform a similar sensitivity study when a different transport predictor is used.

Further research should also address the long-term morphodynamic feedbacks. Spatial differences in sediment transport can lead to different morphologic development, so the effects of roughness that were found in the present study could be amplified or reduced by these feedbacks. Furthermore, the interaction between the different bedform scales should be addressed. Nikora et al. (1997) showed that the migration of larger bedforms (megaripples in our case) could be caused by the movement of small-scale ripples. A time series of measurements of both scales at the same location is, however, not available for our study site.

5.5 Conclusions

For the first time, the effect of small-scale ripples and megaripples on hydrodynamics and sediment transport was quantified in detail using Delft3D scenarios for a mixed wave-current environment (the Ameland ebb-tidal delta). In Delft3D, small-scale ripples and megaripples together constitute the bedform-related roughness. Although small-scale ripples are lower than megaripples, it was found that small-scale ripple-related roughness can make up to 100% of the total roughness on shallow parts of the ebb-tidal delta. Megaripples are more important in the current-dominated channel and further offshore. The bedforms were found to affect the current velocity and sediment transport on the shallow parts of the ebb-tidal delta. Current velocity magnitudes affected by different bedform roughness scenarios resulted in overall differences of 10–20%, both during calm weather and during storm. In sediment transport, on the other hand, small changes in predicted ripple roughness lead to changes of more than 100%. Transport and flow directions were not significantly affected. During a storm, small-scale ripples are predicted to be washed away, but megaripples remain. This process cannot be simulated if a constant roughness value is prescribed, leading to overestimations in sediment transport. Therefore, it is advised to use a spatio-temporally variable bedform roughness in morphodynamic simulations in Delft3D. Future research should address the long-term morphodynamic feedbacks and the interaction between the different bedform scales.

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Chapter 6 | Synthesis

The main aims of this thesis were (1) to determine the spatio-temporal dynamics of bedforms on ebb-tidal deltas on a variety of spatial scales, (2) compare these characteristics to existing bedform roughness predictors and (3) to determine how predicted sediment transport in mixed-energy environments is affected by predicted bedform roughness. Both direct and indirect effects were considered. The direct effects of bedforms on sediment transport are mainly in the bed load regime through migration of relatively large bedforms, while indirect effects are caused by the influence that the bedforms have on bed-shear stresses and vertical mixing, thereby affecting the suspended load transport. Often, numerical models are used to predict the long-term morphological evolution of ebb-tidal deltas. These models rely on bedform predictors, which have often been developed for wave- or current-dominated environments, and have not been tested for mixed energy environments like ebb-tidal deltas. It was thus unclear how well bedform roughness predictors performed in these types of environments, and how the predicted sediment transport was affected by the bedform roughness. This chapter presents a discussion on the ways in which this thesis has improved our knowledge on the interaction between bedforms and sediment transport.

First, the main conclusions of this thesis are summarized and compared to other studies, answering the research questions that were formulated in Chapter 1. Starting with the direct effects of bedforms on sediment transport, bedforms of all spatial scales are characterized, followed by a study of their indirect effects on sediment transport. After that, in Section 6.2 the long-term implications of this research are discussed, as well as the use of bedform roughness in the modelling of sediment transport pathways. This will strengthen the science base for coastal management strategies. This chapter will end with suggestions for future research.

6.1 Main conclusions

6.1.1 What are the characteristics and dynamics of saw-tooth bars on the ebb-tidal deltas in the Wadden Sea and how do they affect sediment transport?

Saw-tooth bars are large sand bars that are found on the downdrift part of most ebb-tidal deltas in the Dutch and German parts of the Wadden Sea. In Chapter 2, they were studied using bathymetries of 9 tidal inlets from 1970 to 2015. On average, saw-tooth bars were 0.74 m high and had an average wave length of 670 m and an average cross-shore width of 1442 m. The bars were located in water depths between 3 and 12 m and migrated with a velocity of 30 m/y on average. Both heights and wave lengths fluctuated over time, with the time scales ranging between 5 and 30 years and decreasing from west to east along the Wadden Sea region. The orientation of the bars was approximately 66° between the bar crest and the underlying depth contours for all inlets.

