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Experiments and new observation techniques related to wave-ice interactions

Thesis submitted for the degree of Philosophiae Doctor

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Preface

This thesis is submitted in partial fulfillment of the requirements for the degree of *Philosophiae Doctor* at the University of Oslo. The research presented here was conducted at the University of Oslo, The University Centre in Svalbard and The Western Norway University of Applied Science in the period between 2018 and 2021, under the supervision of Pr. Atle Jensen, Dr. Jean Rabault and Pr. Aleksey Marchenko. This work was supported by The Research Council of Norway under the PETROMAKS2 scheme (project "DOFI", Grant number 28062), the IntPart project Arctic Offshore and Coastal Engineering in Changing Climate (Project number 274951) and the Arctic Field Grant (project "Investigation of iceberg dynamics in Isfjorden", Grant number 310691), and The Norwegian Society of Graduate Technical and Scientific Professionals under the Harald Boes Grant.

The thesis is a collection of six selected papers, presented in chronological order of writing. The common theme is the investigation of wave propagation and attenuation through sea ice, and the development and testing of new measurement techniques for field observations of the related mechanisms. The papers are preceded by an introductory chapter that relates them to each other, presents the scientific context in which the work was undertaken and provides background information and motivation for the work. A second chapter summarizes the main findings of each paper. I certify that this dissertation is mine and that the results presented are the result of the work of our research group, in which I was involved in all scientific processes and brought significant contribution.

Trygve Kvåle Løken
Oslo, September 2021
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I would like to express my gratitude towards my PhD advisors Pr. Atle Jensen, Dr. Jean Rabault and Pr. Aleksey Marchenko. Their guidance and knowledge have been critical for accomplishing the present work. In particular, the engineering and programming skills of Dr. Jean Rabault in the design and construction of measurement instruments have been greatly appreciated. I also want to acknowledge all my co-authors for their contributions, especially our Laboratory Engineer, MSc. Olav Gundersen and PhD student Thea J. Ellevold, whom I have worked closely with. I would like to thank Dr. David Roger Lande-Sudall, Dr. Jan Bartl and Dr. Gloria Stenfelt for our collaboration at MarinLab and for making my stay in Bergen enjoyable. It has been a great pleasure working with this thesis, both on adventurous field expeditions on Svalbard, in the laboratory of our collaborating university in Bergen and at the office in Oslo, and I would like to thank my good colleagues at the Mechanics Section for all the nice moments.

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Chapter 1

Introduction

1.1 Changing Arctic

Sea ice plays an important role for the polar environments. The ice extension and thickness influences Arctic ecosystems as well as global solar reflection and ocean and atmospheric circulation. However, the average temperature has increased globally and especially in the polar regions over the last couple of decades. Consequently, the arctic sea ice cover has reduced in extent to nearly 30% less at the end of the summer melt period compared to a few decades ago and has shifted from predominantly multi-year to seasonal ice (Meier et al., 2013). The entire Arctic Ocean may soon resemble a marginal ice zone (MIZ) in summer, where surface waves now have a much greater role (Thomson et al., 2013). The MIZ comprises a variety of different forms of ice, from grease ice to large ice floes and compact, semi-continuous ice sheets, and extends between fast ice and the open ocean (Newyear & Martin, 1997; Squire et al., 1995). Different ice types found in and around the MIZ are illustrated in Fig. 1.1.

Economical and geopolitical interest in the Arctic combined with the withdrawal of the ice cover has led to increased human activity in the region. New shipping routes, tourism and exploitation of natural resources are some examples (Feltham, 2015; Smith & Stephenson, 2013). Such activities raise the need for better wave forecasts in the MIZ to ensure safe operations (Fritzner et al., 2019). Also, the changing Arctic environment may alter atmosphere-ocean interactions such as momentum and heat transfer. Improved physical understanding of these mechanisms are important for global circulation and climate models.

A more obvious and direct impact of sea ice on human activity, is the fact that large ice floes and icebergs can cause mechanical failure in collision with platforms or ship hulls. Small icebergs, which are more difficult to observe remotely, can still influence damage on constructions (Marchenko et al., 2020). Even though the ice cover is decreasing, icebergs are observed in almost all Arctic seas (Abramov, 1996). Iceberg and ice floe drift are therefore of importance for safe human operations. In addition, icebergs have a small aspect ratio that makes them prone to tipping and rolling (Crocker et al., 1998), which may cause harm to people and equipment in the direct vicinity. Detailed knowledge about iceberg dynamics is of importance when performing activities on and around them, such as iceberg towing to prevent impact with constructions.
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Figure 1.1: Examples of different ice types encountered during the course of the thesis. a) Land fast ice (Svea Bay, 2020). b) Small iceberg (Tempel Fjord, 2020), c) Small ice floes (Longyearbyen, 2020. Credit: Aleksey Marchenko, UNiS).

1.2 Wave-ice interactions

Wave-ice interactions, which are coupled in a nonlinear manner, are key mechanisms for the polar regions. On the one hand, the retreating ice cover leads to larger areas of open water in the Arctic where more energetic waves are generated due to the increasing fetch, which in turn enhance ice break up and fracturing processes in a positive feedback mechanism (Thomson & Rogers, 2014). On the other hand, experimental studies have shown that waves are exponentially damped in the MIZ (Squire & Moore, 1980; Wadhams et al., 1988), meaning that the presence of the MIZ mitigates the break up process of the ice cover. Especially high frequency wind waves are efficiently damped by the presence of ice floes, while low frequency swell can travel far into the ice pack (Squire, 2007). Recently, Squire, 2018 suggested that the assumption of exponential attenuation may only be true for low amplitude waves and that high amplitude waves may adhere to different attenuation laws.

An example of exponential wave attenuation through the Barents Sea MIZ
Wave-ice interactions

from Løken et al., 2021a is presented in Fig. 1.2. The spectral amplitude $a$ (left panel), which can be interpreted as the wave amplitude associated with a certain wave frequency $f$, attenuates faster for high-frequency waves (wind waves) than for low-frequency waves (swell). The spatial decay coefficient $\alpha$ (right panel) describes wave attenuation and is related to $a$ through

$$\frac{\partial a}{\partial x} = -\alpha a,$$

where $x$ is the wave traveling distance (WTD) through the MIZ. The spatial decay coefficient increases with increasing frequencies. Løken et al., 2021a found from field observations that $\alpha \propto f^{3.51}$, while the two-layer attenuation model of Sutherland et al., 2019, which is also presented in the right panel of Fig. 1.2, concludes that $\alpha \propto f^4$.

Figure 1.2: Exponential wave attenuation through the MIZ (Løken et al., 2021a). Spectral amplitudes vs wave traveling distance (WTD) into the ice (left panel) and spatial decay coefficients vs wave frequency (right panel).

Although the spatial damping coefficient is generally expected to increase with increasing frequencies, high frequency waves may be generated deep into the MIZ through nonlinear energy transfer from the long swells via ice floes to shorter waves due to floe motion and interactions (Wadhams et al., 1988). This phenomenon is called rollover effect and has been reported to result in a flattening of attenuation rate at some frequency and even a decreasing attenuation rate at higher frequencies (Squire, 2018). Figure 1.3 shows an example of surface wave generation due to a moving ice floe investigated by Løken, Marchenko, Ellevold, et al., 2021. The floe motion in surge mode (translation in the axial direction) generated high frequency surface waves associated with floe response in roll and pitch (rotation around the axial and transverse axes, respectively). According to Wadhams et al., 1988, rollover can also be caused by wind forcing in the MIZ where surface waves are regenerated. However, this phenomenon has been debated and Thomson et al., 2021 suggested that rollover may come from artifacts in instrumentation.
1. Introduction

Figure 1.3: Generation of surface waves due to water-ice and ice-ice interactions. The left panel shows an ice floe that was towed back and forth in a pool in the fast ice to simulate wave motion. The right panel shows the floe response in surge (period around 26 s), roll and pitch (period around 2 s). The two figures illustrate that low-frequency surge motion probably induces high-frequency surface waves in the pool.

Despite the disagreements regarding the rollover effect, the general consensus in the society is that waves are damped by the presence of ice. Several phenomena are known to attenuate waves in an ice floe field, such as wave scattering or directional spreading and viscous dissipation in the boundary layer beneath the ice due to shear flow or wake formation caused by a relative velocity between the water and the ice (Liu & Mollo-Christensen, 1988; Wadhams, 1975). However, there is uncertainty associated with the dominating source of wave energy dissipation by sea ice, which depends on the ice concentration and floe size distribution, linked to the incoming ocean waves (Squire, 2018). Detailed knowledge about these processes is required to be able to predict wave properties inside the MIZ, such as wave height, which is an important parameter for human activities. Interactions between waves and ice also contribute to a considerable atmosphere-ice-ocean energy transfer, which is important for understanding and predicting global climate processes.

Scattering, which does not dissipate energy, contributes to wave decay due to energy reflection and spreading, and is known to be of importance in open fields of floes (Squire et al., 1995). Ice floe interactions can lead to wave energy dissipation through different mechanisms and are of relevance in denser fields. Collisions between neighboring ice floes can for example cause momentum transfer and energy absorption (Li & Lubbad, 2018; Shen & Squire, 1998). Floes may also be dragged past each other so that kinetic energy is dissipated at the rough surface. Rabault et al., 2019 showed from wave tank experiments
Wave-ice interactions

Figure 1.4: Eddy injection in the water due to water-ice interactions. Left panel: Wave tank experiment where patches of grease ice collide under the influence of waves, resulting in a large energetic eddy (Rabault et al., 2019). The figure shows a Proper Orthogonal Decomposition (POD) analysis applied on particle image velocimetry data. Right panel: Field experiments where an ice floe collides with fast ice due to simulated wave motion (Løken, Marchenko, Ellevold, et al., 2021).

that colliding chunks of grease ice can generate turbulence that injects eddy viscosity in the water, which leads to enhanced energy dissipation. Examples of eddy generation in the water due to ice-ice interactions are presented in Fig. 1.4. Voermans et al., 2019 measured under-ice turbulence in pancake and frazil ice generated from the relative velocity between the ice and the orbital wave motion, and suggested that turbulence dissipation caused wave attenuation. According to Herman, 2018, it is important to gain knowledge into the details of individual dissipation processes, as this insight is a prerequisite to understanding interactions with other components of the complex system that is the MIZ.

Numerical models seek to explain the different dissipation processes mathematically. Early theoretical descriptions of wave propagation in ice covered seas modeled the ice as a continuous, solid cover that behaves as a thin elastic plate (Liu & Mollo-Christensen, 1988; Squire, 1993; Wadhams, 1973). The core concept of the modeling is to describe wave propagation with potential flow theory (e.g. Newman, 2018), but with modified boundary condition at the surface to account for the elastic effect of the ice, in order to investigate the effect of the ice on the dispersion relation. Several wave damping models build upon this principle, with the addition of an effective viscosity in the water or the ice layer where energy dissipation takes place. The mathematical descriptions grow in complexity from the one-layer (Newyear & Martin, 1997; Weber, 1987) to the two-layer models (De Carolis & Desiderio, 2002; Sutherland et al., 2019; Wang & Shen, 2010).

For the models to agree with observations, the effective eddy viscosity, which is considered a fitting parameter, must be much higher than the molecular water viscosity (Rabault et al., 2017). A drawback with the more recent sophisticated
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models is that they contain several parameters, such as ice thickness, ratio between the thickness of the layers and effective viscosity in the second layer, which can be practically difficult to determine in many situations, particularly in the heterogeneous MIZ. These parameters are usually adapted for the models to fit laboratory or field measurements, meaning that the models do not necessarily describe the underlying mechanisms for wave attenuation, but the good agreement with experimental data could be a mathematical artifact due to more available fitting parameters (Mosig et al., 2015; Rabault et al., 2017).

1.3 Research questions and methodology

Due to the uncertainties associated with the existing wave attenuation models mentioned in the previous section, there is a need for direct observations of the mechanisms in play when wave energy dissipates in the interactions with sea ice. More field measurements are required to validate both mathematical models and remote sensing, such as satellite or airborne altimeter data. The harsh conditions and inaccessibility of the Arctic makes it a challenging place to work in situ (Chang & Bonnet, 2010). Consequently, there are relatively few field experiments on wave propagation in ice, especially considering the vast extension of the polar regions with the rich diversity of ice types and wave conditions (Herman, 2018; Squire, 2007). The development and validation of various field measurement techniques adapted to Arctic conditions has therefore been the primary focus of this thesis. Results obtained from the measurement techniques have been compared with wave attenuation models for validation of the models. Different wave damping mechanisms due to ice have been investigated from the analyzed field data.

During the DOFI (dynamics of floating ice) project, which this thesis has been a part of, two experimental techniques have been introduced to the wave-ice scientific community. The first methodology is a ship-mounted wave measurement device presented in Løken et al., 2021a, which consists of an ultrasonic range meter and an inertial motion unit (IMU) to correct for ship motion, both mounted in the bow of a research vessel. The setup was tested in the MIZ and validated against wave data from special designed IMUs (Rabault et al., 2020) placed on ice floes. Figure 1.5 shows the instrument mounted on R/V Kronprins Haakon during a cruise in the Barents Sea. A great advantage of the ship mounted instrument is that it can gather data continuously and autonomously during expeditions, without dedicating expensive work time or equipment to perform the measurements. Direct observations of waves in ice is much needed to validate models and remote sensing, and this methodology has a potential to increase the amount of accumulated field data available for researchers. Wave parameters obtained from the instrument were used as a comparison with different spectral wave hindcasts in Løken et al., 2020, to evaluate the performance of the different wave propagation models in the MIZ.
The second measurement technique is a particle image velocimetry (PIV) setup designed for field use under Arctic conditions, which was presented and validated in Løken et al., 2021b. It consists of a BlueROV2 remotely operated vehicle (ROV) that is maneuvered below the ice to film tracing particles (BlueRobotics, 2020). The group tested different kinds of passive and environmental-friendly tracers, such as dye and conifer pollen, but learned that it was challenging to obtain a uniform concentration of particles due to the residual ocean current. Therefore, bubbles were used as tracers, and the methodology is referred to as bubble-PIV. Bubbles are advantageous in the sense that they can approximately rise in a 2D plane, which makes a light or laser sheet redundant, but it is of course troublesome to correct for the vertical buoyant bubble velocity. The approach was to perform PIV on the images and remove the vertical bubble motion in the post-processing to obtain the 2D velocity field of the water. The methodology offers the possibility of investigating water kinematics around water-ice and ice-ice interactions with a temporal and spatial resolution that extends beyond conventional instruments, such as ADCPs (acoustic Doppler current profilers). The resulting data may provide insight into important physical mechanisms in the MIZ. The instrumental setup was applied in Løken, Marchenko, Ellevold, et al., 2021 (see Fig. 1.6), where the 2D velocity field of a water jet resulting from ice floe collisions was obtained.

A central working hypothesis of the DOFI project group was that a significant part of the attenuated wave energy in the interactions with ice is dissipated in
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Figure 1.6: Instrumental setup of the bubble-PIV system for field measurements of water kinematics around floating ice. The two panels show two different arrangements for bubble generation, with smaller bubble diameter (1-2 mm) in the left panel and larger bubble diameter (2-4 mm) in the right panel.

turbulence (suggested by e.g. Voermans et al., 2019 and Rabault et al., 2019). This process has been observed directly with the bubble-PIV system described above. In addition, a five beam Nortek Signature1000 ADCP has been used to measure turbulence properties. The instrument was operated in the pulse coherent mode (high resolution mode) that enables cell size down to 2 cm on all beams and a sampling frequency of 8 Hz when all the beams are operated simultaneously. These properties makes the ADCP suitable for turbulence measurements on a much smaller scale than what is found for example in a tidal channel. As relatively new technology is incorporated in the ADCP, it was decided to test its ability to resolve small-scale turbulence under controlled conditions in a towing tank. These experiments are described in Section 1.3.1 and in Løken, Jensen, et al., 2021. The ADCP was also used to measure turbulence dissipation around colliding ice floes in Løken, Marchenko, Ellevold, et al., 2021.

Iceberg 3D dynamics and stability have also been investigated over the course of the thesis. The stability of icebergs defines the safety of the humans and the equipment involved in iceberg operations, such as towing. Previous studies on the topic have mainly used GPS trackers and accelerometers to monitor iceberg dynamics. The DOFI project group went one step further and installed IMUs on the icebergs to obtain information on the six rigid body motion mode response and hydrodynamical stability during drift and towing. Wave parameters were also measured to investigate iceberg response on incoming waves and the additional wave load induced on the iceberg-towline-boat system.
The work is summarized in Løken, Marchenko, Gundersen, et al., 2021. It was found that significant low-frequency iceberg oscillations occurred under weak forcing, which indicates that the restoring forces acting on the iceberg are weak compared with its inertia and that the metacentric height is low. Consequently, the risk of iceberg tipping is high.

1.3.1 Laboratory investigations

The goal of the DOFI project is to describe full-scale phenomena associated with wave-ice interaction in the nature through *in situ* observations. However, laboratory investigations offer some advantages over field work and have therefore been used in some parts of the thesis to validate the introduced methodologies. First and foremost, it is easier to deploy advanced measurement techniques such as particle tracking velocimetry (PTV) in a laboratory in comparison to the Arctic MIZ. A laboratory also offers the possibility of isolating (to some extent) the physical phenomena of interest and reducing sources of error. For example, if the parameter of interest is the water velocity due to surface waves, it is beneficial to avoid the disturbance of tidal current or wind input. In addition, laboratory experiments are performed in a standardized fashion that increase the reproducibility, as opposed to field work where the ice, wave, wind and current conditions most likely vary within a couple of hours.

Two facilities in the Hydrodynamics Laboratory at the University of Oslo were used to investigate the accuracy of the bubble-PIV system presented in Løken et al., 2021b:

- A small glass tank measuring approximately $1.4 \times 0.4 \times 0.3$ m to investigate the bubble rise motion in stagnant water (fresh and salt). Images from a high-speed camera were processed with PTV to obtain a relation between the bubble diameter and terminal velocity. This is an important property as the removal of the vertical buoyant velocity is essential for the bubble-PIV technique. PTV, which uses a Lagrangian frame of reference, was used to track individual bubbles as opposed to PIV, which uses an Eulerian frame of reference.

- A medium size wave tank measuring $24.6 \times 0.5 \times 0.6$ m to investigate the bubbles capability of following the water motion. Waves were generated with a computer-controlled piston type paddle and the wave height was measured with ULS ultrasonic gauges (UG). A comparison of the horizontal bubble velocity obtained from PTV, and the theoretical water velocity under the crests of monochromatic surface waves was performed. It was beneficial to look at the wave crests indicated in Fig 1.7, where the vertical water velocity, which was estimated with third order Stokes waves (Newman, 2018), is zero.
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MarinLab towing tank at the Western Norway University of Applied Sciences in Bergen was used to validate the ability of the ADCP to resolve small scale turbulence properties, which were generated by towing a regular grid through the $50 \times 3 \times 2$ m tank. The motivation for this study was to investigate the ADCP limitations under controlled conditions before applying it in the field to measure turbulence around floating ice, which was done in Løken, Marchenko, Ellevold, et al., 2021. However, only the larger turbulent scales were resolved by the instrument and the results were not published. Hopefully, the group will be able to go back to MarinLab or another towing tank that allows for the generation of larger structures in the future, so that the experiments may be completed with the experience gained from the first attempt. The work is presented in Paper V.