Based on their height, width and migration speed, a first estimate of the saw-tooth bar related direct sediment transport could be given. The data set average volumetric sediment transport by saw-tooth bars was about $9.500 \text{ m}^3/\text{y}$, with maxima of just less than $150.000 \text{ m}^3/\text{y}$. The average amount of transport by saw-tooth bars decreased in the eastward direction, and the maximum was found at the Ameland inlet. This estimation used the assumption that the bedforms are triangular in shape, while in fact they were more or less sinusoidal. Since the area under a sinusoid is larger than the area under a triangle with the same width and height, the transport values were probably underestimated. However, this is still a useful tool to get an indication of the transport patterns over space and time. Ridderinkhof et al. (2016b) studied the longshore transport for the same inlets as in Chapter 2, and found values ranging between 100.000 and $1.400.000 \text{ m}^3/\text{y}$. At the Ameland inlet, the total longshore sediment transport was found to be $1.000.000 \text{ m}^3/\text{y}$, so the maximum contribution of saw-tooth bars was approximately 15%. The average contribution of saw-tooth bars to the longshore transport ranged between 0.5 and 2%, with the largest contribution in the Ameland inlet and the smallest in the Oetzumer Balje inlet. Thus, on average the contribution of saw-tooth bars to the longshore transport was relatively small, but it could become significant in years in which the saw-tooth bar height and migration speed are at their maximum.

Their indirect effect on sediment transport was not studied in this thesis because saw-tooth bars are large enough to be incorporated directly in the bathymetry of numerical models, without parametrization. After Chapter 2 was completed, van der Heijden (2021) used a 2DH Delft3D model to further explore the generation mechanisms of saw-tooth bars. The model results indicated that both the growth and migration of the bars were determined by waves. Saw-tooth bars are formed by high waves during storms, and decay during prolonged periods of fair weather conditions. Since their height is large with respect to the total water depth, the flow and sediment transport change in between the crest and trough. However, this process is opposite on both sides of the crest and there will be no or only a small large-scale averaged effect. On the other hand, saw-tooth bars do affect the longshore drift near the adjacent barrier island coast (van der Heijden, 2021). This drift decreases, because more waves dissipate on the shallow bar crests, which decreases the amount of wave energy in the surf zone.

Saw-tooth bar height and migration speed were positively correlated, which means that larger bars migrated faster than the smaller ones. In most natural systems, this correlation is negative (larger bedforms move slower (e.g. Wilbers and Brinke, 2003)), because for larger bedforms more sediment displacement is needed to move them. A possible explanation for the positive correlation found here is that saw-tooth bar migration is caused by wave action (van der Heijden, 2021); more sediment can be stirred when the bars are higher, because in this case the waves are more likely to feel the bed. However, this would also imply that the shallower bars would move faster, but Chapter 2 showed no significant correlation between depth of occurrence and migration speed.

6.1.2 What are the characteristics and dynamics of megaripples on the Ameland ebb-tidal delta, and how do these depend on time-varying hydrodynamic forcing?

On a spatial scale smaller than the saw-tooth bars, sand waves and megaripples can be found in tidal inlet systems. Sand wave migration can be important for the total amount of sediment transport (Buijsman and Ridderinkhof, 2008b; Ernstsens et al., 2006), but in this thesis their

dynamics could not be studied with the available data sets. Megaripples on the other hand, were observed in the Ameland inlet and its accompanying ebb-tidal delta using multibeam echosoundings, and their characteristics were studied in Chapters 3 and 4. Three multibeam data sets were available: one covering the eastern part of the main channel for multiple years with one bathymetric measurement each year, one covering several parts of the main and secondary channels over the period of a spring-neap tidal cycle during the SEAWAD field campaign in the autumn of 2017 (van Prooijen et al., 2020), and one covering the areas near the frames that were used during the same field campaign. A Delft3D model of the Ameland inlet was available, which was calibrated and validated against the 2017 field campaign. Thus, predicted current velocities were available at all study areas for the SEAWAD data set of megaripples.

It was found that megaripples were between 2 and 28 m long (average of 11.2 m) and between 0.05 and 1 m high (average of 0.46 m). In the deepest part of the main inlet channel, megaripples were superimposed on sand waves. Over a single tidal cycle or a spring-neap tidal cycle, their heights and lengths did not vary considerably. However, over the years, the standard deviation in wavelengths was 4 m, and the standard deviation in heights was 0.1 m. In the channel and on the northern part of the delta, the megaripples had relatively elongated crests, while they were more three-dimensional near the western delta shoal. It was also found that the bed was flattened at a few locations after a storm with wave heights in the study area of up to 3.5 m and near bed current speeds of up to 1 m/s.

Apart from heights and lengths, the asymmetry, slip face angle and migration of megaripples were determined as well. It was found that megaripples were relatively sinusoidal (not asymmetric), with average slip face angles of 8.6° (standard deviation of 3.6°), although in some cases slip faces of up to 20° were found as well. During one tidal cycle, megaripples migrated up to 4 m, but migration only started when the flow exceeded 1.2 m/s. The small variation in megaripple characteristics over a tidal cycle indicated that they do not respond immediately to the changes in hydrodynamics either. Instead, the data presented in Chapter 3 suggest that the height and asymmetry of the megaripples increase with ebb- or flood-dominance (the difference between maximum ebb and flood current velocity per tide). A similar relation was found between the migration and the ebb- or flood-dominance, but this relation showed more scatter than the ones between height and asymmetry and ebb- or flood dominance. These relations indicate that when either ebb or flood is dominant, there is more energy in the flow pushing the sand in a certain direction, leading to higher and more saw-tooth shaped megaripples.

Similar to the saw-tooth bars, megaripples can also directly transport sediment, and based on the measurements in Chapter 3 it was found that the average amount of direct sediment transport by megaripples in the tidal inlet channel was $88 \text{ m}^3/\text{m}/\text{y}$. This is more than 10 times as much as sediment as transported by saw-tooth bars, when the $9603 \text{ m}^3/\text{y}$ is normalized by the average saw-tooth bar width to $6.7 \text{ m}^3/\text{m}/\text{y}$. Yet, the higher transport by megaripples is mainly caused by their much higher migration speeds of approximately $600 \text{ m}/\text{y}$, which was based on measurements from only one tidal cycle. The actual migration of megaripples during a year is not known.

It is expected that the megaripple dimensions and behaviour that were observed in this thesis for the Ameland inlet are representative for other tidal inlets and current- or wave-current-dominated environments. In 2017, a pilot nourishment of 5 Mm^3 was added to the western ebb shoal of the Ameland ebb-tidal delta. First results of the monitoring indicate that

megaripples appear within months on the added sand (van Rhijn, 2019). Once they appear, their characteristics and behaviour are in the same range as the other megaripples observed in the Ameland Inlet system; they are asymmetric, with heights in the order of 0.1-0.25 m and lengths in the order of 5-10 m, and they migrate in the direction of the dominant wave- or tide-induced current. In Chapter 3, it was shown that the hydrodynamics under which the megaripples were observed at the Ameland ebb-tidal delta were current-dominated to wave-current dominated. Furthermore, the megaripples as found in the Ameland inlet had similar characteristics to several other studies, like those of Dalrymple et al. (1978), which was conducted in a tidal inlet as well. Megaripples in wave-dominated surf zones, on the other hand, are much smaller than those in tidal environments; they have maximum heights of 0.1 m and maximum lengths of 2.5 m (Clarke and Werner, 2004; Larsen et al., 2015). Therefore, the results of the present study are probably not applicable to wave-dominated environments.

6.1.3 What are the characteristics and dynamics of small-scale wave-current ripples on the Ameland ebb-tidal delta, and how do these depend on time-varying hydrodynamic forcing?