1.3.2 Fieldwork

Most of the data this thesis is built upon are obtained from field expeditions on and around Svalbard, which are summarized in Table 1.1. As mentioned earlier, direct observations are essential for describing in detail the phenomena occurring in the process of wave-ice interactions, and are required for the validation of numerical models and the calibration of remote sensing techniques. Due to the harsh environment and the remote locations of the Arctic, there is a limited number of in situ observations available in the literature, and more studies are much needed. Also, scaling problems, for example to ensure simultaneous deep-water waves, moderate wave steepness and realistic Reynolds number, are challenging in a wave tank, as pointed out by Rabault et al., 2019. Field observations are therefore necessary to compliment or confirm laboratory experiments.
Research questions and methodology

<table>
<thead>
<tr>
<th>Expedition</th>
<th>Date</th>
<th>Location</th>
<th>Instruments</th>
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<tr>
<td>1</td>
<td>Sept., 2018</td>
<td>Barents Sea</td>
<td>UG, IMU, an.</td>
</tr>
<tr>
<td>2</td>
<td>March, 2019</td>
<td>Longyearbyen</td>
<td>B-PIV, ADCP</td>
</tr>
<tr>
<td>3</td>
<td>April, 2019</td>
<td>Hopen</td>
<td>B-PIV</td>
</tr>
<tr>
<td>4</td>
<td>March, 2020</td>
<td>Svea Bay</td>
<td>B-PIV, ADCP, RM, IMU, LC</td>
</tr>
<tr>
<td>5</td>
<td>Sept., 2020</td>
<td>Tempel Fjord</td>
<td>IMU, LC, ADCP, an.</td>
</tr>
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Table 1.1: Field expeditions over the course of the thesis. The instrument abbreviations are as follows: UG - ultrasonic gauge; IMU - inertial motion unit; an. - anemometer (wind speed and direction); B-PIV - bubble particle image velocimetry; ADCP - acoustic Doppler current profiler; RM - range meter; LC - load cell.

The field expeditions listed in Table 1.1 can be summarized as follows:

1. The first expedition was a cruise in the Barents Sea MIZ with R/V Kronprins Haakon as a part of the Nansen Legacy project (Reigstad et al., 2017), where ship mounted wave measurements were performed. The measurement system was designed and deployed by DOFI project member Jean Rabault, MET. The author was not participating on the expedition but was responsible for the data analysis which was presented in Løken et al., 2021a.

2. This was intended to be a field campaign to the ice-covered Temple Fjord to test the bubble-PIV system. Unfortunately, polar bear and avalanche danger prevented the group from going far outside of Longyearbyen. Therefore, the bubble-PIV system and the ADCP was tested in the Longyearbyen harbor. The work did not result in any publications, but the group gained valuable experience with the instruments and were able to make improvements until the next expedition.

3. Shortly after the testing of the bubble-PIV system in Longyearbyen, a cruise to the vicinity of the Hopen Island with the M/S Polarsyssel was organized by DOFI project member Aleksey Marchenko, UNiS. The bubble-PIV system was deployed by DOFI project member Thea Josefine Ellevold, UiO, who managed to measure the 2D water velocity field next to an ice floe. The author did not participate on the expedition but incorporated the results in Løken et al., 2021b.

4. The next expedition was a snow mobile field campaign to the fast ice covered Svea Bay, arranged by Aleksey Marchenko. Towing experiments with an ice floe were carried out to investigate mechanisms of energy...
dissipation. The water kinematics around the floe were measured with both the bubble-PIV system and the ADCP and presented in Løken, Marchenko, Ellevold, et al., 2021. The towing load was measured with a load cell and the floe motion was measured with a range meter and an IMU.

5. The last fieldwork was a boat campaign to the Tempel Fjord, which contained many small icebergs at the time. Towing experiments with small icebergs were performed to investigate their stability during towing and the effect of waves on the iceberg-towline-boat system (Løken, Marchenko, Gundersen, et al., 2021). The applied towing setup, shown in Fig. 1.8, consisted of a partly submerged towline to reduce the overturning moment and the risk of towline slippage. The iceberg and boat motion, as well as the ocean surface elevation, were measured with IMUs and the iceberg subsurface geometry was investigated with the ROV. The author applied for and was granted funding for the expedition from The Research Council of Norway under the Arctic Field Grant (project "Investigation of iceberg dynamics in Isfjorden", Grant number 310691), and The Norwegian Society of Graduate Technical and Scientific Professionals under the Harald Boes Grant.

A collaboration with Malin Johansson in the Earth Observation group at The Arctic University of Norway (UiT) was initiated prior to the fifth expedition. Their group provided satellite images of the field site and our group provided photos and GPS coordinates of various icebergs that they were interested in recognizing on the satellite images, which is an example of the use of in situ
observations for validation of remote sensing.

The data from all the experiments (including the laboratory work) were processed in Matlab and Python. PTV was performed with DigiFlow software (Dalziel, 2006) and PIV with the in-house HydrolabPIV software developed at the University of Oslo (Kolaas, 2016).

1.4 Summary

Wave-ice interactions define the polar areas and are of importance for the global climate because of atmosphere-ocean energy transfer. The decrease in the Arctic sea ice cover that has been observed over the last couple of decades has allowed for more human activities in the region. Accurate wave forecasts and knowledge about ice dynamics are necessities to ensure safe human operations in the Arctic. Numerical forecast and climate models seek to incorporate wave damping processes in ice-covered areas, but there are uncertainties associated with the dominating mechanisms of wave attenuation. More direct observations of wave-ice interactions are required to develop and validate wave damping models and remote sensing used for predictions of wave parameters in ice-covered oceans. The works included in this thesis therefore present new observations and techniques to measure the physical mechanisms related to wave attenuation through ice. Error analysis, on several occasions obtained from laboratory investigations, have been included to validate the introduced methodologies. The results obtained from several field expeditions are compared with numerical models to investigate their accuracy.

In addition to wave propagation through continuous ice covers or fields of ice floes, drifting icebergs due to forcing from waves, current and wind are defining aspects of the polar regions. Icebergs pose a direct threat to ships and constructions because they can cause damage in the event of a collision. Drift models and satellite surveillance of icebergs can aid sailors in navigational tasks. However, iceberg towing may be required if an iceberg approaches a platform or drilling vessel. Icebergs usually have small aspect ratios and they may roll as a consequence of the external forcing applied during such an operation, which could cause harm to the crew and equipment. Iceberg stability is poorly understood because few studies have analyzed the dynamics of icebergs during towing. Therefore, a part of the project objective was to investigate iceberg hydrodynamic stability and the effect of towing in a wave field. This was obtained by an extensive instrumentation in the experimental setup, where the icebergs’ six rigid body motion mode response, the towing force and the wave conditions were measured.
1. Introduction

References


1. Introduction


Chapter 2

Summary of papers

This chapter presents an overview of the studies that were published or submitted to journals and conferences during the course of the thesis, and also unpublished work. For each paper, the main findings are outlined. The papers are put into context and an explanation of how they fit together follows to help the reader understand the combined results. Some reflections around how the studies could be improved in the future are included at the end of the chapter.

2.1 Papers included in the thesis

As mentioned in Chapter 1, the overall objective of this thesis is to improve our understanding of wave damping due to the presence of sea ice through an experimental approach. Hence, an indirect objective is to develop and validate field measurement techniques adapted to Arctic conditions and apply these to make direct observations of attenuation mechanisms caused by wave-ice interactions.

Paper I: Wave measurements from ship mounted sensors in the Arctic marginal ice zone

presents a ship-mounted wave measurement technique and data from a field campaign into the marginal ice zone (MIZ). The main objective of the paper was to test and validate the presented methodology against waves-in-ice (WII) sensors (Rabault et al., 2020) and the spectral wave forecast and hindcast models WAM-4 (Saetra & Bidlot, 2004) and ERA5 (Janssen & Bidlot, 2002), respectively. Figure 2.1 presents time series of wave parameters from the observations and the models when the ship was located at the open ocean. In terms of wave height and period, the measured values deviated most of the time less than 20% from the models.

Inside the MIZ, wave spectra from the ship measurements agreed with spectra from the independent in-situ WII sensors (shown in Fig. 2.2), which illustrates a satisfactory performance of the new methodology in ice-covered seas. The wave attenuation coefficient estimated from the observations was found to be proportional to the wave frequency $f$ raised to the power of 3.51. In comparison, the theoretical two-layer ice model of Sutherland et al., 2019 is proportional to $f^4$, which implies that the model predicts a slightly higher increase in attenuation with increasing wave frequency compared to the observations. The authors released the
Figure 2.1: Long time comparison between observations (bow) and the models WAM-4 and ERA5. Significant wave height from time series and spectra (upper), and periods from spectra (lower). Time periods where total instrument saturation was below 10% and ship speed over ground was below 0.5 m/s are highlighted with green background color and the period with an ice cover (hence no model data) is highlighted with red background color.

electronics and code as open source in the hope that the technique can be used by other researchers to increase the collective amount of much needed wave measurements in ice-covered regions.

**Paper II: A comparison of wave observations in the Arctic marginal ice zone with spectral models**

is a continuation of Paper I. A central function of wave observations, besides investigation of physical processes, is to validate mathematical and numerical models so that they can be improved. The models are under continuous development and in-situ data from the MIZ offer a rear opportunity to evaluate the performance of the algorithms that describe wave propagation through sea ice. In this paper, the wave data obtained within the MIZ in Paper I were compared with the spectral hindcast models WAM-3 and WW3, which both estimate wave attenuation in ice covered seas. Spectra and peak frequencies are shown in Fig. 2.2. The ship-mounted and WII sensors show a similar decay in wave energy with increasing ice concentration (IC, from upper to lower panels), while the spectra from WW3 barely decrease. The peak frequency estimated with WAM-3, decrease with increasing IC in accordance with the observations.
Figure 2.2: Power spectra from ship (bow) measurements (black) and WII instruments (blue) with their respective 95% confidence intervals shaded. Spectra from WW3 (magenta) and peak frequencies from WAM-3 (green) with increasing ice concentration from upper to lower panels.

The IC2 setting in WW3 was applied, which models the sea ice as a continuous thin elastic plate, where the waves are damped through dissipation caused by bottom friction below the plate (Ardhuin et al., 2015). It was found that the WW3 model underestimated wave attenuation due to the ice by approximately 75% with respect to the observations, when the spatial attenuation coefficient was estimated from the significant wave height. The WAM-3 model on the other hand, which was configured with a two-layer ice parameterization (Sutherland et al., 2019), was in better agreement with the measured data when considering the estimated wave attenuation coefficient. The results suggest that WAM-3 performs better than WW3 under certain conditions (ice floe diameter $\sim 10-100$ m and ice concentration $10-100\%$) with the applied model parameterization.

**Paper III: Bringing optical fluid motion analysis to the field: a methodology using an open source ROV as camera system and rising bubbles as tracers**

is to a large extent a methodology paper that presents a new field measurement technique. 2D velocity fields in the water, e.g. around ice boundaries, can be obtained with the introduced method. The objective of the paper was to develop a tool to investigate water kinematics related to mechanisms of wave attenuation in the MIZ. An example of a velocity field obtained
Figure 2.3: Horizontal, non-dimensional velocity profile divided into 10 equally spaced vertical bins below 1.8 Hz wave crests. Blue dots: observed horizontal bubble velocity $u_b$. Green error bars: mean and $2\sigma$ uncertainty of the observed $u_b$ within each bin. Red line: theoretical water velocity $u_w$. Orange shaded region: expected $u_b$ from estimated bubble size.

Next to a heaving ice floe in the Barents Sea is included to illustrate how the methodology can be used in field measurements. It was originally intended to use the results from laboratory tests to compensate for the buoyancy driven vertical bubble velocity in the field experiments. However, it was found that an on-site calibration was necessary because the different conditions in the field (temperature and pressure) substantially altered the bubble properties.

Another central part of the study was the analysis of bubble dynamics and response in a wave field, which is of importance in the emerging field of measurement techniques that utilize bubbles as tracers. Figure 2.3 shows the tracked bubble velocities and the theoretical water velocity from third order Stokes waves (Newman, 2018) in the horizontal direction below a wave crest. Laboratory experiments showed that the presented methodology was able to reproduce the water velocity with an accuracy in the order of 10%. The deviation was mainly attributed to the slip velocity caused by the bubble inertia due to the added mass effect, and theoretical estimates corresponded well with the observations, as seen from the orange shaded region in Fig. 2.3. The spread in bubble size distribution and secondary bubble motion probably contributed to the data scattering.

**Paper IV: Iceberg stability during towing in a wave field**

is the only paper related to icebergs and iceberg towing operations. The objectives of the study were to develop a towing configuration that would reduce the risk of iceberg tipping and towline slippage, and to apply motion sensors to investigate the hydrodynamic stability and iceberg response to
waves during towing. Figure 2.4 shows a sketch of the applied towing setup, which reduced the overturning moment due to the submerged towline compared with the conventional floating towline.

Transitions between stable static positions were observed in the iceberg roll data, and in some situations, this caused tipping events. The towing load showed oscillations with period around 6 s, which were attributed to either the towline elastic properties or the iceberg heave motion. These oscillations were found to increase in amplitude when the towing was performed parallel to the principal wave direction and could be important to consider in a full-scale operation due to the increased peak load, which may lead to towline rupture.

**Paper V: Grid turbulence measurements with an acoustic Doppler current profiler**

Presents results from towing tank experiments where grid-induced turbulence was measured with a Nortek Signature1000 (kHz) acoustic Doppler current profiler (ADCP) and an acoustic Doppler velocimeter (ADV) for validation. The motivation for the study was to investigate the abilities and limitations of this specific ADCP regarding fine-scale turbulence measurements. The DOFI (dynamics of floating ice) research group utilized the instrument for turbulence measurements around water-ice interactions in Paper VI. Therefore, the ADCP was tested in a laboratory facility, where turbulence was generated from grid towing under controlled conditions.

It was found that the ADCP resolved turbulent kinetic energy (TKE) reasonably well when energetic and large turbulent eddies were generated, i.e. the largest grid towed at a high velocity. However, TKE was overestimated by a factor of 2-5 by the ADCP compared with the ADV for less energetic flow, probably due to instrument noise. Figure 2.5 shows the decay of TKE as function of downstream distance. The observed decay was similar to the experimental results of Sirivat and Warhaft, 1983 and Liu, 1995 (dashed line). The TKE spectra estimated from the ADCP showed that
only the largest eddies were detected by the ADCP. The inertial subrange, which was clearly visible from the ADV data, was not visible in the ADCP spectra due to the intrinsic Doppler noise and the insufficient temporal and spatial instrument resolution.

**Paper VI: Turbulent kinetic energy dissipation from colliding ice floes** contains elements from several of the previous papers and it naturally concludes the thesis. The objective of the study was to investigate and directly observe mechanisms that contribute to attenuation of waves in the interaction with ice. This was obtained with the bubble tracers presented in Paper III, the motion sensors applied in Paper I and IV and the ADCP that was tested on grid-generated turbulence in Paper V. A full-scale ice floe was towed back and forth in a pool by electrical winches to simulate wave motion. The extensive instrumentation that was installed on and around the colliding floe allowed for comparison of the energy input rate with dissipation rates from various processes related to energy loss.

The rate of kinetic energy absorbed in the collisions between the ice floe and the surrounding ice walls was found to be 7.5% of the total input energy rate from the electrical winches. The bubble tracers visualized the jets that were generated in the collision events, which improved the authors’ conceptual understanding of the different turbulent regimes. Figure 2.6 presents profiles of TKE dissipation rates, decaying with depth, that were
Papers included in the thesis

Figure 2.6: Estimated TKE dissipation rate $\epsilon$ profiles along the depth $z$. a) ADCP placed 0.5 m from the pool edge. b) ADCP placed 0.25 m from the pool edge and on the ice floe center (brown). Dashed lines show curve fits to the ADCP data. Confidence intervals are indicated with shaded regions. The inset plots show the vertical beam correlation data for the ADCP profiles (percentage of time series with correlation < 50%). ADV data are presented as large dots. Estimated TKE dissipated in the water was estimated to be more than 1/3 of the total input energy rate, which indicates that a substantial part of the wave energy is dissipated in turbulence during interaction with sea ice.

All the papers seek to investigate physical aspects of wave-ice interactions or mechanisms related to these phenomena. The dynamics of floating ice is the main topic of Papers I, II, IV and VI, and wave attenuation due to floating ice is described in Papers I, II and VI. Instruments or measurement techniques adapted to Arctic field conditions have been developed and validated in Papers I and III, and these methodologies or data obtained from them have been used to analyze mechanical features of ice or the surrounding water in Papers II and VI, respectively. Although Paper V is not directly focused on waves in ice, turbulence and TKE dissipation could be an important contribution to wave attenuation, as shown in Paper VI. The experience gained in Paper V related to fine-scale turbulence measurements with an ADCP proved beneficial in the field experiments presented in Paper VI. While most of the studies can be considered fundamental research where the knowledge outcome may contribute to improve climate or wave models (as shown in Paper II), Paper IV, which
focus on iceberg towing and stability in a wave field, has a more direct approach to human activities in the polar regions.

2.2 Relevant paper not included

In addition to the selected works mentioned above, one publication written during the project is not included in this thesis. This is a conference paper on turbulence around ice floes (Løken et al., 2021), which has been extended and completed by Paper VI.

2.3 Future perspectives

Several adjustments could be made to improve the experimental setups and techniques presented in this thesis. In the following, some ideas and suggestions for future development are outlined.

The shipborne wave measurement technique presented in Paper I could be arranged such that the range meter is mounted on a hinged pole in order to always keep the instrument axis approximately vertical, even when the ship is rolling and pitching. Such an arrangement would reduce the error introduced from the roll and pitch angles (although the introduced error is mitigated in the post-processing). The hinge would probably have to consist of a combination of spring and damping component to reduce system oscillations. Another possibility is to use a laser instead of an ultrasonic range meter, which may reduce the noise level in the signal and at least reduce the measurement footprint on the ocean surface. Some of the DOFI project members are currently improving the system for a near future expedition, but the author is not directly involved in this work.

In the field application for particle image velocimetry with bubbles as tracers presented in Paper III, the main challenge is to generate bubbles of uniform size. This is advantageous because the bubble rise velocity, which depends on the bubble diameter, needs to be subtracted from the measured vertical velocity to obtain the water velocity. It is also desirable to work with small bubbles, because these have a rectilinear rising path (Haberman & Morton, 1953). These two criteria may be possible to achieve with electrolysis in water, where preliminary tests carried out by the DOFI project group have shown that the generated hydrogen bubbles are small and relatively uniform in size. Further investigations are required to confirm this, but it would be interesting to test a setup with electrolysis in a wave tank in the future, to see if it is possible to obtain better results than the ones presented in Paper III.

During the iceberg towing experiments in Paper IV, the iceberg rolled around, and the motion sensors attached to it by means of ice screws ended up under water and was not retrieved. The DOFI project group is planning to carry out new iceberg towing experiments in the future, where this situation should be
avoided. The sensors ought to be attached to the iceberg with a mechanism that allows for remote releasing. For example, there exist ice screws for ice climbing applications that can be unwound with a thin rope from a distance. Another possibility is to freeze the instruments to the iceberg, and if the iceberg rolls over, the central topic of ice will thaw relative quickly in contact with sea water and the instruments will surface due to the buoyant casings. This method, which of course requires air temperatures below freezing, was used to attach accelerometers to the ice floe in Paper VI.

The turbulent kinetic energy dissipation around a colliding ice floe was estimated from measurements with an acoustic Doppler current profiler in Paper VI. The instrument was only deployed on three locations due to time limitations in the field, although the pool in which the floe moved was 6×4 m. If the experiment is to be repeated in the future, it would be beneficial to deploy the instrument on several locations around the pool and on the ice floe to improve the spatial resolution of the estimates.

References


2. Summary of papers

Wave measurements from ship mounted sensors in the Arctic marginal ice zone

Løken, T.K., Rabault, J., Jensen, A., Sutherland, G., Christensen, K.H., and Müller, M.

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WAVE MEASUREMENTS FROM SHIP MOUNTED SENSORS IN THE ARCTIC MARGINAL ICE ZONE

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May 5, 2021

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ABSTRACT

This study presents wave measurements in the marginal ice zone (MIZ) obtained from ship mounted sensors. The system combines altimeter readings from the ship bow with ship motion correction data to provide estimated single point ocean surface elevation. Significant wave height and mean wave period, as well as one-dimensional wave spectra are derived from the combined measurements. The results are compared with integrated parameters from two spectral wave models over a period of eight days in the open ocean, and with spectra and integrated parameters derived from motion detecting instruments placed on ice floes inside the MIZ. Mean absolute errors of the integrated parameters are in the range 13.4-29.9% when comparing with the spectral wave models and 1.0-9.6% when comparing with valid motion detecting instruments. The spatial wave damping coefficient is estimated by looking at the change in spectral wave amplitude found at discrete frequency values as the ship was moving along the longitudinal direction of the MIZ within time intervals where the wave field is found to be approximately constant in time. As expected from theory, high frequency waves are effectively dampened by the presence of sea ice. The observed wave attenuation rates compare favourably with a two-layer dissipation model. Our methodology can be regarded as a simple and reliable way to collect more waves-in-ice data as it can be easily added to any ship participating to ice expeditions, at little extra cost.

Keywords: Sea ice dynamics, wave measurements, marginal ice zone, wave attenuation

1 Introduction

Sea ice is a major feature in the polar environments, it has importance for Arctic ecosystems as well as for global ocean and atmospheric circulation. A decline in the Arctic ice cover has been observed over the past decades (Feltham 2015). Interactions between sea ice and surface gravity waves play an important role in breakup and reduction of ice cover. The decline in ice cover leads to a larger fetch where waves build up more energy and enhance the breakup and melting process in a positive feedback mechanism (Thomson & Rogers 2014). The mixture of icebergs, floes and grease ice found in the interface between solid ice, such as land fast ice or pack ice, and the open ocean, is called the marginal ice zone (MIZ). Previous studies have found that high frequency wind waves are effectively dampened in the MIZ (Weber 1987, Wadhams et al. 1988), which reduces the ice cover break up rate.

Recent changing conditions in the polar regions have allowed for increased human activities, which raises the importance of better forecast models and improved physical understanding of the environment to ensure safe operations (Fritzner et al. 2019). In situ wave measurements can increase our understanding of global climate systems and provide data

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for calibration of numerical models. Also, mathematical models to describe wave attenuation in ice, e.g. Weber (1987), Newyear & Martin (1997) and Sutherland et al. (2019), need experimental data for validation and improvement. However, experimental data are relatively sparse due to the inaccessibility of the regions where sea ice is present, combined with the harsh and dangerous environment for both researchers and instruments (Squire 2007).

We present here results from shipborne wave measurements in the MIZ. This system has not up until now been used deep into the MIZ, therefore, the novelty of our observations. The methodology, first described in Christensen et al. (2013), combines a bow mounted altimeter and a motion correction device. We provide wave spectra and integrated parameters from spectra, which are important quantities when considering wave-ice interactions. Significant wave height and mean periods are compared with a spectral wave forecast model over a period of eight days in the open ocean. We also compare measurements in the MIZ with data from wave measuring instruments consisting of inertial motion units (IMUs) placed on ice floes (Rabault et al. 2020). From the spectra, the spatial damping coefficient is found as a function of wave frequency, which can be compared to attenuation models.

In this paper, the data acquisition and processing methods are described in Section 2. The results are presented in Section 3, which is divided in two parts. A comparison with spectral models and in situ instruments is outlined in Section 3.1. In Section 3.2, we present results on wave attenuation and the spatial damping coefficient is compared to the theoretical model of (Sutherland et al. 2019). Finally, a discussion follows in Section 4 and the concluding remarks are given in Section 5.

## 2 Data and Methods

The data were obtained during a research campaign in the Barents Sea with R/V Kronprins Haakon in September 2018 as part of the Nansen Legacy project (Reigstad et al. 2017). The vessel is 100 m long and with a beam of 21 m. In total, the cruise lasted two weeks, during which continuous measurements were made. The results from the MIZ presented here were recorded on September 19 when the ship ventured approximately 28 km into the MIZ. Four stops were made to deploy in situ waves-in-ice (WII) instruments on ice floes (Rabault et al. 2020), which were used to validate our ship mounted system. Upon the return to the open ocean, a total of seven stops were made with more or less equal spacing on a close to straight south-southwest heading where measurements for wave damping estimates were carried out. Twenty minutes samples were recorded as the ship was freely drifting on each station. The location, starting time and wave travel distance (WTD) for each measurement in the MIZ are summarized in Table 1. Prefix 1 denotes the four stops into the MIZ while prefix 2 denotes the seven stops out of the MIZ.

<table>
<thead>
<tr>
<th>Stop</th>
<th>Time</th>
<th>Position (N/E)</th>
<th>WTD [km]</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>04:22</td>
<td>82.126/20.736</td>
<td></td>
</tr>
<tr>
<td>1.2</td>
<td>06:28</td>
<td>82.246/20.245</td>
<td></td>
</tr>
<tr>
<td>1.3</td>
<td>08:59</td>
<td>82.355/19.803</td>
<td></td>
</tr>
<tr>
<td>1.4</td>
<td>12:20</td>
<td>82.436/19.674</td>
<td></td>
</tr>
<tr>
<td>2.1</td>
<td>13:11</td>
<td>82.421/19.579</td>
<td>92.7</td>
</tr>
<tr>
<td>2.2</td>
<td>14:32</td>
<td>82.359/19.544</td>
<td>69.2</td>
</tr>
<tr>
<td>2.3</td>
<td>15:40</td>
<td>82.294/19.389</td>
<td>58.3</td>
</tr>
<tr>
<td>2.4</td>
<td>16:54</td>
<td>82.228/19.275</td>
<td>20.8</td>
</tr>
<tr>
<td>2.5</td>
<td>18:00</td>
<td>82.163/19.183</td>
<td>6.7</td>
</tr>
<tr>
<td>2.6</td>
<td>19:08</td>
<td>82.099/19.046</td>
<td>0</td>
</tr>
<tr>
<td>2.7</td>
<td>20:09</td>
<td>81.994/18.982</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 1: Time and location where measurements were carried out inside the MIZ. Prefix 1 for stop number indicate deployment of WII instruments while prefix 2 indicate measurement of damping coefficient. WTD through the MIZ are listed for the stops with prefix 2.

Figure 1a is a MODIS Corrected Reflectance satellite image, which shows the ship trajectory into (red) and out of (green) the MIZ and the ice edge at approximately 82.2 °N below the thin cloud cover. The ice concentration for the four stops 1.1-1.4 was estimated to be roughly 10%, 30%, 90% and 100% respectively in Rabault et al. (2020). From Fig. 1a the ice edge is roughly located by visual inspection along the relevant longitudes and recreated in Fig. 1b (blue) along with the location of the wave attenuation measurements, i.e. stops 2.1-2.7 (black). This figure also indicates the
2.1 Data acquisition

The instrument setup consisted of an ultrasonic gauge (UG) that measured ocean surface elevation relative to the ship bow. Figure 2a and 2b shows the downward facing UG mounted on a rigid pole. Estimated absolute surface elevation was obtained after correcting for ship motion by means of an IMU placed on deck, also in the bow section of the ship. The horizontal and vertical distances between the UG and the IMU were $x_{LEV} = 2.5$ m and $y_{LEV} = 6$ m respectively. We used a UG (Banner QT50ULB) with approximately 0.2-8 m range. The instrument emits 75 kHz ultrasonic pulses at a 10.4 Hz sampling rate. Distance $L$ between the UG and the ocean surface is calculated internally from the time delay of the echo, where the speed of sound in air is temperature compensated with an integrated thermometer. Mean $L$
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Figure 3: Saturation proportion (i.e. proportion of time when the UG range was exceeded) for stops 2.1-2.7, where "total" saturation proportion is the sum of "low" and "high" saturation proportion. Area of discarded data due to total saturation exceeding 10% threshold is shaded.

was approximately 5 m and the effective beam width, i.e. the diameter of the circular area where the signal was reflected on the ocean surface, was approximately 0.9 m. This footprint is very low compared to the wavelength $\lambda = 156$ m of a typical 0.1 Hz ocean wave, as the one shown in Fig. 4. The wavelength is found from $\lambda = \frac{2\pi}{k}$, and the wavenumber $k$ from the linear deep water dispersion relation:

$$\omega^2 = gk,$$

(1)

where $\omega = 2\pi f$ is the angular frequency, $f$ the wave frequency and $g$ the acceleration due to gravity. The depth dependency is neglected here since the water depth was $\approx 3500$ m in the region. The use of (1) in the MIZ can be justified by the fact that the dispersion relation in ice deviates little from the deep water dispersion relation in the frequency band where wave motion is present (Marchenko et al. 2017).

A feature of the UG called Auto-Window was enabled, meaning that a 1 m sensing window was centered around a taught length. The taught length $\bar{L}$ was found automatically by time averaging $L$ over a couple of wave periods before each measurement was initiated. We later display the output signal of the instrument as $D = L - \bar{L}$ in Figures 4, 5a and 5b. The Auto-Window feature was chosen out of practical reasons and to increase instrument accuracy at the expense of a reduced range ($|D| < 0.5$ m), because the intention was initially to only measure in the Barents Sea MIZ where low wave heights are normally expected. However, it was decided to measure continuously during the whole cruise, also in the open ocean where the waves were generally larger than in the MIZ. Consequently, the instrument range was exceeded (saturated) at times with large wave amplitudes. Note that saturation only occurred when $|D| > 0.5$ m, not necessarily when the trough-to-crest amplitude exceeded 1 m, depending on ship response. For future expeditions we would recommend to pre-calibrate the UG to allow for a larger sensing window if rougher waves are expected, for example in the Southern Ocean MIZ, to reduce the probability of saturation. With the right instrument calibration, the measurement range could rather be limited by the actual instrument range and the mean vertical distance between the instrument and the surface (0.2-8 and 5 m respectively in our case). Christensen et al. (2013) reports an equivalent setup of the system to be useful for significant wave heights up to 4 m.

We define a low saturation when $D < -0.5$ m and a high saturation when $D > +0.5$ m. In these cases, the output signal of $D$ is simply $\pm 0.5$ m, although the distance is actually smaller or greater, and this can be seen as a "cut" graph in Fig. 5b. We define the saturation proportion as the ratio of time where saturation occurs over total sample time. Figure 3 shows the saturation proportion of each sample for stops 2.1-2.7, where "total" saturation proportion is the sum of "low" and "high" saturation proportion. We accept up to 10% total saturation and will therefore discard the sample recorded at stop 2.7, which had a total saturation of approximately 20%, in our wave attenuation analysis in Section 3.2.

The system can measure trough-to-crest amplitudes up to 1 m before the range of the UG is exceeded. This constraint sets an upper limit for how large significant wave height the instrument theoretically can record in a time series without exceeding the measurement range. Significant wave height ($SWH$) from time series is defined as:

$$SWH = 4\sigma,$$

(2)
where $\sigma$ is the standard deviation of the surface elevation $\eta$. In the simplified case of a pure sine wave, the trough-to-crest amplitude is equal to $2\sqrt{2}\sigma$. In this idealized example where ship motion is not considered, the limiting upper value of $SWH$ is approximately 1.4 m for the UG.

As motion correction device, we used an IMU (VectorNav VN100) similar to the ones in Rabault et al. (2020). It features 3-axis accelerometers and 3-axis gyroscopes measuring at a rate of 800 Hz. After an internal Kalman filtering, the instrument gives an output frequency of 80 Hz. The gyros yield rotation angles about all three axis directly. Vertical acceleration is integrated twice to obtain ship vertical displacement about the mean, where positive values are displacement above the mean. Details on the integral scheme, data filtering and other technical information on the instrument can be found in Rabault et al. (2020).

In order to obtain time series of the surface elevation, UG and IMU data at the same time instance were needed. We solved this by defining a common sampling rate of 10 Hz, which should be sufficient for resolving all relevant ocean surface features. All data were then interpolated on the common time base for the analysis.

2.2 Ship response

In general, if the wavelength $\lambda$ is large when compared with the diameter $d$ of a floating body, the body will tend to follow the path of a fluid particle at the free surface. In the opposite case when $\lambda/d$ is small, the waves are losing their influence on the behavior of the body. The response amplitude operator (RAO) of a vessel (or any object) is the ratio of response amplitude over incoming wave amplitude. RAOs in all modes for R/V Kromprins Haakon are reported in Yiterland (2016). RAOs depend on the wave heading angle $\beta$ which is defined as the relative angle between the ship heading and wave direction. Hence, $\beta = 0^\circ$ corresponds to following seas, $\beta = 90^\circ$ corresponds to beam seas, and $\beta = 180^\circ$ corresponds to head seas.

2.3 Motion correction

Downwards facing altimeters mounted on fixed structures are common for measuring ocean surface elevation $\eta$, which is a function of time (Reistad et al. 2011, Magnusson & Donelan 2013). Correction is required when the altimeter is mounted on a floating structure like a ship. The parameter $\eta$ is a function of the position in the horizontal plane in addition to time when the ship is free to move, but we assume the horizontal drift velocity to be much smaller than the phase speed of the waves, so that the dependency of position can be neglected. We define a coordinate system with the $(x, y, z)$ axis to be aligned horizontally in the direction from stern to bow, vertically in upward direction and horizontally in the direction from port to starboard, respectively. The origin coincides with the IMU. Along axis translation are surge, heave and sway while rotation about the axis are roll, yaw and pitch, respectively. Only heave $\xi$, roll angle $\phi_R$ and pitch angle $\phi_P$ affect the distance $L$ measured by the UG and need to be addressed when compensating.

In the (imaginary) case where the ship lies completely still, as if it was fixed in position, surface elevation will be given by $\eta = \bar{L} - L$. When heave response is present, we simply subtract the vertical displacement $\xi$, which is positive when the ship is above its mean level and negative when the ship is below. When rotation about the horizontal axis is included, two aspects need to be considered. First, the UG measures the distance to the surface with an angle about the vertical axis. The true vertical distance is obtained by multiplying $L$ with the cosine of both $\phi_R$ and $\phi_P$ as shown in the second term on the R.H.S. of (3). Second, there is the lever effect due to the horizontal and vertical distance between the UG and the IMU, $x_{lev}$ and $y_{lev}$ respectively. The UG is displaced in vertical direction relative to the ocean surface when the UG rotates around the IMU about the $x$-axis and the $z$-axis, which corresponds to ship roll and pitch respectively. We name these displacements roll and pitch elevation effects, and define them respectively as $y_{lev}(1 - \cos(\phi_R))$, which is always positive, and $x_{lev}\sin(\phi_P)$, which is positive when $\phi_P$ is positive and negative otherwise.

All corrections are combined and we obtain an expression for the ocean surface elevation:

$$\eta = \bar{L} - L\cos(\phi_R)\cos(\phi_P) - \xi - y_{lev}(1 - \cos(\phi_R)) - x_{lev}\sin(\phi_P), \quad (3)$$

which is similar to Eq. 1 in Christensen et al. (2013) except from the roll and pitch elevation effects in the fourth and fifth term on the R.H.S. of (3), which are new considerations.

A typical example of a 150 s time series inside the MIZ is shown in Fig. 4. Trough-to-crest amplitudes up to approximately 0.4 m and frequencies of around 0.1 Hz are found in this specific sample, as seen in the upper panel. Ship heave response $\xi$ and UG reading $D = \bar{L} - L$ are also presented in the upper panel and are in the same order of magnitude as the surface elevation. The lower panel shows roll and pitch elevation effects (fourth and fifth term on R.H.S. of (3) respectively) in addition to the UG angle correction factor corresponding to the second term on R.H.S. of (3), here presented as $1 - \cos(\phi_R)\cos(\phi_P)$. These results indicate that the roll elevation effect and the UG angle
correction factor are $O(10^{-3})$ relative to the wave amplitude and thus negligible in this case. The pitch elevation effect is on the other hand approximately $0.02 \text{ m}$ and $O(10^{-1})$ relative to the wave amplitude and is therefore important to include in the processing.

A section of the time series from stop number 1.1 is presented in Fig. 5a, where the mean value of the surface elevation $\eta$ jumps from zero up to about $0.1 \text{ m}$ after roughly $35 \text{ s}$ and then jumps back to zero at roughly $100 \text{ s}$. This step is most likely caused by an ice floe drifting under the sensing window of the UG for a short period, which for example also occurs at stop number 2.4 as illustrated in Fig. 2b, and demonstrates the UG’s ability to receive reflected signals off an ice cover. The crest-to-trough amplitude is approximately $0.4 \text{ m}$ for the waves and $0.1 \text{ m}$ for the step presented in Fig. 5a. If the step is interpreted as a long wave, its frequency is less than $0.008 \text{ Hz}$, which is much lower than the lower cut-off frequency $f_{\text{min}}$ explained in Section 2.4, and should therefore not affect the spectral analysis presented in this study.

An example of a saturated sample can be seen in Fig. 5b. This time series is extracted from stop number 2.7. The UG reading $D$ is clearly flattened out at the peaks where the range of the instrument is exceeded. Total saturation for this sample was approximately $20\%$.

### 2.4 Statistical parameters and spectrum

We obtain surface elevation from (3). Power spectrum density $PSD(f)$ of the surface elevation is obtained with the Welch method (Earle 1996) where the samples are subdivided in $q$ consecutive segments and ensemble averaged. Segment size is set to $200\text{ s}$ with $50\%$ overlap. With $20\text{ min}$ sampling time at $10\text{ Hz}$ common sampling rate, this gives a segment size of $2000\text{ sampling points}$ and a total number of $12000\text{ sampling points}$ for each measurement. A Hanning window is applied to each segment to reduce spectral leakage.