In order to study small-scale ripples, five locations on the Ameland ebb-tidal delta were monitored with 3D profiling Sonars, which made scans of the bed once per hour for 6 consecutive weeks (Chapter 4). At the same locations, wave- and current velocities were measured using ADVs. The ratio between wave- and current-related shear stress proved that most locations, except the inlet channel, are indeed in a mixed energy regime, but wave-dominated and current-dominated conditions were observed as well.

Small-scale ripples were found on all studied locations, apart from during the storm. Often, they were superimposed on megaripples. In general, ripple heights ranged between 0.005 and 0.025 m, with an average of 0.015 m, while the ripple lengths were in between 0.06 and 0.18 m, with an average of 0.11 m. The ripples were always highly three-dimensional. Because the range in length, height and three-dimensionality was small in both time and space, the ripple characteristics and the hydrodynamic variables were largely unrelated. Only the ripple height and the wave-related near bed orbital velocity were related to each other. With increasing orbital velocities, the ripple heights decreased. It was observed that the ripples do migrate and change in planform, even over the course of an hour, but their characteristics would still remain the same in that case.

It is difficult to determine how representative the ripple measurements on the Ameland ebb-tidal delta are for other ebb-tidal deltas and other coastal environments. In environments where waves or currents are dominant, for example in estuaries or shallow straight coasts, the ripples are straight-crested (two-dimensional) with a large range in heights and lengths (Dolphin and Vincent, 2009; Wengrove et al., 2018). In this thesis, ripples were observed for a large range of wave-dominated, current-dominated and wave-current dominated conditions, but no transition was found to larger and more two-dimensional ripples. In Chapter 4, it was hypothesized that the low variability in ripple size and three-dimensionality was related to the small grain size that was found in the Ameland inlet (0.211 mm). This is supported by the studies of Krämer and Winter (2016) and Bolaños et al. (2012), who found relatively small and three-dimensional bedforms as well in environments with grain sizes <0.255 mm (on a continental shelf and in a sandy embayment, respectively). In both the Dutch and German parts of the Wadden Sea, the grain size on most ebb-tidal deltas is often lower than 0.300

mm (RIKZ and GEODAN IT, 1998; Son et al., 2010), so the results described in this thesis are likely to be representative of all Wadden Sea ebb-tidal deltas.

6.1.4 What are the effects of medium and small-scale bedforms on hydrodynamics and sediment transport on ebb-tidal deltas?

Bedforms on all scales affect both hydrodynamics and sediment transport. While the direct effects by migration were discussed in Sections 6.1.1 - 6.1.3, the indirect effects could only be solved by numerical modelling. Bedforms with length scales smaller than approximately 50 m, so megaripples and ripples, are often parametrized in numerical models, because they are smaller than the typical grid cell size. Bedforms influence the vertical mixing of sediment and the drag experienced by the flow, with higher bedforms causing more drag for the same flow velocity. To include this effect in model simulations, bedform predictors are used. In Delft3D the van Rijn (2007) predictor is used, in which the roughness depends on the bedform height. In the model, contrary to the natural situation, the small-scale ripple roughness height affects the sediment transport directly through the reference height for suspended load transport and indirectly through the affected current and vertical mixing, while the megaripple-related roughness height only affects the sediment transport indirectly. In Chapters 3 and 4, the bedform roughness height as predicted by van Rijn (2007) was compared to measured heights of megaripples and small-scale ripples, respectively. In both cases, the predicted roughness height was lower than the measured ripple or megaripple height. For the megaripples, the predicted roughness height was approximately half of the measured roughness height, while for the small-scale ripples the predicted roughness height was more in the order of 75% of the measured roughness height. Also, both for megaripple and small-scale ripples predicted roughness heights were much more variable in time and space than measured values. The roughness predictors depend on the current velocity, which changes during a tidal cycle, while the field data showed no or only weak sensitivity to current velocity.