Since we are later comparing our measurements with data from a spectral model, we find the statistical parameters from spectra. The mean wave period and the significant wave height are obtained from the spectral moments:

$$m_j = \int_{f_{\text{min}}}^{f_{\text{max}}} f^j PSD(f) df,$$

where the cutoff frequencies $f_{\text{min}}$ and $f_{\text{max}}$ are set to $0.04 \text{ Hz}$ and $1.0 \text{ Hz}$ respectively, which should include the most energetic ocean waves. The same cutoff frequencies are applied in the spectral model we have compared our
results with. In our comparison of integrated parameters from bow instruments and WII instruments at stops 1.1-1.4, we use 0.05-0.25 Hz, which corresponds to the cutoff frequencies applied by the on-board processing unit of the WII instruments. The spectral moments can be used to estimate the mean ($T_{m01}$) and zero up-crossing ($T_{m02}$) periods from

$$T_{m01} = \frac{m_0}{m_1},$$  \hspace{1cm} (5)

and

$$T_{m02} = \sqrt{\frac{m_0}{m_2}},$$  \hspace{1cm} (6)

respectively.

We use significant wave height $H_s$ from spectra for most of our analysis here. $SWH$ from time series of surface elevation (Eq. 2) is used for comparison in a redundancy check for the measurements, as the two methods ideally should yield the same result. $H_s$ is defined as:

$$H_s = 4\sqrt{m_0}. \hspace{1cm} (7)$$

Wave attenuation is dependent on frequency. It is therefore convenient to have a measure for wave amplitude as function of frequency when finding the spatial damping coefficient. We define a spectral amplitude $a$ at discrete forcing frequency $f_0$ as:

$$a(f_0) = \sqrt{\int_{f_0 - \Delta f}^{f_0 + \Delta f} PSD(f) df}, \hspace{1cm} (8)$$

where $\Delta f$ is the bandwidth frequency set to 0.005 Hz. Six discrete forcing frequencies (0.076-0.128 Hz with 0.0104 Hz increments) are chosen based on the spectra presented in Section 3.2. Equation 8 is similar to the definition of Meylan et al. (2014), used for attenuation analysis in the Antarctic MIZ. The spectral amplitude $a(f_0)$ is not the physical wave amplitude in the classical sense, but an interpretation of the energy content in a finite frequency proportion of the PSD with unit meter.

Confidence intervals for both spectra and the integrated parameters, significant wave height and amplitude, are calculated from the Chi-squared distribution, following the methodology presented in Young (1995). For spectra, the total degree
of freedom (TDF) is calculated as \( TDF = 2q \), and for significant wave height, TDF is found with Eq. 9 below (ITTC 2017).

\[
TDF = \frac{2q \left[ \int_{f_{\text{min}}}^{f_{\text{max}}} PSD(f) df \right]^2}{\int_{f_{\text{min}}}^{f_{\text{max}}} [PSD(f)]^2 df}.
\]  

Equation 9 is also used when finding the confidence intervals for \( a(f_0) \), with the small modification of multiplying the bandwidth frequency \( \Delta f \) to the denominator.

We use the mean absolute percentage error (MAPE) to compare our system with either the spectral model or the WII instruments. Systematic bias is described with the mean percentage error (MPE). The error statistics are defined as:

\[
MAPE = \frac{100\%}{n} \sum_{t=1}^{n} \left| \frac{X_t - Y_t}{X_t} \right|
\]  

\[
MPE = \frac{100\%}{n} \sum_{t=1}^{n} \frac{X_t - Y_t}{X_t}
\]

where \( Y_t \) are parameters obtained from bow measurements and \( X_t \) are reference parameters.

2.5 Wave models

WAM-4 is a third generation spectral wave model (Komen et al. 1996, Saetra & Bidlot 2004), run operationally by the Norwegian Meteorological Institute. Its performance has been validated against in-situ and EnviSat Radar Altimeter observations (Carrasco & Gusdal 2014, Gusdal et al. 2011). Spatial and temporal resolution of the model is 4 km and 1 hour, respectively. It runs twice a day with a forecast period of 66 hours per run. In this study, we have concatenated forecasts from consecutive runs to provide a continuous time series. A hard ice boundary based on satellite images (ice concentration over 3/10th) is defined in the model.

ERA5 reanalysis of global atmosphere and ocean waves is produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). ERA5 combines historical observations into past forecasts using data assimilation systems to model meteorologic and oceanographic parameters. The operational wave model at ECMWF assimilates satellite synthetic aperture radar observations and altimeter wave heights into the numerical scheme of the WAM spectral wave model, which is coupled with the IFS atmospheric model to account for wind forcing (Janssen & Bidlot 2003). The sea ice concentration (SIC) is based on coupled atmosphere ocean simulations of the IFS model combined with satellite observations (Hirahara et al. 2016), and wave parameters are not given for \( SIC > 50\% \). Reanalysis wave data from ERA5 have hourly output and a horizontal resolution of 40 km.

No wave simulations are performed where the assumed ice cover is present, although the models still provide wind information everywhere within the domain. We have used model data from WAM-4 and ERA5 as a comparison to our wave observations outside the MIZ. The model ice edge was defined at 82.065° N and 82.175° N by WAM-4 and ERA5 respectively at the relevant longitude on September 19. The horizontal resolution of the models are indicated with a white and orange square in Fig. 1a for ERA5 and WAM-4 respectively. From the models, we extract total mean wave direction \( THQ \), \( H_S \), \( T_{m01} \), \( T_{m02} \) (from WAM-4 only) and wind parameters. \( THQ \) is defined as the mean of the two-dimensional wave spectrum over all frequencies and directions. The wave parameters are integrated parameters from spectra, but the raw spectra were not available.

2.6 Waves-in-ice instruments

Four in situ WII instruments (Rabault et al. 2020), placed on ice floes close to the ship at the stops 1.1-1.4, have been used to cross-validate our system inside the MIZ. The instruments measure waves with an integrated IMU, and they have been tested and used in a series of previous works (Rabault et al. 2016, 2017, Sutherland & Rabault 2016, Marchenko et al. 2017). The WII instruments are autonomous and designed to be deployed for long durations. Compressed power spectra from 20 min time series of surface elevation were sent via Iridium satellite and used as comparison with the spectra obtained from the bow measurements. We also compare the integrated parameters \( H_S \) and \( T_{m02} \).

The WII instruments measured with five hours intervals to conserve battery power. Data from the first measurement were corrupted due to disturbances during carrying and placement on the ice and were discarded. The results presented
here are from the second measurement of each instrument, i.e. five hours after the respective bow measurement. The temporal change of sea state over these five hour periods are addressed in Section 3.1. The WII instruments drifted 2.4 km, 1.7 km, 3.5 km and 2.2 km during the five hours between first sample taken at stop 1.1-1.4 respectively, and second sample. These are relatively small displacements compared to the distance between the WII instruments, which were an order of magnitude larger. Hence, the measurements with the two different systems were made at approximately the same place with five-hour time delay.

2.7 Attenuation modeling

Previous field measurements indicate that waves decay exponentially in ice (Squire & Moore 1980, Wadhams et al. 1988). For each frequency bin corresponding to the six forcing frequencies $f_0$, the spectral amplitudes $a$ are fitted to decreasing exponentials on the shape $a = Ce^{-\alpha x}$, where $C$ and $\alpha$ are estimated parameters and $x$ is wave traveling distance through the MIZ, by means of non-linear least squares. From the fitted curves, the spatial damping coefficients $\alpha$, which describe wave attenuation are determined from:

$$\frac{\partial a}{\partial x} = -\alpha a,$$

(12)

for each frequency bin. Non-linear least squares are applied to fit a power function $\alpha = Af^p$ to the measured values of $\alpha$ as function of frequency $f$, where $A$ and $p$ are the best-fit parameters. The standard error in the exponent is obtained from the square root of the variance of $p$.

Spatial damping coefficients found in the field measurements are compared to the model of Sutherland et al. (2019), which presents a parameterization for wave dissipation that allows for a two-layer structure within the ice. The lower layer with thickness $h_i$, where $h_i$ is the total ice thickness, is defined as a highly viscous layer where wave motion exists. The upper layer is defined as impermeable. With a no-slip boundary condition imposed on the bottom of the ice, Equation (16) of Sutherland et al. (2019) can be written as:

$$\alpha_{mod} = \frac{1}{2} \frac{\epsilon}{h_i} k^2,$$

(13)

where $k$ is the wavenumber, here calculated with (1). Total ice thickness is set to $h_i = 1.1$ m based on visual observations from the field campaign. We use $\epsilon = 0.02$ due to the large ice thickness, meaning that wave motion is assumed to exist in a small fraction of the total ice thickness. Note that (13) is proportional to frequency to the fourth power.

3 Results

3.1 Data comparison

Independent in-situ wave measurements were not available for validation of the observations in the open ocean on this cruise. Therefore, we compare our results with WAM-4 and ERA5 spectral model data. It is worth mentioning that Christensen et al. (2013) performed and documented validation of an equivalent setup of the system in open water against a moored Waverider buoy, model data from the ERA Interim Reanalysis and satellite observations from AVISO.

Bow measurements outside the MIZ are compared with spectral model data at the ship location in Fig. 6. The period with no available model data due to the ice cover is highlighted with red background color and is not included in the comparison between observations and the spectral models. Valid measurements, i.e. time periods where total UG saturation is below 10% and ship speed over ground (SOG) is below 0.5 m/s, are highlighted with green background color. The whole comparison spans over eight days, and periods of valid measurements are found 23% of the time. Error statistics are summarized in Table 2 where MAPE describes the mean absolute error and MPE describes the mean error. In general, there is a good agreement between observations and models, also for the conditions not defined as valid. The MPEs of WAM-4 do not exceed ±10% for $H_S$ and $T_{m01}$ in the valid periods, which indicate low systematic bias. ERA5 systematically overestimates $H_S$ compared to the observations, but matches the observed $T_{m01}$ better than WAM-4. A surprising result is that significant wave height has a higher MAPE in the valid periods than outside, although the difference is not large (18.9% vs 13.3% for WAM-4 and 29.2% vs 21.3% for ERA5). MAPEs for $T_{m01}$ are about the same inside the valid periods as outside. Note that significant wave height does not exceed the theoretical limit (for sine waves) of 1.4 m as described in Section 2.1 within the valid data periods. MAPE between measured $H_S$ and SWH was 11.7% during valid measurements (not shown in Table 2).
Figure 6: Long time comparison between observations and the models WAM-4 and ERA5. Significant wave height from time series and spectra (upper), and periods from spectra (lower). Time periods where total saturation is below 10% and ship speed over ground (SOG) is below 0.5 m/s are highlighted with green background color and the period with an ice cover (hence no model data) is highlighted with red background color. Note that ERA5 provides wave data further into the MIZ than WAM-4.

<table>
<thead>
<tr>
<th></th>
<th>WAM-4</th>
<th>ERA5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Valid MAPE [%]</td>
<td>18.9</td>
<td>15.0</td>
</tr>
<tr>
<td></td>
<td>17.2</td>
<td>17.2</td>
</tr>
<tr>
<td></td>
<td>29.2</td>
<td>13.4</td>
</tr>
<tr>
<td></td>
<td>-14.4</td>
<td>3.4</td>
</tr>
<tr>
<td>Not Val. MAPE [%]</td>
<td>13.3</td>
<td>15.7</td>
</tr>
<tr>
<td></td>
<td>14.7</td>
<td>16.4</td>
</tr>
<tr>
<td></td>
<td>21.3</td>
<td>16.4</td>
</tr>
<tr>
<td></td>
<td>28.5</td>
<td>3.4</td>
</tr>
<tr>
<td>Valid MPE [%]</td>
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<td>-8.9</td>
</tr>
<tr>
<td></td>
<td>-14.4</td>
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<td>3.1</td>
</tr>
<tr>
<td></td>
<td>19.5</td>
<td>3.1</td>
</tr>
</tbody>
</table>

Table 2: Long time error statistics for significant wave height $H_S$ and periods $T_{m01}$ and $T_{m02}$ when comparing bow measurements with the WAM-4 and the ERA5 model. Valid periods are defined as when total UG saturation is less than 10% and ship speed over ground is less than 0.5 m/s (23% of the time). All other times are considered not valid.

A stationary sea state over the measurement period inside the MIZ is an advantage in our analysis. The sea state is investigated in Fig. 7 where time series of $H_S$, $T_{m01}$ and $THQ$ from WAM-4 are presented. The model data are extracted from the six points arranged in a grid configuration outside the MIZ, shown in cyan in Fig. 1a. Time series from each grid point are plotted in gray and the mean value in a different color. All parameters were close to constant between 11:00 and 17:00 but varying with time outside of this period. This indicates a variability in sea state in the comparison of bow and WII instruments, due to the five-hour time delay between the measurements. However, the change was not dramatic. From 04:22 to 09:22 at stop 1.1, the standard deviation in the mean samples (containing five values) were 0.06 m, 0.10 s and 1.00° for $H_S$, $T_{m01}$ and $THQ$ respectively. From 12:20 to 17:20 at stop 1.4, the standard deviation in the mean samples are an order lower for all parameters. The results in Fig. 7 also indicate a variability in sea state over the period from 13:11 to 19:08 where wave damping was measured. In this period, the standard deviations in the mean samples (containing seven values) were 0.03 m, 0.10 s and 1.64° for $H_S$, $T_{m01}$ and $THQ$ respectively.
A comparison of ship measurements and available WAM-4 model data from stops 2.1-2.6 is presented in Fig. 8. The upper panel shows ship heading and speed over ground (SOG), extracted from the ship navigation system. Forecasted wind direction and speed presented in the lower panel are compared to 10 min average values (corrected for ship motion) acquired by the ship anemometer. The six stops when measurements are performed are clearly visible as when the SOG graph flats out at close to zero value. These periods are highlighted with green background color.

As the ship cruised between stops, the heading direction was about 190°. During the measurements, the heading direction was 240° at stop 2.1, roughly 140° at stop 2.2-2.5 and about 315° at stop 2.6. The wind direction was almost constantly 45° over the six-hour period. It is clear that the ship was oriented approximately perpendicular to the wind direction when it drifted freely during the stops, except from stop 2.1 where the high ice concentration most likely prevented the ship from rotating. For stop 2.2-2.5, the wind came in on port side of the vessel, while it came in from starboard side at stop 2.6.

There is generally a good agreement between forecast and measured wind direction (WD) and speed (WS), although the model underestimated WS by 50-75% in the period from roughly 15:30 to 18:30. THQ from model and HS from model and bow measurements are also presented in the lower panel of Fig. 8. The overestimated wind speed could possibly explain why forecast HS is 95% higher than measured value at 20:09 when wave model data were available, as WAM-4 is forced with winds at 10m height from the atmospheric model UM4. Also, occasional ice floes were observed at this point and WAM-4 does not take attenuation caused by the presence of ice into considerations. However, the measured wave data were saturated at this point and cannot be completely trusted. THQ was 282.1° at 20:09. Due to the low variation in THQ (in time and space) over the period when attenuation measurements were sampled, as seen in Fig. 7, this value (282.1°) is used to find the WTD through the MIZ, presented in Table 1 and Fig. 1b. The wave heading angle \( \beta \) was 40° at stop 2.1, 140° at the four intermediate stops 2.2-2.5 and 35° at stop 2.6. Wave heading angle will be further assessed in Section 4.

Power spectra from bow measurements (black) and WII instruments (blue) with their respective 95% confidence intervals are compared in Fig. 9, going successively from stop number 1.1 (upper panel) to 1.4 (lower panel). The power spectra from the two methods are consistent for most frequencies in the two first samples closer to the open ocean. The third sample is consistent for some frequencies up to 0.00 Hz. For higher frequencies, the WII spectrum flattens out while the bow spectrum peaks at 0.1 Hz. Investigation of bow spectra from samples taken respectively twenty and forty minutes after the PSD presented here while the ship was still stationary around the WII instruments, reveals similar spectral peaks at the same frequency. This consistency substantiates the validity of the bow measurements. A discussion on this discrepancy follows in Section 4. The fourth sample furthest into the ice displays a fair agreement between the two methods.
Figure 8: Comparison of observations and WAM-4 model. Ship heading and SOG (upper), and wind direction (WD) and wind speed (WS) from ship measurements and model, total mean wave direction $THQ$ from model and significant wave height $H_S$ (multiplied with a factor of 10) from bow measurements and model (lower). Directions increase clockwise from geographic north (zero degrees). Times with valid data are shaded.

Figure 9: Validation of bow measurements in the MIZ with WII instruments placed on ice floes. PSD are presented from bow measurements (black) and from WII instruments (blue) with their respective 95% confidence intervals as shaded regions, going further into the ice zone from upper (stop 1.1) to lower panel (stop 1.4).
Figure 10: Power spectra at different latitudes with 95% confidence intervals, and ship resonance frequencies in vertical modes from Ytterland (2016).

Integrated parameters from the spectra in Fig. 9 are summarized in Table 3. There is a good match for both significant wave height and mean period where MPEs do not exceed ±10% for all stops except 1.3 where the error in $H_S$ is −87.8%. We use MPE as defined in (11), although it is not strictly a mean error in this context where only single point parameters are compared.

<table>
<thead>
<tr>
<th>Stop</th>
<th>Significant wave height</th>
<th>Mean wave period</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$H_{S, bow}$ [m]</td>
<td>$H_{S, WII}$ [m]</td>
</tr>
<tr>
<td>1.1</td>
<td>0.36</td>
<td>0.39</td>
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<tr>
<td>1.2</td>
<td>0.29</td>
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<td>1.3</td>
<td>0.16</td>
<td>0.09</td>
</tr>
<tr>
<td>1.4</td>
<td>0.04</td>
<td>0.04</td>
</tr>
</tbody>
</table>

Table 3: Significant wave height and zero up-crossing period from bow measurements inside the MIZ are compared with WII measurements. Errors are included.

3.2 Wave attenuation

One dimensional power spectra of the surface elevation with 95% confidence intervals from stops 2.1-2.6 are presented in Fig. 10. Most of the energy is found at either low (swell) or high (wind wave) frequencies. Averaged peak frequency is found to be 0.076 Hz for swell and 0.128 Hz for wind wave. Spectral amplitudes $a(f_0)$ for a set of six finite frequency bins are found from (8) and investigated at stops 2.1-2.6. The highest and lowest $f_0$ are set to 0.076 Hz and 0.128 Hz respectively. Four intermediate forcing frequencies are evenly distributed between the swell and wind wave frequency.

It is evident that wave energy content increases as the ship approaches the open ocean. Figure 11a shows an exponential decrease in $H_S$ (with 95% confidence intervals) as function of WTD through the MIZ for stops 2.1-2.6. Ideally, the measurements should have been performed simultaneously to ensure stationary wind forcing throughout the attenuation run. Fortunately, wind conditions did not change dramatically over the period from 13:11 to 19:08 as can be seen in Fig. 8, and we therefore assume the wave generation to be approximately stationary.