In Chapter 5, the effect of the bedform roughness parametrization on currents and sediment transport was tested by means of a sensitivity study. Several roughness scenarios were compared to each other in a Delft3D model of the Ameland Inlet. In the base scenario, the van Rijn (2007) roughness predictor was used. In 8 other scenarios the megaripple or ripple roughness were doubled, the ripple roughness was kept constant or did not react immediately to the hydrodynamics, or a constant Chézy or Manning coefficient was used instead of the ripple roughness prediction.

The model correctly predicted that the megaripples were mainly found in the channels and the deeper area offshore of the Ameland ebb-tidal delta. On the shallow shoals, mainly small-scale ripples were found. Current velocities were in general affected by up to 20% when the roughness scenario was changed, with maximum changes of up to 70% at the edges of the delta during a storm with high megaripples or when a constant Chézy or Manning coefficient were used. The choice of roughness parametrization has a larger effect on the sediment transport than on the current. Predicted bed load transport was always 15-20 times smaller than the suspended load, so even though changes of up to 100% were found in the bed load transport, its effect on the total amount of sediment transport was small. For the suspended load transport, however, small changes in ripple roughness, for example using a 180-minutes relaxation time, could also result in transport changes of up to 100% during a storm. These changes were mainly found on the shallow shoals and offshore; the locations where the transport was largest in the base scenario, in the channels and at the edges of the

delta, were affected by less than 5%. Most change compared to the base scenario was found when a Manning or Chézy roughness coefficient was used, or when the ripples were replaced by only grain-related roughness. During low energy conditions (significant wave height in the order of 0.1 m), maximum changes of up to 300% were found, especially when the ripple roughness was set to a constant high value. However, the largest changes during calm weather were found at locations where the transport in the base scenario was near zero.

Even though only the roughness predictor of van Rijn (2007) was used, the study is representative of other models and predictors as well. For example, the van Rijn (2007) predictor can also be used in other models, such as TELEMAC. Furthermore, most other roughness predictors, like the ones created by Nielsen (1979), Tanaka and Dang (1996) and Soulsby et al. (2012), are also highly dependent on current velocities (Chapter 4). Moreover, the predictors mentioned here incorporate the waves as well, albeit all in a slightly different way. Thus, even though different predictors will result in different roughness height and therefore also sediment transport rates, the fact they have similar sensitivity to waves and currents will result in similar patterns of sediment transport in space and time.

6.2 Discussion and perspectives

6.2.1 The cumulative effect of the parametrization of bedforms and other parameters in numerical sediment transport modelling

In this thesis, bedforms and their effect on sediment transport have been considered, but the question remains how they affect the large-scale and long-term morphology. Figure 6.1 shows that spatial and temporal scales of the tidal inlet morphology are related to each other. In the central part of the figure, numerical morphodynamic models are represented, showing that they only explicitly solve certain spatial and temporal scales. Processes acting on smaller spatial and temporal scales, like megaripples and small-scale ripples, are parametrized, and the effects of this parametrization on sediment transport were quantified in Chapter 5. However, bedforms do not interact solely with hydrodynamics, but also with sedimentological and ecological properties. This is well-reflected in the other SEAWAD subprojects (Figure 6.1). Some of these processes are parametrized separately, but the cumulative effect of all parametrizations is unknown.

The first SEAWAD subproject studied the effect of tidal currents on wave-related transport. The tidal currents across the Ameland ebb-tidal delta affect the wave shape, especially on the shallow shoals (de Wit et al., 2019). Since the wave-related sediment transport is determined by the wave shape, changes in shape might also lead to changes in transport. However, single waves are not resolved in morphodynamic models, so their shape and the amount of energy they contain are parametrized as well. Boechat Albernaz et al. (2019) found that the wave-related transport parametrization that is used by default in Delft3D, which was also used in the present study, requires strong calibration. Waves can transport sediment both in the onshore direction (in the direction of wave propagation), and in the offshore direction (against the direction of wave propagation). In 2DH models, vertical offshore processes such as the undertow are often not included (Boechat Albernaz et al., 2019). As a result, the onshore component of wave-related sediment transport should be reduced in order to prevent unrealistic nearshore sediment accretion. However, by doing so the amount of longshore transport significantly decreases (by more than a factor 2), which in the case of an ebb-tidal delta would also lead to lower predicted migration of the shoals. For longer-term modeling,

Figure 6.1 Sketch of the relation between spatial and temporal scales in the morphodynamic system of a tidal inlet and its accompanying ebb-tidal delta. Chapters and processes which are the topic of this thesis are shown by solid lines, other processes are shown by dashed lines.

Boechat Albernaz et al. (2019) showed that a different wave parametrization leads to a more realistic evolution of the morphology, which in the case of the Ameland inlet would lead to an increased amount of sediment transport and an increased shoal migration velocity. Most locations that were studied in this thesis were dominated by both waves and currents (Chapter 3 and 4), so the effect of wave shape parametrization on ebb-tidal deltas is probably of comparable significance as the bedform parametrization.

Second, research was done in a SEAWAD subproject on the importance of grain size on sediment transport. In Chapter 4 it was suggested that the lack of variation in small-scale ripple dimensions was caused by the small grain size, but ripple dimensions will increase with increasing grain sizes above $200\ \mu\text{m}$ (Bartholdy et al., 2015). Herrling and Winter (2014) showed that sediment transport pathways depend on the grain size of the sediment, although the smallest grain size they used in their study was $150\ \mu\text{m}$, which is still fine sand. On the Ameland ebb-tidal delta the mud content was less than 1% by volume (Pearson et al., 2019), but during ebb, mud is transported from the Wadden Sea towards the North Sea, leading to temporary high mud contributions of up to 98% of the total sediment load in the water column (Pearson et al., 2021). It is unknown if and if so, where, this mud settles down, but if it does, it will affect the bedform dynamics, as sand and mud behave differently. In muddy environments it is difficult for bedforms of all scales to develop. For example, sand waves are ubiquitous in estuaries, except at the location of the estuarine turbidity maximum, where the mud content in the water is highest (Herrling et al., 2021). Megaripple lengths and heights linearly decrease with an increasing mud content, and they disappear when the mud content

of the bed exceeds 15% by volume (Schindler et al., 2015). Small-scale current ripples can form at least until the mud content of the bed reaches 18% (Baas et al., 2013), but ripple formation takes more time when the bed contains more mud (Wu et al., 2018). Across the larger bedforms, mud deposits are often patchy and variable in space (10s of meters) and time (single tidal cycle), and a part of the mud might consolidate into a thin, erosion-resistant layer (Schrottke et al., 2006). Local resistance to erosion would change the gradients in sediment transport, which might lead to variations in bedform shape. This might also happen on the sand waves in the Ameland inlet and on the saw-tooth bars on the Ameland ebb-tidal delta. Yet, the resolution of the measurements of the saw-tooth bars was too large to verify this, and the measurements of the sand waves were too scarce to determine if this process is relevant. It is expected that the effect of mud is mainly important in the back-barrier basin, which is mud-dominated, but changes in the basin might also affect the ebb-tidal delta.

The last SEAWAD subproject studied the interaction between benthic ecology and sediment transport. Previous research found that sand wave areas with complex sediment transport pathways are associated with high biodiversity (van Lancker, 2016). First, bedform crests are subjected to shallower water depths and larger current velocities than the troughs. As a result, Ramey et al. (2009) found that benthic species communities were different between ripple crests and troughs, even on small-scale ripples with heights of 5-15 cm and lengths of 10-100 cm. This was also found for sand waves, although the macro-scale morphology and the seasons are more important (Baptist et al., 2006; Damveld et al., 2018). Second, pressure gradients in the flow over bedforms (caused by flow compression over the crest) drive pore water flow within the bedform (Ahmerkamp et al., 2015). This enhances the oxygen flux through the bedform, which supports benthic life, although less oxygen is supplied to the benthos when bedforms migrate. Yet, the benthos can also affect the bedforms, for example when the benthic species trap extra fine material (Passchier and Kleinhans, 2005) or when they bury themselves in the sand and disturb the bed, thereby destroying small-scale ripples, while also bringing small amounts of sediment into suspension. The cumulative effect of benthos on bedforms on the larger spatial scale of the ebb-tidal delta is not yet known.