Spectral amplitude for swell, the third intermediate wave, and wind wave are plotted versus wave travel distance through the MIZ and presented in Fig. 11b along with their 95% confidence intervals. A clear exponential attenuation for wind wave can be seen. The exponential shape is less obvious for the intermediate frequency and swell, but the trend is visible. Also, swell amplitude seems to decrease from stop 2.5 to 2.6, which is not physical, and this problem is addressed in Section 4. Exponential fits lie within the confidence intervals of $a(f_0)$ for almost all frequencies.
Figure 11: Wave damping with 95% confidence intervals. a) Significant wave height $H_S$. b) Spectral amplitudes $a(f_0)$ for swell, one of the intermediate wave frequencies and wind wave. Exponential decay functions are fitted to the data.

Figure 12: Spatial damping coefficient $\alpha$ as function of frequency from measurements (black) and two-layer model (blue). Power coefficients for the fitted curve and the model are displayed in the legend. 1σ errors associated with the power coefficient estimate for the fitted curve (gray) are shown as the gray shaded region.

The spatial damping coefficients $\alpha$ are plotted versus their respective $f_0$ with black asterisks in Fig. 12. We obtain values of $\alpha$ from $0.67 \times 10^{-5} \text{m}^{-1}$ at 0.076 Hz to $4.44 \times 10^{-5} \text{m}^{-1}$ at 0.128 Hz. These values are within the same order of magnitude as other attenuation observations within the same frequencies reported in the literature where the measurements are done over approximately the same wave traveling distance into the MIZ, e.g. Squire & Moore (1980) ($\alpha = 2.7 - 8.6 \times 10^{-5} \text{m}^{-1}$) and the February 26, 1983 event reported in Wadhams et al. (1988) ($\alpha = 1.6 - 4.8 \times 10^{-5} \text{m}^{-1}$).

Observed wave attenuation increase with frequency in the shape of a power function. The fitted power curve is plotted with a dashed gray line in Fig. 12, and the $1\sigma$ uncertainty associated with the power law exponent is marked with a gray shaded area. The fitted exponent is $3.51 \pm 0.41$, where the uncertainty is the standard error from the curve fit. Values of $\alpha$ obtained from (13) are plotted with a blue line in Fig. 12. The two-layer model of Sutherland et al. (2019) describes the frequency dependence of $\alpha$ quite well for the frequencies considered here. It can be noted that the model slightly overestimates the frequency dependence for frequencies above 0.12 Hz.
4 Discussion

A culture of sharing the design of instruments within the scientific community would enable a more cost effective and increased quantity of data collection in the Arctic. We would like to stimulate researchers to use off-the-shelf sensors and open-source code and therefore make all our material available (See Appendix A). The components used to obtain wave data in this study are relatively cheap compared to industry made black-box instruments. The system can record autonomously, hence, expensive expedition time does not have to be allocated as measurements can be done in parallel. Also, cruises in the Arctic with other purposes than measuring waves in ice, for example studies within biology or meteorology, could benefit from installing the system presented here, which provides wave measurements without the need for human intervention. Other devices such as floaters and buoys offer the advantage of multiple simultaneous measurements at different locations, which is not possible with bow mounted sensors. However, bow mounted sensors do not require costly and time-consuming deployment and retrieval operations, and there is low risk associated with losing instruments at sea.

The primary focus of this study is to present a new method for wave measurements in the MIZ. Spectral wave models have been included only as a supplement for comparison with the observations. Nevertheless, we provide a short discussion on the difference between the models and how well they correspond to the bow system. As mentioned in Section 2.5, ERA5 is a reanalysis of previous observations and simulations, and is therefore considered more reliable than a forecast model, such as WAM-4. However, WAM-4 matches the observations better than ERA5 in terms of significant wave height, especially in the vicinity of the MIZ. No wave models are the true realization of the ocean, but there are arguments that support the accuracy of WAM-4 over ERA5 in our situation. ERA5 provides wave information for sea ice concentration up to 50%, while the hard ice boundary in WAM-4 is defined at 30%. In other words, wave information is available further north from ERA5 than from WAM-4 as seen from Fig. 6, although none of the models consider the effect of wave damping due to the presence of sea ice. Consequently, the modeled significant wave height furthest into the MIZ deviates the most from the real values. Additionally, ERA5 is a global model with a horizontal resolution of 40 km, while WAM-4 is 10 times finer resolved, as seen in Fig 1a. Considering the fact that the measurements were done relatively close to the Svalbard and Franz Josef Land archipelagos, the horizontal resolution is likely to be more important for the accuracy of the models than for example in the middle of the Pacific.

Data recorded while a ship is in motion will contain a Doppler shifted frequency according to the wave heading angle $\beta$ (Collins III et al. 2017), which could be problematic in the calculation of periods. Peak and mean periods were reported strongly biased by the Doppler shift in Christensen et al. (2013), while the estimates of significant wave heights were reasonable since they only depend on the sea surface variance. In the present study, we consider the measurements invalid either when the ship is cruising (and a Doppler shift could be introduced depending on $\beta$) or when the UG is saturated (exceeded measure range), as we have not differentiated between these two cases. We see little difference in MAPE inside versus outside the time periods we consider valid, when comparing periods from bow measurements with model data. MPES are actually higher in the periods where data is considered valid. There is also a slightly better match outside the valid periods in the comparison of significant wave height. This is somewhat counter intuitive, but there is of course uncertainty associated with the spectral models, and $\beta$ is not included in the analysis. It is possible to correct the Doppler shift, but this requires accurate observations of the wave direction, or more precisely $\beta$ (Collins III et al. 2017). This would also allow for possibly improved mean period estimates during cruising.

Significant wave energy for $f > 0.1$ Hz was detected by the bow mounted sensors and not by the WII instruments at stop 1.3 as seen in Figure 9. Yiew et al. (2016) have performed experiments with thin floating disks and applied two theoretical models to determine the hydrodynamic responses of ice floes with diameter $d$ in regular waves with wavelength $\lambda$. They find RAO in heave to be 0.5 at approximately $\lambda/d = 1.5$ and rapidly decreasing for smaller $\lambda/d$. For the peak frequency measured by the bow sensor, i.e. $f = 0.1$ Hz, $\lambda$ is found to be 156 m with Eq. 1. We have no exact measure of the size of the ice floe at stop 1.3, but deeming from Fig. 1a and Fig. 9 in Rabault et al. (2020) its diameter seems to be at least in the order of hundred meters. The flexural rigidity of the ice floe will of course differ from the disks applied in Yiew et al. (2016), but the results in this paper suggest a low heave response of such a large floe at this short wavelength. As the WII instruments essentially calculate the power spectra from the heave motion of the floe, a plausible explanation for the discrepancy between the two instruments could be that waves in this frequency range were effectively damped by the floe. The bow sensor on the other hand, measured at points where the ice was broken up by the ship and was therefore able to detect higher frequencies and give a more accurate observation of the sea state.

A relevant problem is the response of the ship itself. Figure 10 shows the resonance frequencies of the ship in the vertical modes with $\beta = 150^\circ$/$30^\circ$ (due to symmetry) (Ytterland 2016), which almost corresponds to $\beta$ at stops 2 1-2-5 (10° deviation) and stop 2.6 (5° deviation). The RAO values are 1.8 and 1.9 for these angles, meaning a possible maximum amplification of 80% and 90% for roll and pitch motion respectively. Heave mode has only got a peak in RAO for $\beta = 90^\circ$, which did not occur during stops 2.1-2.6 and is therefore not shown. None of the spectral peaks
in Fig. 10 coincides with the natural frequencies in roll. However, the high frequency peaks are in the same range as the ship’s natural frequency in pitch. Pitch motion was likely amplified up to 90% when exposed to 0.12 Hz waves. However, this effect should be compensated for by the pitch motion correction performed on the data.

Stop 2.6 has been included in our wave attenuation analysis even though it was located outside of the defined ice edge in Figure 1b. This ice edge was estimated from coarsely resolved satellite images and is therefore not very accurate. Reflected wave energy from the ice edge could have been a problem at stop 2.6. However, the ice edge was not a strictly defined line, it was rather observed as a gradually decreasing ice concentration towards the open ocean and ice floes were still present around stop 2.6. Swell energy was observed to decrease from stop 2.5 to 2.6 where the opposite is expected. Our assumption of stationary wave field may not have been valid in this case, although wind measurements and WAM-4 model data suggest otherwise. There is also uncertainty associated with the total mean wave direction \( THQ \). There could have been discrepancies between \( THQ \) and mean swell direction \( THQ_{swell} \), although \( 90^\circ < THQ_{swell} < 270^\circ \) (swell traveling south) is unlikely. Another possible explanation is that ship yaw (i.e. rotation in the horizontal plane) and/or a changing \( \beta \) might have influenced the low frequencies in the measurements. At stop 2.6 the ship heading increased continuously from approximately 200° until it stabilized at 315° towards the end of the 20 min sampling period, whereas both ship heading and \( \beta \) were almost constant for the other stops, as seen in Figure 8.

We emphasize that we are not trying to show a universal power law in Fig. 12, which would not make sense for frequencies spanning over less than a decade. Our intention is to compare the observed wave attenuation with the two layer waves-in-ice model of Sutherland et al. (2019), and not to verify the model. The power function curve fit gives a frequency dependence of \( \alpha \propto f^{3.51\pm0.41} \), which is similar to (13) where \( \alpha \propto f^4 \). Wind energy input is not considered by the model. With a wind speed of approximately 6 m/s on average over the measurement period as seen in Fig. 8, it is reasonable to assume that wave generation has taken place, which will reduce the slope of the observed attenuation at higher frequencies. This could explain why the model predicts the observation quite well for the lower frequency bins, while it slightly overestimates attenuation for the higher frequency bin, based on the six data points.

5 Conclusions

We have presented shipborne wave measurements from the MIZ with a system combining an altimeter (UG) and a motion correction device (IMU). The UG was able to reflect signals off an ice-covered surface. The current system was designed for the present cruise in the MIZ of the Barents Sea where waves are normally quite small. Hence, the choice of the present UG and its auto calibration with a 1 m range. If some cruise would plan to go to rougher seas, one could either use the same UG over its full range (8 m), or another UG with larger range. The methodology is both cost effective as components are off-the-shelf and it can be installed on a cruise with other objectives. The system has proven to be robust as instruments have measured continuously over the two-week period without suffering any damage. Our setup provides single point time series of ocean surface elevation outside of and inside the MIZ, which enables us to produce 1D power spectra and integrated parameters.

Measured data have been compared to the WAM-4 and the ERA5 spectral wave models over a period of eight days. We have found good agreement in mean and zero up-crossing periods and significant wave height. Mean absolute errors were about the same during periods where measurements were not considered valid, indicating that the system is insensitive to exceeded instrument range and/or Doppler shifted wave frequencies. Forecast significant wave height was 95% higher than the value measured inside the MIZ. This deviation might have been caused by an overestimated model wind speed compared to wind speed measured with the ship anemometer, or by the fact that occasional ice floes were present, which were not considered by the forecast model. Power spectra and integrated parameters from bow measurements have been compared with values obtained from waves-in-ice instruments placed on ice floes. A good match was found when the sensors were placed on smaller floes further out in the MIZ. The spectra agree fairly well further into the MIZ. A major discrepancy was likely caused by a large ice floe that filtered out higher frequencies where the waves-in-ice instrument was placed. Errors in \( H_S \), \( T_{m02} \) were smaller than 10% in all cases except for one of the measurements.

We observed an exponential wave attenuation when going through the MIZ. The spatial damping coefficients obtained from measurements were within the same order of magnitude as observations reported in previous studies with a similar wave traveling distance through the ice (Squire & Moore 1980, Wadhams et al. 1988). We have found a strongly frequency dependent attenuation, where \( \alpha \propto f^{3.51\pm0.41} \), which is similar to a two-layer attenuation model where \( \alpha \propto f^4 \) (Sutherland et al. 2019).
Acknowledgement

The authors are grateful to Øyvind Breivik for inviting us to the cruise. We also thank the crew of RV Kronprins Haakon for their assistance and Yuriy Batrak at MET Norway for providing satellite images. The Nansen Legacy project helped funding the cruise and loggers. Funding for the experiment was provided by the Research Council of Norway under the PETROMAKS2 scheme (project DOFI, Grant number 28082). The data are available from the corresponding author upon request.

Appendix A: Open source code and designs

All the designs and files used for setting up the system, including the code used to extract and process data, and general instructions for mounting the instruments, are made available on the Github of the author under a MIT license that allows full re-use and further development [https://github.com/jerabau129/Ultrasound_IMU_boat_waves_system [Note: The material will be available upon publication in peer-reviewed literature]). All software and designs are based entirely on open source tools, so that the designs can be easily modified and built upon.

References


Earle, M. D. (1996), ‘Nondirectional and directional wave data analysis procedures’, *NDBC Tech. Doc. 96(002).*


Paper II

A comparison of wave observations in the Arctic marginal ice zone with spectral models

Løken, T.K., Rabault, J., Thomas, E.E., Müller, M., Christensen, K.H., Sutherland, G., and Jensen, A.

A comparison of wave observations in the Arctic marginal ice zone with spectral models

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Increased economic activity and research interest in the Arctic raise the need for better wave forecasts in the marginal ice zone (MIZ). Mathematical and numerical models of wave propagation in sea ice would benefit from more in situ data for validation. This study presents shipborne wave measurements from the MIZ where altimeter readings are corrected for ship motion to obtain estimated single point ocean surface elevation. From the combined measurements, we obtain significant wave height and zero up-crossing period, as well as one-dimensional wave spectra. In addition, we provide spectra and integrated parameters obtained from inertial motion units placed on ice floes inside the MIZ. The results are compared with integrated parameters from the WAM-4 spectral wave model over a period of three days in the open ocean. We also compare our measurements outside and inside the MIZ with hindcast data from the new pan-Arctic WAM-3 model and the Wave Watch III (WW3) model for the European Arctic, which both model wave attenuation in sea ice. A good agreement is found with WAM-4 and WW3 in zero up-crossing period and significant wave height outside the MIZ, where deviations are less than 23%. WAM-3 is on the other hand up to 60% higher than observations. WW3 and WAM-3 are able to estimate the trends for significant wave height and zero up-crossing period inside the MIZ, although the discrepancies with respect to the observations are larger than in the open ocean. Wave damping by sea ice is investigated by looking at the spatial attenuation coefficients. Predicted attenuation coefficients are found to be 72-83% smaller for WW3 and 3-64% larger for WAM-3 compared to observations. Hence, further model tuning is necessary to better estimate wave parameters in the ice.
1. Introduction
The recent decline in Arctic ice cover has allowed for increased human activities in the region, which raises the importance of better forecast models and improved physical understanding of the environment to ensure safe operations (Fritzner et al., 2019). This also applies in the interface between solid ice, such as land fast ice or pack ice, and the open ocean, called the marginal ice zone (MIZ). In situ wave measurements can increase our understanding of global climate systems and provide data for calibration and validation of numerical and mathematical models to describe wave attenuation in ice. However, experimental data are relatively sparse due to the harsh and dangerous environment for both researchers and instruments, combined with the inaccessibility of the regions where sea ice is present (Squire, 2007).

In this study, we present results from shipborne wave measurements in the MIZ. The methodology, first described in Christensen et al. (2013), combines a bow mounted altimeter and a motion correction instrument. We provide estimated power spectra from ocean surface elevation and integrated parameters from spectra, which are important quantities when considering wave-ice interactions. The results are compared with the spectral wave models WAM-4, Wave Watch III (WW3) and WAM-3 in the open ocean and in the MIZ, as a validation for the capability of modelling wave attenuation by sea ice. We also compare measurements in the MIZ with data from wave measuring instruments consisting of inertial motion units (IMUs) placed on ice floes. From the significant wave height, the spatial damping coefficient from the observations and models are found and compared with each other.

In this paper, the data acquisition and processing methods are described in Section 2. The results are presented in Section 3 followed by a discussion in Section 4. Finally, the concluding remarks are given in Section 5.

2. Data and Methods
The data were obtained during a research campaign in the Barents Sea with R/V Kronprins Haakon in September 2018. Shipborne wave measurements were made continuously during cruising in the open ocean and in the MIZ, into which the ship ventured on September 19.

<table>
<thead>
<tr>
<th>Stop</th>
<th>Time</th>
<th>Position (N/E)</th>
<th>WTD km</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Bow</td>
</tr>
<tr>
<td>1.1</td>
<td>04:22</td>
<td>82.126/20.736</td>
<td>0</td>
</tr>
<tr>
<td>1.2</td>
<td>06:28</td>
<td>82.246/20.245</td>
<td>28</td>
</tr>
<tr>
<td>1.3</td>
<td>08:59</td>
<td>82.355/19.803</td>
<td>61</td>
</tr>
<tr>
<td>1.4</td>
<td>12:20</td>
<td>82.436/19.674</td>
<td>82</td>
</tr>
<tr>
<td>2.1</td>
<td>13:11</td>
<td>82.421/19.579</td>
<td>77</td>
</tr>
<tr>
<td>2.2</td>
<td>14:32</td>
<td>82.359/19.544</td>
<td>65</td>
</tr>
<tr>
<td>2.3</td>
<td>15:40</td>
<td>82.294/19.389</td>
<td>46</td>
</tr>
<tr>
<td>2.4</td>
<td>16:54</td>
<td>82.228/19.275</td>
<td>18</td>
</tr>
<tr>
<td>2.5</td>
<td>18:00</td>
<td>82.163/19.183</td>
<td>7</td>
</tr>
<tr>
<td>2.6</td>
<td>19:08</td>
<td>82.099/19.046</td>
<td>0</td>
</tr>
<tr>
<td>2.7</td>
<td>20:09</td>
<td>81.994/18.982</td>
<td>0</td>
</tr>
</tbody>
</table>
Four stops were made on the way into the MIZ to deploy in situ waves-in-ice (WII) instruments on ice floes (Rabault et al., 2020). The sea ice concentration was visually estimated to be 10, 30, 90 and 100% for stops 1.1-1.4 respectively, and the ice thickness of the floes was approximately 1 m. Upon the return to the open ocean, a total of seven stops were made to carry out measurements unaffected by cruising speed. The location, starting time and wave travel distance (WTD) through the ice for each measurement in the MIZ are summarized in Table 1. WTD indicated “Bow” is found from the approximated ice edge from Fig. 1 and the estimated wave direction from WAM-4, while “WW3” and “WAM-3” are found by combining the ice edge and the wave direction estimated by the respective model. Prefix 1 denotes the four stops into the MIZ while prefix 2 denotes the seven stops out of the MIZ.

Figure 1(left) shows the ice edge at approximately 82.2 °N below the thin cloud cover and ship trajectory into (red) and out of (green) the MIZ. The cyan line indicates the longitudes over which the wave directions are averaged to find the mean. The ice edge is roughly recreated from the satellite image and shown in Fig. 1(right). This figure also indicates the ice edge and the mean wave direction at the ice edge at noon on September 19 estimated by the spectral models, which is further described in Section 2.2. We use the "going-to" convention for wave direction and define directions as clockwise rotation from the geographic north.