6.2.2 The importance of bedforms on larger spatial and temporal scales

Processes that are important on the long term and on large spatial scales are the bathymetry and the climate, and they were assumed to be constant during the model simulations (Figure 6.1). Climate change might affect the sediment transport on ebb-tidal deltas, but the spatio-temporal scales of these effects are much larger than those considered in this thesis. Thus, the effect of climate change on bedform-induced sediment transport cannot be determined here. On time scales of years to decades, small differences in the parametrizations of bedforms used in the model might lead to large variations in sediment transport and morphology. Therefore, it is difficult to say how sediment transport pathways change on the long term. In Chapter 5 the effect of roughness parametrizations on sediment transport was studied. Wang et al. (2016) studied the representation of dunes (sand waves) by roughness in Delft3D. They found that dune roughness as calculated by van Rijn (2007) varied with tidal cycles, just like the ripple and megaripple roughness as modelled in Chapters 3 and 5. Similar to the ripples and megaripples studied in this thesis, dunes in the field data of Wang et al. (2016) show much less variability in space and time than the van Rijn (2007) predictor suggests. On the long term, reduced roughness variability in the model leads to lower depth-averaged current velocities and thus to less incision of the tidal inlet channel (Wang et al., 2016). Similar effects

are expected for the ripple- and megaripple-related roughness. All across the delta, lower current velocities would lead to more wave dominance, as the current-related shear stress decreases while the wave-related shear stress remains the same. On the shallow shoals, where the ripple-related roughness was found to be most important, landward sediment transport caused by waves and tides is dominant over seaward transport, which is caused by only tides (Herrling and Winter, 2014; Lenstra et al., 2019; Ridderinkhof et al., 2016a). Thus, more wave dominance will probably lead to more predicted onshore transport, because the averaged seaward transport caused by tides decreases.

Gradients in sediment transport are important indicators for morphologic change. In the default scenario in Chapter 5, the largest gradients in both bed load and suspended load transport were found at the edges of the shoals during storms. On the long term, these gradients will lead to the migration and deformation of the shoals. In the other scenarios that were tested in Chapter 5, the locations and amounts of maximum sediment transport did not change, but in the surroundings differences were found. This means that the gradients in sediment transport will change if a different roughness predictor is used. In all scenarios, except the one with no ripples, the gradients decreased. Thus, in most scenarios, the morphological changes will be slower, thereby increasing the typical timescales associated with moving shoals, and only when the ripple roughness height is equal to the grain size, morphological change will be faster. The largest decrease in shoal migration velocity is expected if a constant Chézy or Manning value is used as roughness parameter, because they result in the smallest gradients in sediment transport. On short timescales, this effect is expected to be small, because the absolute difference in gradients between the scenarios was small as well, but on longer time scales it might become important because of morphodynamic feedbacks.

In this respect, it is important to consider the roughness predictors that are used in morphodynamic studies. Lenstra et al. (2019) used a constant Chézy value of $63 \text{ m}^{0.5}/\text{s}$ in modelling the Ameland ebb-tidal delta, while Ridderinkhof et al. (2016a) used the van Rijn (2007) roughness predictor while studying the same inlet. Ridderinkhof et al. (2016a) found a shoal migration of 629 m/y , which is much larger than observed in nature. It is hypothesized that the shoal migration as modelled by Lenstra et al. (2019) will be lower, but it was not determined in that study.

6.2.3 Suggestions for future research

For coastal management purposes, it is important to understand the interaction between bedforms and large-scale ebb-tidal delta nourishments. Ebb-tidal deltas can be nourished in order to increase their size, and thus their protective effect, or to enhance sediment transport into the basin. The large-scale effects of an ebb-tidal delta nourishment on the cyclic channel-shoal dynamics were discussed by Lenstra (2020), and they are different for different sizes and nourishment locations. Of course, if the size and shape of the ebb-tidal delta are altered by a large nourishment, this also affects the wave-current ratio. Furthermore, the location and rotation of the tidal inlet channels might be altered after a nourishment (Lenstra, 2020), which might also affect the ebb- or flood dominance at certain location. If this happens, megaripple-related roughness will change, and the sediment transport caused by migrating megaripples will also change in direction. However, since there are numerous unknown variables in this prediction, this should be addressed by future research using a morphodynamic model including morphologic updates.

Apart from the different sizes and locations of future nourishments, their composition is important as well. In Chapter 4, it was hypothesized that the relatively constant ripple height on the Ameland ebb-tidal delta was caused by the fine sand at this location. Thus, when a nourishment with coarser sediment is applied, the bedforms on the nourishment might be different in shape and behaviour, possibly causing different sediment transport pathways. In the pilot nourishment at the Ameland ebb-tidal delta, fine sediment of 0.175 mm was used (van Rhijn, 2019), so in this case no such differences are expected. Future research should address the importance of grain size for the characteristics and behaviour of bedforms and their effect on sediment transport.

Furthermore, the effects of roughness calibration on sediment transport should be studied. Chapters 3-5 have shown that a calibrated model using a bedform-related roughness that may deviate considerably from the measured bedform heights, still gives plausible results in terms of hydrodynamics and sediment transport. The question remains whether the measured roughness is incorrect (since roughness cannot be measured directly, which is a form of underdetermination (Kleinhans et al., 2010)), or that all other processes incorporated in the model, like wave-driven and current-driven sediment transport processes, require an increased bed roughness to function. This last option is a type of equifinality (Von Bertalanffy, 1950) so it is probably very difficult to untangle all processes in the model. It might also be the case that the connection between bedform height and its related roughness is weaker than presumed, and that factors like slip face steepness and the relation between megaripple roughness and tidal current dominance should be incorporated into roughness predictors as well. This requires more measurements of ebb and flood dominance, because this dominance will vary over space and over spring-neap tidal cycles. Together with the bedform steepness, this might lead to highly varying roughness values between ebb and flood, which in turn may considerably affect sediment transport.

This thesis has provided insight into the underlying processes in numerical modelling. When this insight is improved even further, the behaviour of nourishments on ebb-tidal deltas can be predicted with more confidence, which will eventually improve coastal safety.

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About the author

Laura Brakenhoff was born on the 23rd of August 1992 in Blaricum, the Netherlands. From a very young age, she wanted to become an environmental specialist, in order to reduce climate change and make the world a better place. Therefore, she started studying Earth Sciences at Utrecht University in 2010. During the bachelor, she got involved in a KNAW internship in the Coastal Group, which sparked her interest in the Dutch coast. Therefore, in 2013 she enrolled in the master's programme Earth Surface and Water, again at Utrecht University, specializing in coastal morphodynamics and fluvial systems. This culminated in a master's thesis on the interaction of ground water and surface moisture at the Dutch beach near Egmond. After obtaining the master's degree in 2015, she started working at a consultancy company, but she returned to the UU six months later to pursue a PhD. Besides writing this thesis, she also chaired the PhD council of the faculty of Geosciences for a year and contributed to outreach to elementary school children during several editions of the Utrecht Summer School Junior. Furthermore, she was involved in organizing the yearly conference of the Dutch coastal community, the NCK days, in 2019. She also did two projects for Rijkswaterstaat, to help them incorporating the methods used in her PhD in their daily activities.

List of publications

Journal papers

Brakenhoff, L.B., R. Schrijvershof, J.J. van der Werf, B. Grasmeijer, B.G. Ruessink and M. van der Vegt (2020). From Ripples to Large-Scale Sand Transport: The Effects of Bedform-Related Roughness on Hydrodynamics and Sediment Transport Patterns in Delft3D. *Journal of Marine Science and Engineering*, 8(11): 892. <https://doi.org/10.3390/jmse8110892>.

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