The instrument setup consisted of an ultrasonic gauge (UG) mounted on a rigid pole that measured ocean surface elevation relative to the ship bow. We used a UG (Banner QT50ULB) with approximately 0.2-8 m range. The instrument emits 75 kHz ultrasonic pulses at a 10.4 Hz sampling rate. Estimated absolute surface elevation was obtained after correcting for ship motion by means of an IMU placed on deck, also in the bow section of the ship. As motion correction device, we used an IMU (VectorNav VN100). It features 3-axis accelerometers and 3-axis gyroscopes measuring at a rate of 800 Hz. After an internal Kalman filtering, the instrument gives an output frequency of 80 Hz. The gyros yield rotation angles about all three axis directly. Vertical acceleration is integrated twice to obtain ship vertical displacement about the mean. Details on the integral scheme, data filtering and other technical information on the instrument can be found in Rabault et al. (2020).

In order to obtain time series of the surface elevation, UG and IMU data at the same time instance were needed. We solved this by defining a common sampling rate of 10 Hz, which should be sufficient for resolving all relevant ocean surface features. All data were then
interpolated on the common time base for the analysis. See Løken et al. (2021, Eq. 3) for details on the motion correction and estimates for the ocean surface elevation. The UG range was exceeded (saturated) at times with large wave amplitudes. We accept up to 10% saturation in a time series and will therefore discard the sample recorded at stop 2.7 in further analysis, which had a total saturation of approximately 20%.

2.1. Spectrum, statistical parameters and wave attenuation

Power spectra $S(f)$ of the surface elevation is obtained with the Welch method, following Earle (1996). Samples are subdivided in $q$ consecutive segments and ensemble averaged. Segment size is set to 200 s with 50% overlap. 20 min sampling time at 10 Hz yields a segment size of 2000 and a total number of 12000 sampling points for each measurement. A Hanning window is applied to each segment to reduce spectral leakage. Peak frequency $f_p$ is defined as the frequency at the spectral peak. The mean zero up-crossing period and the significant wave height are obtained from the spectral moments:

$$m_j = \int_{f_{min}}^{f_{max}} f^j S(f) df,$$

where the cutoff frequencies $f_{min}$ and $f_{max}$ are set to 0.04 Hz and 1.0 Hz, respectively, which should include the most energetic ocean waves. Approximately the same cutoff frequencies are also applied in the spectral models. Zero up-crossing periods $T_{m02}$ are estimated from:

$$T_{m02} = \sqrt{\frac{m_0}{m_2}}. \quad [2]$$

We investigate significant wave heights $H_S$ estimated from spectra, defined as:

$$H_S = 4\sqrt{m_0}. \quad [3]$$

Confidence intervals for both spectra and significant wave heights are calculated from the Chi-squared distribution, following Young (1995). For spectra, the total degree of freedom (TDF) is calculated as $TDF = 2q$, and for significant wave height, TDF is found as described in ITTC (2017, p. 5). We use the mean absolute percentage error (MAPE) to compare model data with bow measurements. Systematic bias relative to observations is described with the mean percentage error (MPE).

Previous field measurements indicate that waves decay exponentially in ice (Squire and Moore, 1980; Wadhams et al., 1988; Marchenko, 2018). We have investigated the attenuation of the significant wave height by fitting decreasing exponentials on the shape $H_S = Ce^{-\alpha x}$, where $C$ and $\alpha$ are estimated parameters and $x$ is wave traveling distance through the MIZ, by means of non-linear least squares. From the fitted curves, the spatial damping coefficients $\alpha$, which describe wave attenuation per meter, are determined from:

$$\frac{\partial H_S}{\partial x} = -\alpha H_S. \quad [4]$$
2.2. Wave models
We have investigated the performance of the three wave prediction models; WAM-4, WW3 and WAM-3 in comparison to our observations. WAM-4 does not give wave parameter where there is assumed to be an ice cover present, whereas the two latter models contain wave attenuation through the ice cover. WW3 estimates power spectra, while the WAM models only provide integrated parameters.

The WAM-4 spectral wave forecast model is run operationally by the Norwegian Meteorological Institute (Carrasco and Gusdal, 2014). Spatial and temporal resolution of the model is 4 km and 1 hour, respectively, and it runs twice a day. A hard ice boundary based on satellite images (ice concentration larger than 30%) is defined in the model.

The WAM-3 model is a pan-Arctic wave forecasting system with an effective resolution of 3 km, operated in the framework of the Copernicus Marine Environmental Monitoring System’s Arctic Marine Forecasting Center. The WAM-3 wave hindcast used in the present study is forced by hourly 10 meter winds from ERA-Interim reanalysis (Dee et al., 2011) with a horizontal resolution of around 80 km. The wave spectra from the ERA-Interim are used at the lateral boundaries. The daily sea-ice concentration, ice thickness, and surface current information is taken from the TOPAZ ocean model system (Sakov et al., 2014). In areas with a sea-ice concentration larger than 20%, for the wave propagation, the frictional dissipation by the overlying ice sheet is considered as a function of the sea-ice thickness and wavenumber (Sutherland et al., 2019). Epsilon, which describes the thickness ratio of the two layers, is 0.13.

The WW3 model is a two-way coupled atmosphere-wave numerical weather prediction system in the Arctic. The atmosphere model AROME-Arctic (Müller et al. 2017) is coupled to the 3rd generation spectral wave model WW3 v5.16 using the OASIS3 model coupling toolkit. To obtain the sea-ice variables required by the wave model, we utilize the simple sea ice scheme (SICE) within SURFEX (Batrak et al., 2018). The WW3 setup used in this study uses the ST3 physics setting (equivalent to WAM4 physics) and estimates wave damping by sea ice through dissipation caused by bottom friction below a continuous thin elastic plate of ice (IC2 setting). We assume no scattering by sea ice in this framework. The coupled model is configured over the AROME-Arctic domain with a horizontal grid resolution of 2.5 km. AROME-Arctic uses 1 min. time steps while WW3 uses 5 min. time steps. The coupling frequency is every 30 min.

3. Results
Bow measurements are compared with model data interpolated to the location of the ship in Fig. 2(left). Valid measurements inside the MIZ, which we define as time periods where total UG saturation is below 10% and ship speed over ground (SOG) is below 0.5 m/s, are highlighted with gray background color. The whole comparison spans over three days. Error statistics are summarized in Table 2, where MAPE describes the mean absolute error and MPE describes the mean error. In general, there is a good agreement between observations and model. The WAM-4 model performs best outside the MIZ compared to the bow measurements with deviations less than 23% and 8% for \( H_S \) and \( T_m02 \) respectively. Deviations are approximately equally high when comparing to WW3, while WAM-3 overestimates both wave height (approx. 60%) and period (approx. 17%), especially the day before the ship ventured into the MIZ. Inside the MIZ, both models capture the larger trends when comparing them to the observations. \( H_S \) is actually better predicted when the measurements are considered not valid for both models. \( T_m02 \) is more sensitive to the motion of the ship and the models are closer to the observations within the valid times.
A stationary sea state over the measurement period inside the MIZ is an advantage in our analysis. The sea state is investigated in Fig. 2(right) where time series of \( H_S \), \( T_{m02} \) and wave direction from the models are presented. The model data are extracted from the ice edge defined in the respective models over the range of longitudes indicated in Fig. 1(left), and the standard deviation over this range is shown as shaded areas. None of the parameters change dramatically over the time period, and the data is overall quite consistent in space.

Table 2. Error statistics for \( H_S \) and periods \( T_{m02} \) when comparing model data with bow measurements.

<table>
<thead>
<tr>
<th>Error</th>
<th>WAM-4</th>
<th>WW3</th>
<th>WAM-3</th>
</tr>
</thead>
<tbody>
<tr>
<td>( H_S )</td>
<td>22.9</td>
<td>19.5</td>
<td>59.7</td>
</tr>
<tr>
<td>( T_{m02} )</td>
<td>7.8</td>
<td>11.4</td>
<td>17.1</td>
</tr>
<tr>
<td>( H_S )</td>
<td>107.5</td>
<td>58.2</td>
<td>85.7</td>
</tr>
<tr>
<td>( T_{m02} )</td>
<td>15.9</td>
<td>53.7</td>
<td>32.7</td>
</tr>
<tr>
<td>( H_S )</td>
<td>32.8</td>
<td>67.2</td>
<td>39.0</td>
</tr>
<tr>
<td>( T_{m02} )</td>
<td>3.2</td>
<td>32.7</td>
<td>29.9</td>
</tr>
</tbody>
</table>

Power spectra from bow measurements (black) and WII instruments (blue) with their respective 95% confidence intervals are compared to spectra from WW3 (magenta) and peak frequencies from WAM-3 (green) in Fig. 3. Stops into the MIZ are presented to the left, going
successively from stop number 1.1 (upper panel) to 1.4 (lower panel) and stops out of the MIZ.

The observed power spectra from bow measurements and WII instruments in Fig. 3(left) are consistent for most frequencies for all stops in the left figure, except for stop 1.3, where the spectra deviates substantially for frequencies higher than 0.1 Hz. See Løken et al. (2021) for a discussion on this discrepancy. The spectrum from WW3 underestimates the wave energy content at stop 1.1, is consistent up to approximately 0.125 Hz at stop 1.2 and overestimates the energy at stops 1.3-1.4. The peak frequencies from WAM-3 decrease further into the ice in accordance with the observations, except for stop 1.4 where it is overestimated.

In Fig. 3(right), WW3 overestimates the wave energy content at stops 2.1-2.3 furthest into the MIZ. The model predicts a dual peak spectrum with the low frequency peak in the same frequency range as the observed spectra, but no high frequent peak was detected in the bow measurements. For stops 2.4-2.6, the spectra are comparable in strength and frequency. WAM-3 predicts an increasing peak frequency, but the growth in $f_p$ is smaller than what was observed with the bow measurements.

**Figure 3.** Power spectra from bow measurements (black) and WII instruments (blue) with their respective 95% confidence intervals shaded. Spectra from WW3 (magenta) and peak frequencies from WAM-3 (green). Left: going into the ice zone from upper (stop 1.1) to lower panel (stop 1.4). Right: going out of the MIZ from upper (stop 2.1) to lower panel (stop 2.6).

It is evident that waves are attenuated as they travel through the MIZ. Figure 4 shows the decrease in $H_s$ from observations (with 95% confidence intervals) and the models as function of WTD through the MIZ on the way into (left) and out of (right) the ice. Decreasing exponentials are fitted to the data and the spatial damping coefficients are found from the fitted curves with Eq. 4. The damping coefficients from the different models and observations are summarized in Table 3. The bow measurements and the WII instruments are quite consistent. WAM-3 predicts a larger damping than observations into the MIZ (attenuation coefficient 64% higher than bow measurements), but is very close to observations out of the MIZ (attenuation coefficient 3% higher than bow measurements). WW3 substantially underestimates the damping compared to observations both into and out of the MIZ (attenuation coefficient 72-83% lower than bow measurements).
4. Discussion

The consistency in spectra between the bow and the WII instruments substantiates the validity of the wave measurements in the ice. The WTD through the MIZ is found from the mean wave direction and the location of the ice edge predicted by each model and therefore differs, as seen in Table 1. WAM-3 performs reasonably well compared to the observations in the MIZ when considering integrated parameters. Wave attenuation through the ice is satisfactorily modeled in WAM-3 with the two-layer model of Sutherland et al. (2019), although a bit overestimated in terms of the spatial damping coefficient compared to observations. This deviation can be partly explained by the fact that the WTD found from WAM-3 is considerably smaller than the WTD found from WW3, which in again is smaller than the WTD found from WAM-4. WW3 also gives fair estimates of the integrated parameters, especially outside the MIZ, but the spectra and the damping coefficient reveals that the thin elastic plate modeling of the sea ice so far fails to estimate the attenuation of the waves.

Data recorded while a ship is in motion will contain a Doppler shifted frequency according to the wave heading angle (Collins III et al., 2017), which could be problematic in the calculation of periods. In the present study, we observe that measured $T_m02$ is closer to the predicted value from WAM-3, which performs best of the models inside the MIZ, when the ship is stationary. Observations of $H_S$ on the other hand, seem to be less affected by the moving ship when compared to the models, the deviations are in fact larger between models and observations at the times considered not valid. This is in agreement with the results of Cristensen et al. (2013), which reported that peak and mean periods were strongly biased by the Doppler shift in the

Table 3. Spatial damping coefficients from observations and models into (stop 1) and out of (stop 2) the MIZ.

<table>
<thead>
<tr>
<th>$\alpha \ [10^{-5} \ m^{-1}]$</th>
<th>Bow</th>
<th>WII</th>
<th>WW3</th>
<th>WAM-3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stop 1</td>
<td>1.65</td>
<td>2.15</td>
<td>0.28</td>
<td>2.71</td>
</tr>
<tr>
<td>Stop 2</td>
<td>2.64</td>
<td>0.74</td>
<td>2.72</td>
<td></td>
</tr>
</tbody>
</table>

Figure 4. $H_S$ as function of WTD with fitted exponential decays from bow measurements (black), WII instruments (blue) with their respective 95% confidence intervals shaded, WW3 (magenta) and WAM-3 (green). Left: into the MIZ. Right: out of the MIZ.

Data recorded while a ship is in motion will contain a Doppler shifted frequency according to the wave heading angle (Collins III et al., 2017), which could be problematic in the calculation of periods. In the present study, we observe that measured $T_m02$ is closer to the predicted value from WAM-3, which performs best of the models inside the MIZ, when the ship is stationary. Observations of $H_S$ on the other hand, seem to be less affected by the moving ship when compared to the models, the deviations are in fact larger between models and observations at the times considered not valid. This is in agreement with the results of Cristensen et al. (2013), which reported that peak and mean periods were strongly biased by the Doppler shift in the
time series, while the estimates of significant wave heights were reasonable since they only depend on the sea surface variance.

5. Conclusions
We have presented shipborne wave measurements from the MIZ with a system combining an altimeter (UG) and a motion correction device (IMU). Our setup provides single point time series of ocean surface elevation outside of and inside the MIZ, which enables us to produce 1D power spectra and integrated parameters.

Observations have been compared with estimates from the spectral models WAM-4 (integrated parameters) in the open ocean, WW3 (integrated parameters and spectra) and WAM-3 (integrated parameters) in the open ocean and in the MIZ. We have found good agreement in zero up-crossing periods and significant wave height outside the MIZ. Predictions from WAM-4 and WW3 deviate with less than 23% over this timespan while $H_S$ predicted by WAM-3 is up to 60% off. From the time series, it is clear that the two models which predict wave attenuation through ice are able to estimate the trends for $H_S$ and $T_{m02}$ inside the MIZ, although the errors with respect to the observations are larger here. Observed $H_S$ actually matches the model prediction better when the measurement conditions are considered not valid, i.e. either when the altimeter measurement range was exceeded or when the ship was cruising. Measured $T_{m02}$ matches model predictions better during the times considered valid, most likely due to the Doppler shift induced in the time series during cruising.

Both observations and models show an exponential decay in $H_S$ through the MIZ. From the spectra and the spatial attenuation coefficients, we can conclude that WW3 underestimates the wave damping through the ice. The attenuation modeling of WW3 predicts coefficients 72-83% smaller than observations, and further tuning is therefore necessary to better estimate wave parameters in the ice. Spectra were not available from WAM-3, but the attenuation coefficients from this model match observations better and are 3-64% larger than bow measurements.

Acknowledgments
The authors are grateful to Øyvind Breivik for inviting us to the cruise and to the crew of RV Kronprins Haakon for their assistance. The Nansen Legacy project helped funding the campaign. Funding for the experiment was provided by the Research Council of Norway under the PETROMAKS2 scheme (project DOFI, Grant number 28062). The data are available from the corresponding author upon request.

References


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Paper III

Bringing optical fluid motion analysis to the field: a methodology using an open source ROV as camera system and rising bubbles as tracers

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BRINGING OPTICAL FLUID MOTION ANALYSIS TO THE FIELD: A METHODOLOGY USING AN OPEN SOURCE ROV AS CAMERA SYSTEM AND RISING BUBBLES AS TRACERS

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ABSTRACT

Detailed water kinematics are important for understanding atmosphere-ice-ocean energy transfer processes in the Arctic. There are few in situ observations of 2D velocity fields in the marginal ice zone. Particle tracking velocimetry and particle image velocimetry are well known laboratory techniques for measuring 2D velocity fields, but they usually rely on fragile equipment and pollutive plastic tracers. Therefore, in order to bring these methods to the field, we have developed a new system which combines a compact open-source remotely operated vehicle as an imaging device, and air bubbles as tracing particles. The data obtained can then be analyzed using image processing techniques tuned for field measurements in the polar regions. The properties of the generated bubbles, such as the relation between terminal velocity and diameter, have been investigated under controlled conditions. The accuracy and the spread of the velocity measurements have been quantified in a wave tank and compared with theoretical solutions. Horizontal velocity components under periodic waves were measured within the order of 10% accuracy. The deviation from theoretical solutions is attributed to the bubble inertia due to the accelerated flow. We include an example from an Arctic field expedition where the system was deployed and successfully tested from an ice floe. This work is an important milestone towards performing detailed 2D flow measurements under the ice in the Arctic, which we anticipate will help perform much needed direct observations of the dynamics happening under sea ice.

1 Introduction

The development of field studies of fluid flow in the ocean has increased considerably in the last decades thanks to the technological advances within measurement equipment. Nonetheless, working in the Arctic environment remains yet difficult. Not only because of the technical aspects to execute a successful controlled experiment in the field, but also because the possibilities to use regular laboratory equipment are limited by the harsh temperatures and conditions. On top of that, working on an ice layer brings new challenges to the design of a field setup. Adaptation of well known techniques to the field by accurately testing them in the laboratory first is a reasonable approach. This paper introduces a new system for high resolution velocity measurements of fluid flow specially designed for Arctic conditions, where elements from already established methods are combined. The accuracy of the instrumentation is determined to understand its capabilities in order to adapt them to the desired object of study.

Although laboratory experiments can yield useful results due to the high level of control over variables, scaling between wave tank conditions and the real ocean is challenging as several nondimensional numbers are necessary to describe the...
problem. In wave tank experiments, it is for example desirable to moderate nonlinear effects and ensure deep-water waves, as this is the most common situation in the Arctic. These conditions are achieved by respectively reducing the wave steepness and the ratio of wavelength over depth. As pointed out by [1], it is with such constraints difficult to simultaneously obtain a realistic Reynolds number for turbulence under ocean waves. It is therefore necessary to perform field measurements to complement or confirm laboratory experiments, where scaling problems are usually inevitable.

Ultrasonic velocimeters, such as acoustic Doppler velocimeters (ADVs) and acoustic Doppler current profilers (ADCPs) are well known instruments for fluid velocity measurements in the ocean. However, these instruments provide only single point velocity (ADVs), or in the best case an array of along-beam velocities (ADCPs), which in many cases is insufficient to give a detailed description of the dynamics of certain flow phenomena that occur in the Arctic. An example of such a phenomenon is wave propagation underneath an ice layer [2, 3], which is interesting for validation of wave attenuation models through the marginal ice zone, e.g. [4, 5, 6]. Another example is the investigation of turbulent structures or jets induced by colliding ice floes [1], which may contribute to ice-ocean energy transfer and increased eddy viscosity and viscous dissipation. In these cases it is desirable to resolve a 2D velocity field for a better understanding of the underlying physics.

Within fluid dynamics and wave studies in particular, researchers have used two very established laboratory techniques to understand flow dynamics: particle image velocimetry (PIV) and particle tracking velocimetry (PTV). In controlled environments, these techniques can accurately represent the velocity fields in a wide range of flows. Conventional PIV and PTV systems normally utilize a powerful light source, such as lasers, to illuminate a thin sheet of tracer particles. In this way it can be ensured that the particles of interest are situated in a plane with a fixed distance to the camera. In later years it has been shown that these techniques can be applicable in field measurements with certain reserves, PIV-laser systems have been developed and deployed off docks and from ships [7, 8]. However, it is difficult to arrange such a setup in the field, especially in an Arctic environment, due to the large weight, high cost and fragility of lasers.

Another complication in the use of traditional PIV and PTV in field measurements is the use of tracers. As addressed in [7], some oceanic areas can be analyzed by the use of the natural suspended particles. But, particularly in the Arctic Ocean, the water is mostly clear during the winter. Therefore, the introduction of environmental-friendly tracers is required. The use of passive tracers is problematic in field applications because the residual water motion, which is always present due to tides and currents even though the conditions appear calm, will transport the tracers away from the measurement plane. It is very challenging to maintain an acceptable concentration of particles and at the same time avoid intrusive mixing motion. The adaptation of PIV using bubbles as tracers has become popular throughout different applications [9, 10, 11, 12, 13], and tested against more commonly seeding methods [14]. The mentioned studies have brought to light different aspects that need to be calibrated and considered to apply these techniques outside a controlled environment. In this study, an array of small air bubbles has been introduced as tracers, which provide a plane of particles and thus enables us to find a two-dimensional velocity field without the need of a light sheet.

A good way to assist underwater measurements is the use of ROVs (remotely operated vehicles). Since the early 1980s, the interest of utilizing ROVs for field work has increased. Towards the end of that decade, the first major field works using ROVs to retrieve information and samples were made [15]. Afterwards, ROVs have been used extensively for observations and sample retrieving. By using an open source ROV, it is possible to control tools underwater and retrieve images and information from different sensors, while keeping costs down to a reasonable level. It is only in the last few years that ROVs have been considered a tool in different fluid dynamics applications and in situ measurements, where the reliability and limitations of the sensors mounted remain the main issues [16, 17].

In this work, a methodology to perform reliable field measurements using PIV techniques is described. Extensive laboratory tests have been included to show the reliability and range of confidence of the bubble curtain as a substitute for passive tracers and light sheets. Two type of experiments were performed under controlled conditions in the laboratory; one to evaluate the terminal velocity and the maximum oscillating velocity of the bubbles as function of the bubble size, and the other to investigate the effect of accelerated flow on the bubble slip velocity. PTV was used in these experiments to track and investigate the motion of each individual bubble. The use of an ROV as an imaging and light source system in PIV field measurements has been introduced and tested in extreme Arctic conditions. The field experiments were performed and analyzed independently of the laboratory results. The laboratory results were not used to calibrate the field experiments as the conditions were very different, but as general observations of the system to as good as possible estimate some sources of error and the accuracy of the method. For simplicity, we introduce the abbreviation PV (particle velocimetry) to address PTV, which was used in the laboratory, and PIV, which was used in the field, and name the introduced technique "ROV-PV system".

The paper is organized in the following manner. Section 2 describes the components and the assembly of the ROV-PV system, the algorithms used for image processing and how fluid velocity is interpreted from measured bubble velocity.
Section 3 presents the results from wave tank experiments. An example of a field deployment in the Arctic is included in Section 4. The methods and results are discussed in Section 5 and conclusions are drawn in Section 6.

2 Data and methods

The ROV-PV system presented here is designed to capture dynamics generated from air-water and ice-water interactions in the upper ocean. It is a portable system built for Arctic field campaigns at remote locations and harsh environments. It is emphasized that the accuracy of the methodology is sub-optimal for laboratory application standards. The system is developed for field use where applied measurement techniques usually have a much higher tolerance for error. The core concept is to perform optical recordings with the camera of an ROV. A plane of rising air bubbles illuminated by the ROV’s headlights are used as tracing particles. The bubbles rise inside a thin aluminum frame with indicated reference coordinates. As it is difficult to maintain a constant relative position between the ROV and the aluminum frame, all images are calibrated individually to obtain real world units. Image processing is applied to produce the 2D in-plane velocity field of the bubbles. During post-processing, the vertical buoyancy driven velocity component of the bubbles is subtracted to obtain the water velocity. The same procedure also applies for the laboratory experiments which are carried out to validate some aspects of the method, except that the relative position between the camera and the measurement plane is constant in these situations, and the same coordinate transformation can be used for all images.

2.1 ROV-PV system

An open source ROV has been chosen for image acquisition that will be used for the PV analysis, as it is easily maneuvered below and around sea ice. The use of open source instruments is an increasing trend in geophysics, see e.g. [18, 19, 20]. Open source instruments provide flexibility in the sense that they offer the possibilities for extensive modification to specific needs and easy interfacing with other open source systems. Another advantage is the significantly lower cost than similar closed source solutions. An ROV allows for preliminary inspections of the site to find the most suitable location for measurements, as opposed to a stationary camera. The high-performance BlueROV2 from BlueRobotics [21] with open source software and electronics has been used. It is installed with eight thrusters to increase the stability and obtain full six degrees of freedom control. The ROV is physically connected to a field computer by a tether, which carries and transfers video and data signal. An open source application called QGroundControl (QGC) works as the user interface and provides live video stream and various types of information for the pilot.

The camera is a wide-angle low-light HD USB camera with a 2.97 mm focal length lens with low distortion (1%). It is mounted on a servo motor for user control of tilt angle. The digital chip has a 1920×1080 pixel resolution. The highest available frame rate \( \Delta f = 30 \) frames/s is used for maximum temporal resolution. Camera settings such as exposure, brightness and gain are adjusted in QGC to optimize the images in different environments and conditions, to ensure that the bubbles appear as clear, circular particles and not as blurry streaks. The ROV is equipped with four controllable headlights with a total capacity of 6000 lumens.

In the setup, which is illustrated in Fig. 1, bubbles are generated with a 5 m long flexible silicon rubber hose. At the one end, the hose is fed with air of approximately 0.5 bar from a 1.1 kW ABAC compressor. At the other end, the last 0.75 m is perforated every 1 cm on the upward facing side with a 0.3 mm needle and the tube is sealed with a plug to prevent air leakage. This configuration provides an array of relatively small bubbles with a diameter of 0.6 – 1.4 mm. Bubbles have also been generated with a thin carbon fiber pipe perforated with a 0.1 mm drill. This device was not used in the experiments because the bubbles grew too large while sticking to the pipe before they detached. The headlights of the ROV have proven sufficient to make the bubbles visible during fieldwork. During laboratory experiments, the ROV has been placed outside of the water tank due to space limitations, and the bubbles have been illuminated with external LED lamps to avoid headlight reflections in the glass wall.

In order to obtain quantitative physical information from the experiments, it is necessary to implement a coordinate system to be able to convert from pixels to real-world units in the post processing. The coordinate system needs to be in the framework of the bubble plane. A thin aluminum grid with a combination of grid bars and woven string has been used as coordinate indicators. The grid measures 55×45 cm and the coordinate resolution is 8 – 12 cm in the horizontal and 9 cm in the vertical direction. In a traditional particle tracking setup, the camera and the light-sheet are fixed in one position throughout the whole experiment, making it easy to take a picture of a coordinate system in advance of the measurements. In the current system, the ROV will always be in some movement which leads to a varying position and angle relative to the coordinate grid. Therefore, the perforated hose is integrated at the bottom part of the grid so that each single image can be converted into real-world units with a third order coordinate transformation, e.g. [22], meaning that the mapping function of the calibration is unique for every image. Upon deployment in the ocean, the
perforated hose and coordinate system are suspended from a constructed frame or from the sea ice. Heavy bolts are attached to the bottom of the grid to keep it vertical and in place when exposed to waves and ocean currents.

2.2 PTV and PIV methodology

In this work, PTV and PIV were used in laboratory and field observations, respectively. PTV differs from PIV in a fundamental way. PIV relies on pattern matching in an essentially Eulerian way, whereas PTV seeks to identify individual particles and follow them in a Lagrangian sense. Therefore, it is important to choose the most suitable technique based on the characteristics of the experiment, and on the information which is desirable to obtain. A small review of the PTV and PIV algorithms used for this work follows.

DigiFlow Software [23] has been utilized for PTV in this study. After the image acquisition, the image is scanned for blobs that have an intensity satisfying some threshold parameters. If a blob is found, its characteristics are determined and compared with a set of requirements for the blob to be considered a particle. If the blob satisfies these requirements, it is recorded as a particle, if it does not, it is discarded. Once all the particles in an image have been found at $t = t_n + 1$, they need to be related back to the previous image $t = t_n$. DigiFlow uses a modification of what is known in operations research as the Transportation Algorithm, which was developed by [24]. The algorithm chooses a set of particles $P$ at $t = t_n$, and the other the set of particles $Q$ at $t = t_n + 1$. At $t = t_n$, $p_i$ for $i = 1, 2, ..., M$ represents the set of particles, while at $t = t_n + 1$, they are labeled $q_j$ for $j = 1, 2, ..., N$. Each $p_i$ or $q_j$ contains the location of the particle and other characteristics such as size, shape, intensity, or any other desired information. A set of association variables $\alpha_{ij}$ is defined. When $\alpha_{ij} = 1$ then $p_i$ at $t = t_n$ is the same particle as $q_j$ at $t = t_n + 1$. If $\alpha_{ij} = 0$, then $p_i$ and $q_j$ represent different physical particles. The number of particles in the images may be different at $t = t_n$ and $t = t_n + 1$. To overcome this problem, $\alpha_{ij}$ and $\alpha_{0j}$ are defined as dummy particles at times $t = t_n$ and $t = t_n + 1$, respectively. Unlike ordinary particles, more than one value of $j$ or $i$ may give a nonzero value of $\alpha_{0j}$ and $\alpha_{i0}$, respectively. In this case, a nonzero value of $\alpha_{0j}$ indicates that particle $p_i$ at $t = t_n$ has been lost from the image by $t = t_n + 1$, either by moving out of the image or for some other reason. Similarly, $\alpha_{ij} = 1$ represents a particle $q_j$ present at $t = t_n + 1$ which was not there at $t = t_n$.

The in-house HydrolabPIV software developed at the University of Oslo has been used for PIV in this work [22]. This software combines different techniques presented in previous works to optimize the PIV algorithm [25, 26, 27, 28]. Given two consecutive images of a seeded flow, it is desirable to find an Eulerian description of the velocity field. To achieve this, the images are divided into a regular grid of subwindows, usually of size from $8 \times 8$ to $64 \times 64$ pixels. The use of a certain percentage of overlap is also common, usually between 25% and 75%. For each subwindow in the
first frame \( I_1 \), a subwindow \( I_2 \) with a similar pattern is searched in the second frame. A metric of the similarity of the patterns is required, a common choice is the normalized cross-correlation

\[
R_{ncc} = \frac{\sum I_1 I_2 - \sum I_1 \sum I_2}{\sqrt{\left(\sum I_1^2 - \frac{\left(\sum I_1\right)^2}{N}\right) \left(\sum I_2^2 - \frac{\left(\sum I_2\right)^2}{N}\right)}}.
\]

To speed up the processing, fast Fourier transform is often used to calculate the cross-correlation. An ensemble averaged velocity is found by dividing the optimal match displacement of the pattern by \( \Delta t \). After the velocity is found, the result can be made more robust by checking the quality of the image, and further validation of the vectors using outlier detection. This detection uses a \( 3 \times 3 \) normalized local median filter. In addition, a cubic B-spline is fitted to the velocity field using an iterative weighted least squares fit. The local residuals are used with the biweight function in the first iteration. The fit is then iterated a few times where the residuals are used to update the weights in the least squares fit. The normalized residuals from the last iteration together with the residuals from the local median filter are used to mark the outliers. These outliers can be removed and replaced by using the fitted B-spline.

2.3 Air bubbles as tracers

Bubbles as tracing particles are challenging because they do not passively follow the fluid motion as opposed to conventional tracers. The buoyancy driven motion of the bubbles must be subtracted in order to achieve the absolute water velocity. In the following, the \( x = (x, y) \), \( V = (u, v) \) and \( F = (F_x, F_y) \) conventions will be used for 2D position, velocity and force, respectively. The objective of this study is to introduce a technique where the velocity of bubbles \( V_b \) is measured in order to determine the velocity of the water \( V_w \). An important quantity in this method is the slip velocity \( V_{\text{slip}} \) which is the relative velocity between the water and the bubbles, defined as

\[
V_{\text{slip}} = V_b - V_w.
\]

Small bubbles are beneficial because they have rectilinear vertical trajectories, whereas larger bubbles tend to oscillate in the horizontal direction [29], which raises the need for further velocity corrections. It is additionally advantageous to generate uniform bubbles along the array, which will allow for subtraction of the mean buoyant velocity across the entire velocity field. As pointed out by [30], it is difficult to produce small bubbles of uniform size and rate. In the present study, the variability in bubble diameter was large over the span of the perforated hose. From visual inspection, the bubble diameter seemed to be randomly distributed in size. It is likely that the difference in bubble size was caused by the non-uniform size of the outlet holes, since the silicon rubber material may have contracted differently from hole to hole after the needle was subtracted.

For steady uniform water flow, bubbles will approximately follow the water motion in the horizontal direction after a relaxation time. An additional vertical component is present due to the buoyancy force \( B = \frac{1}{2}(\rho_w - \rho_b)\pi R^3 g \), where \( \rho_w \) denotes the water density, \( \rho_b \) the bubble density, \( g \) the acceleration due to gravity and \( R \) the bubble radius. In this case, \( V_{\text{slip}} = v_{\text{slip}} \) only has a vertical component driven by \( B = B_v \). On the other hand, \( V_{\text{slip}} \) also depends on the bubble inertia when the water flow \( V_w(x, t) \), where \( t \) denotes time, is accelerated. Since \( \frac{\rho_w}{\rho_b} \ll 1 \), the effect of added mass becomes important to determine the inertia. Drag force on a spherical bubble is expressed as \( D = \frac{1}{2} \rho_w \pi R^2 C_D V_{\text{slip}} |V_{\text{slip}}| \). The drag coefficient \( C_D \) depends on the Reynolds number \( Re = \frac{2R \sqrt{\nu v}}{\sqrt{D}} \), where \( \nu \) denotes the water kinematic viscosity. Following [31], an estimate of the slip velocity can be obtained from the momentum balance of a bubble. Their Eqs. (2) are modified to include the added mass effect to obtain

\[
D(V_{\text{slip}}) - B \approx -M_b \frac{\partial V_w}{\partial t} |_{\text{max}},
\]

where \( M_b = \frac{1}{2} \rho_w \frac{4}{3} \pi R^3 \) is the added mass of a spherical bubble, and the mass of the bubble is neglected. Equation (3) contains the leading order terms in Eq. (43) of Maxey and Riley [32]. The slip acceleration \( \frac{\partial V_{\text{slip}}}{\partial t} |_{\text{max}} \) is conservatively approximated with the maximum water acceleration \( \frac{\partial V_w}{\partial t} |_{\text{max}} \). The estimate for \( V_{\text{slip}} \) presented in Eq. (3) depends on the relation between \( C_D \) and \( Re \) on and on the maximum water acceleration. In Section 3, the horizontal velocity component under the crests of water waves, where there exist analytical approximations for the acceleration, is investigated. An analytical relation for \( C_D(Re) \) is used to solve Eq. (3) for \( V_{\text{slip}} \) and it is shown that the observations are in agreement with the error analysis.
Analytical solutions exist for $C_D(Re)$ for spherical bubbles moving in infinite mediums at low Reynolds numbers. Stokes’ solution, i.e. $C_D = \frac{24}{Re}$ is a good approximation for $Re < 1$ [29]. For moderate Reynolds numbers, boundary-layer theory approximations have resulted in analytical solutions. Batchelor assumed an irrotational flow outside a thin boundary layer around spherical bubbles and use potential theory to obtain $C_D = \frac{48}{Re}$ [33], which is a fair estimate for $20 < Re < 200−500$, depending on the fluid. Bubbles tend to stay spherical up to and including moderate Reynolds numbers due to sufficiently strong surface tension forces. In the spherical regime, the bubble trajectory is usually rectilinear and the vertical velocity increases with the radius squared. Above $Re \approx 450$, bubbles enter the elliptical regime where they start to deform and flatten as viscous and hydrodynamic forces become more prominent. Consequently, the drag increases, and approximations of $C_D$ primarily rely on numerical simulations and experimental data. In the ellipsoidal regime, the rising motion is typically helical or oscillatory and the terminal velocity has been observed to increase more slowly or even decrease with radius [34]. Haberman and Morton reported a smooth transition in vertical velocity between the spherical and the elliptical regime in fresh water [29]. For $Re > 5000$, bubbles normally take form as spherical caps, but this region is not relevant for the present study.

The terminal velocity $v_T$ of a bubble can be found theoretically by balancing the drag and the buoyancy force in the vertical direction. For a comparison with the observed bubble terminal velocity presented next, the analytical relation for $C_D(Re)$ of [33] is used, i.e. $C_D = \frac{48}{Re}$. This gives a drag force $D_y = 12\mu \pi R v_T$. The force balance is rearranged and solved for the terminal velocity

$$v_T = \frac{(\rho_w - \rho_b) g R^2}{9\mu w}.$$ (4)

The relation between the size and the rising velocity of bubbles produced by the perforated hose has been investigated in the Hydrodynamical Laboratory at the University of Oslo. The experiments were performed in a small glass tank measuring 1.5 m long and 0.4 m wide with water depth $h = 27$ cm. The perforated hose was placed on the bottom, parallel to the long wall. In this experiment, the motion of the rising bubbles was of interest, and not the ROV-PV system as a whole. Therefore, the ROV camera was substituted with a Photron FASTCAM high speed camera for increased accuracy. Fresh water was used in the first case and salt water with salinity equal to typical Arctic conditions was used in the second case. An LED lamp placed on the back side of the tank illuminated the bubbles. An integrated laboratory Kaeser compressor produced a stable airflow to the perforated hose. A schematic of the experimental setup is shown in Fig. 2. The water temperature was 18°C. Other fluid parameters are summarized in Table 1.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Fresh water</th>
<th>Salt water</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density $[kg/m^3]$</td>
<td>$\rho_w = 998$</td>
<td>$\rho_w = 1026$</td>
</tr>
<tr>
<td>Dynamic viscosity $[Pa \ s]$</td>
<td>$\mu_w = 1.027 \times 10^{-3}$</td>
<td>$\mu_w = 1.103 \times 10^{-3}$</td>
</tr>
</tbody>
</table>

Table 1: Parameters used in the experiments. These values are used for calculating the theoretical terminal velocity in both fresh and salt water.

The image resolution was $1024 \times 1024$ pixels. The frame rate and shutter speed were set to 500 frames/s and 1/9000 s, respectively. The camera was located outside the tank, 15 cm from the tank wall. The field of view (FOV) for the
validation was $40.1 \times 40.1$ mm in fresh water and $48.8 \times 48.8$ mm in salt water. Before the experiments were initiated, a Perspex plate marked with reference coordinates was placed vertically on the same location as the bubble plane and recorded with the camera. The reference coordinates had a horizontal and vertical resolution of 1 cm. From the reference coordinates, pixel coordinates were converted to real world coordinates for all the images with a linear transform [22], which should suffice since the camera axis was perpendicular to the measurement plane. To be certain that the terminal velocity was reached, the center of FOV was placed 18 cm above the tank bottom. The vertical displacement of a bubble between two frames varied $5 - 20$ pixels, depending on the bubble speed.

Throughout the experiments, bubbles interacted with each other and in some situations, the tracking algorithm mistook two bubbles for one. These bubbles were filtered out. Additionally, bubbles which entered an elliptical regime were removed in order to compare the results with theoretical solutions for spherical bubbles. Elliptical bubbles and misinterpretations were filtered out by applying a ratio parameter $\beta = \frac{\text{vertical radius}}{\text{horizontal radius}}$. $\beta$ contains information on the bubble shape, e.g. $\beta \approx 1$ implies a spherical shape and $\beta < 1$ implies a horizontal elliptic shape. Table 2 lists the threshold values of $\beta$ used in the validation. The horizontal and vertical bubble diameters were determined with 4 pixel accuracy, which corresponds to an error up to approximately 14%. Figure 3 shows an example of a raw image where the bubbles tracked by the PTV algorithm are indicated.

The instantaneous horizontal ($u_i$) and vertical ($v_i$) bubble velocities were calculated from the position $(x_i, y_i)$ for each frame $i$ a bubble was tracked. Here, $i = 1, 2, ..., N - 1$, where $N$ is the number of frames, typically $50 - 200$ depending on the bubble speed. The minimum, mean and maximum horizontal and vertical velocity were found for each bubble. The standard deviation of vertical velocity ($\sigma_v$) and horizontal radius ($\sigma_r$) for each bubble were used as indicators for consistency, and bubbles which did not satisfy the quality parameters listed in Table 2 were not included in the analysis. The bubbles included in the results were very consistent with $\sigma_v$ and $\sigma_r$ both in the order of $10^{-3}$ in fresh and salt water. This indicates that the vertical velocity was quite constant and that the terminal velocity was reached.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Fresh water</th>
<th>Salt water</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ratio</td>
<td>$0.6 \leq \beta \leq 1.0$</td>
<td>$0.6 \leq \beta \leq 1.0$</td>
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<tr>
<td>$\sigma_r$ [mm]</td>
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<td>$\sigma_r \leq 0.015$</td>
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<tr>
<td>$\sigma_v$ [ms$^{-1}$]</td>
<td>$\sigma_v \leq 0.007$</td>
<td>$\sigma_v \leq 0.01$</td>
</tr>
</tbody>
</table>

Table 2: Threshold parameters applied when processing the data from the bubble terminal velocity experiments.

Figure 4 presents the tracked terminal velocity of the rising bubbles (black dots) in fresh water (left panel) and salt water (right panel). A logarithmic increase in terminal velocity with respect to the bubble diameter can be observed. A
logarithmic curve fit has been applied to the terminal velocity in both stagnant fluids by means of linear least squares (red line). Tracked maximum and minimum absolute horizontal velocities are displayed with purple and blue dots, respectively. There is a slight increase in maximum absolute horizontal velocity with respect to bubble diameter in the case of fresh water. This increase is more prominent in salt water. The theoretical terminal velocities calculated from Eq. (4) with values listed in Table 1 and $\rho_b = 1.2 \text{ kg/m}^3$ are displayed as blue lines.

![Figure 4: Measured bubble terminal velocity vs diameter (black dots) from experiments in fresh water (left panel) and salt water (right panel) at 18°C. Only bubbles satisfying the quality criteria in Table 2 are included. The theoretical model of [33] (blue line) is a fair estimate up to $Re \approx 300$ for fresh water and $Re \approx 500$ for salt water. Reynolds numbers of the tracked bubbles are indicated. The experimental results of [35] (Fig. 15, curve I) are included for comparison with the fresh water experiment (black circles) and agree well with the current observations. A logarithmic curve fit has been applied to the tracked terminal velocity (red line), and is used in Section 3 as a relation between bubble terminal velocity and diameter. Blue dots show minimum and purple dots show maximum tracked absolute horizontal velocity.](image)

Aybers and Tapucu conducted a similar study on the terminal velocity of rising bubbles in stagnant water [35]. Their results (Fig. 15, curve I), included in Fig. 4 (left panel) as black circles for comparison, are in agreement with the present findings. Aybers and Tapucu found that the bubbles entered the ellipsoidal regime for $Re \approx 530$, where the terminal velocity started to decrease [35]. Whether the present bubbles enter the ellipsoidal regime in fresh water is not clear from visual inspections, the change in regime may occur at a higher Reynolds number than investigated here. The salt water bubbles possibly enter the ellipsoidal regime above $Re \approx 540$, where we observe the terminal velocity to flatten out. The theoretical terminal velocity from Eq. (4) is a fair estimate up to $Re \approx 300$ for fresh water and $Re \approx 500$ for salt water, although the curve shape is quite different, most likely because the bubbles investigated are in the transition between the spherical and ellipsoidal regime.

### 3 Validation in a wave tank

The ROV-PV system was evaluated under propagating periodic surface waves. A comparison to theoretical solutions to determine the accuracy of the measurements follows. The experiments were carried out in a wave tank in the Hydrodynamical Laboratory at the University of Oslo.

Plane progressive waves are characterized by the acceleration of gravity $g$, the angular wave frequency $\omega$, the amplitude $a$, the wave number $k$ and the corresponding wavelength $\lambda$, where $\omega = 2\pi f$ and $\lambda = 2\pi/k$. In the case of $h > \frac{1}{2}\lambda$, the effects of the bottom are negligible. Periodic waves with steepness $ak \ll 1$ can be well approximated by third order Stokes theory, where the velocity potential $\phi$ is given by

$$\phi = \frac{ag}{\omega} e^{k y} \sin(kx - \omega t) + O(a^4),$$  

\[5\]
where \( x \) denotes the horizontal axis pointing in the direction of wave propagation and \( y \) the vertical axis pointing upwards with \( y = 0 \) as the undisturbed free surface [36]. Equation (5) is accurate up to and including terms of \( O(a^3) \) as long as the second order dispersion relation

\[ \omega^2 = gk(1 + a^2k^2) + O(a^3), \]

which describes the relation between \( \omega \) and \( k \), is applied [36]. From Eq. (5), the following terms for horizontal and vertical velocity components, respectively, are obtained up to and including third order

\[ u_w = \frac{\partial \phi}{\partial x} = \frac{gak}{\omega} e^{kx} \cos(kx - \omega t), \]

\[ v_w = \frac{\partial \phi}{\partial y} = \frac{gak}{\omega} e^{kx} \sin(kx - \omega t). \]

Following [37], the velocities are nondimensionalized with \( gak/\omega \) for the comparison of waves with different amplitudes. The velocities in vertical columns directly underneath the wave crests are investigated in the further analysis. Here, the phase functions are \( kx - \omega t = 2\pi n \), where \( n \) is any integer. The nondimensional form of Eqs. (7)–(8) evaluated under the wave crest reduce to \( u_w/(gak/\omega) = e^{ky} \) and \( v_w/(gak/\omega) = 0 \), respectively. In the following, the non-zero horizontal velocity component is investigated.

In Section 2.3, it was shown from the momentum equation that the slip velocity is related to the bubble inertia due to added mass. Here, it is attempted to estimate \( u_{slip} \) by evaluating the momentum equation in the horizontal direction. As a wave undergoes an entire period, the maximum horizontal water acceleration up to and including third order terms is

\[ \left[ \frac{\partial u_w}{\partial t} \right]_{max} = gako e^{ky}. \]

It is assumed that viscosity controls fluid resistance to bubble motion, due to the small bubble size and horizontal slip velocity. Therefore, the drag force \( D_x \) acting on the bubbles is approximated with the Stokes drag, i.e. \( D_x = 6\rho_wv\pi Ru_{slip} \). By applying the Stokes drag approximation, a linear relation between the drag force and the slip velocity is obtained. It should therefore be reasonable to perform a decomposition of the drag force into the vertical and horizontal direction. Equation (9) and the drag force approximation are inserted into Eq. (3), which is rearranged to estimate the slip velocity

\[ u_{slip} \approx -\frac{R^2}{9\nu} \left[ \frac{\partial u_w}{\partial t} \right]_{max} = -\frac{R^2}{9\nu} gako e^{ky}, \]

where the buoyancy term is neglected, as we are only looking at the horizontal component. Equation (7) and (10) are inserted into Eq. (2) to obtain an estimate for the horizontal bubble velocity

\[ u_b \approx \frac{gak}{\omega} e^{ky} \left( 1 - \frac{\omega R^2}{9\nu} \right). \]

Figure 5 shows a schematic drawing of the experimental setup in the wave tank. It measures 24.6 m long and 0.5 m wide and was filled with fresh water to a depth \( h = 0.6 \) m. Waves were generated in one end of the tank with a computer controlled hydraulic piston wave maker. The perforated hose was centered on the bottom of the tank in the span-wise direction 11 m from the wave maker. Two powerful LED lamps were suspended from the tank walls and angled downwards to illuminate the bubbles from the sides. The ROV was placed outside the tank with the camera axis perpendicular to the bubble plane. The camera lens was located 30 cm from the tank wall, meaning that the distance between camera and bubbles was approximately 55 cm. The FOV in the tank center was approximately 80×45 cm. Before the experiments were initiated, a Perspex plate marked with reference coordinates was placed vertically on the same location as the bubble plane and recorded with the camera. The reference coordinates, 88 in total, had a horizontal resolution of 5 cm and a vertical resolution of 2 cm close to the surface and 5 cm further down in the depth. For all images in the experiment, pixel coordinates were converted to real world coordinates with a cubic transform [22] from the reference coordinates. Fluid motion was analyzed within a section of 45 cm in the span-wise direction and 30 cm in the vertical direction, which is indicated in Fig. 7. This section was well within the domain of the pixel-to-world
Figure 5: Schematic drawing of the experimental setup in the wave tank. A coordinate system is defined with the \((x, y)\) axes aligned horizontally in the direction of wave propagation and vertically in upward direction, respectively, with \(y = 0\) as the undisturbed free surface. The velocity under the waves is estimated from the motion of the bubbles, which is recorded with the ROV camera.

Figure 6: Time series of surface elevation \(\eta\) obtained with ultrasonic gauges for 1.4 Hz monochromatic waves. The amplitude was approximately 7, 15 and 23 mm for top, middle and bottom panel, respectively. The shaded area marks the 10 periods investigated, which were periodic, almost constant in amplitude and unaffected by any reflected waves from the beach.

transformation. The distortion effect of the camera, which is most prominent in the outer edges, was also reduced in the center of the FOV.

A series of about 70 monochromatic waves was generated in each run. After a short transient build-up, the wave train became periodic and this is the part included in the analysis. An absorbing beach damped the waves at the far end of the tank and the analysis was terminated before any reflected waves reached back to the FOV. Figure 6 shows the surface elevation \(\eta\) in the position of the ROV with respect to time for 1.4 Hz waves and three different amplitudes. The 10 investigated periods are highlighted. Surface elevation at the test section of the tank was measured with an array of three ULS Advanced Ultrasonic wave Gauges (UGs) from Ultralab. The UG situated directly above the ROV was mainly used, and the two others were applied as redundancy in case the first one failed, which happened in a few runs. The UGs sampled at 250 Hz and they have a technical resolution of 0.18 mm.

Wave amplitude \(a\) from each run was determined as the mean value of \(\eta_{\text{max}}\) over the 10 periods investigated. Three different wave frequencies \(f = 1.4, 1.6\) and \(1.8\) Hz, and three different amplitudes of approximately 7, 15 and 23 mm were investigated. Each combination was repeated three times, which means 27 runs in total. The wavenumber \(k\) of each run was determined from Eq. 6, which was used to find the wavelength \(\lambda\). For all frequencies, \(\lambda (0.51 – 0.82 \text{ m})\) was comparable to \(h\). Hence, the waves were considered to be deep-water waves. In most runs, the wave steepness \(ak < 0.2\). Only the two runs where the highest amplitude and the two highest frequencies were combined yielded \(0.2 < ak < 0.24\). Grue et al. showed that the velocity profile beneath a wave crest is very close to Stokes third order theory for \(ak = 0.23\), and still a good approximation for \(ak < 0.30 [38]\). Therefore, it is reasonable to use Eq. (7) as an approximation of theoretical \(u_w\).
Only bubbles located in vertical columns directly under the crest within 4% of the wavelength were considered. This was achieved by manually locating the position of the first incoming crest highlighted in Fig. 6 in the first relevant image frame. Frame 1. The crest position in the consecutive frames was found from $c_p/\Delta f$, where $c_p = \omega/k$ is the wave phase speed. Figure 7 shows the location of the first analyzed wave crest in Frame 1 (left panel) and Frame 6 (right panel) of the 1.4 Hz and 23 mm amplitude waves. The tracked particles within the blue shaded region were analyzed when the crest was located within the red dashed line, which marks the calibrated region of the image. When the first crest had passed out of the FOV, the next crest was found one wavelength upstream. This procedure was repeated for all 10 wave crests within a run. For each bubble detected directly below a crest, the vertical position and horizontal velocity was extracted. The inset figures in Fig. 7 shows the tracked bubbles within the white rectangle.

Figure 7: Frame 1 (left) and Frame 6 (right) of the 1.4 Hz and 23 mm amplitude waves propagating towards the left. The green star marks the crest location in the tank center and the blue shaded region highlights the 0.04λ wide column directly below the crest, which was considered. Only when the crest was located within the red dashed line (length: 45 cm, height: 30 cm), which marks the calibrated portion of the image, were the tracked particles within the crest column analyzed. In the inset figures, indicated with white rectangles, the tracked particles are marked with green circles.

In the following results, $u_b$ of the smallest tracked bubbles is presented in order to minimize the horizontal oscillating motion [29]. Since the FOV is too large to accurately determine the bubble size in the wave tank (as seen from Fig. 7), the bubble radius was estimated from the observed vertical velocity, based on the findings on the relation between $v_b$ and $R$ presented in Section 2.3. The vertical velocities accepted were $0.15 < v_b < 0.25 \text{ m/s}$, which should correspond to bubble radius $0.33 < R < 0.47 \text{ mm}$ according to the logarithmic curve fit applied in the left panel of Fig. 4. This range of bubble radius is marginally smaller than the bubbles investigated in Section 2.3, most likely due to either a slightly different input pressure on the perforated hose or the short duration (approximately 1 s) and the small FOV (approximately 4 cm) of the experiments presented in Section 2.3. For the 1.4 Hz waves, each run contained on average 683 analyzed bubbles which satisfied the quality criteria (small bubbles located directly below the crest) during the 10 periods investigated. In the case of 1.4 Hz waves, the 10 investigated periods consisted of approximately 214 image frames, meaning that 3.2 relevant bubbles were analyzed on average per image frame.

Figure 8 shows the horizontal velocity profiles below wave crests with increasing frequencies from upper to lower panels and increasing amplitudes from left to right panels. All three repetitions of the same frequency-amplitude combination are included in each panel. The blue dots indicate observed $u_b$. Each profile is divided into 10 equally spaced vertical bins, in order to assess statistical properties of the tracked particles along the profile. A global filter was applied to remove outliers which were mostly situated around $y = 0$, probably caused by local surface tension effects or from bubbles accumulating on the surface. Horizontal velocities which deviated more than a certain threshold off the mean of the same bin were discarded. The threshold was set to 70% of the total mean $u_b$ of all observed particles in the whole vertical profile. For the 1.4 Hz waves, this filter removed on average 2.9% of the observations. After filtering, the mean and two standard deviations ($2\sigma$) of $u_b$ within each bin were calculated (green error bars). Each bin must contain at least 10 tracked velocities for the error bar to be shown. The theoretical fluid velocity given in Eq. (7) is displayed with a red line. The expected bubble velocity was calculated from Eq. (11), with the range of estimated bubble radius and the kinematic viscosity $\nu = 10^{-6} \text{ m}^2/\text{s}$ inserted, and displayed as an orange highlighted region.

From Fig. 8, it can be seen that the horizontal bubble velocity in general has a smaller magnitude than the fluid velocity, but the exponential profile is clearly visible. The simple error estimate given in Section 2.3 and Eq. (11) seems to be a good approximation for the slip velocity, as the observed mean lies within the expected bubble velocity for most of the
Figure 8: Horizontal velocity profiles divided into 10 equally spaced vertical bins below wave crests. Wave frequency: 1.4, 1.6 and 1.8 Hz from upper to lower row, respectively. Wave amplitude: approximately 7, 14 and 21 mm from left to right column, respectively. Blue dots: observed horizontal bubble velocity $u_b$. Green error bars: mean and $2\sigma$ uncertainty of the observed $u_b$ within each bin (at least 10 observations must be present within the bin for the error bar to be displayed). Red line: theoretical horizontal water velocity $u_w$ given by Eq. (7). Orange shaded region: expected $u_b$ from estimated bubble size and Eq. (11). The observed mean $u_b$ lies within the expected region for most of the vertical bins.

The theoretical bubble velocity deviates 10.6$−21.6$, 12.2$−24.7$ and 13.7$−27.7\%$ from the theoretical fluid velocity for the 1.4, 1.6 and 1.8 Hz waves respectively, depending on the bubble radius. Statistics from the second vertical bin from the top of all the 1.4 Hz waves are displayed in Table 3. This bin was chosen because the highest bin in terms of $y$-position and velocity does not contain enough tracked bubbles in the case of the highest wave amplitude. Here, the mean observed $u_b$ deviates with $1.3−6.5\%$ from the expected $u_b$ of the average estimated bubble radius. The spread in the observed data, exemplified through $\sigma$ in Table 3, slightly increases with increasing fluid velocity (i.e. higher wave amplitude). However, the relative standard deviation of the observations, that is $\sigma$ divided by mean observed $u_b$ in the respective bin, decreases from 22.4$\%$ to 10.9$\%$ from the lowest to the highest amplitude for the 1.4 Hz waves. Deeming from Fig. 8, the relative standard deviation of the observed velocities decrease in general with increasing fluid velocity (i.e. higher wave amplitude and/or frequency). The mean observed $u_b$ is $11.7−16.0\%$ smaller than theoretical $u_w$.

The velocity profiles were nondimensionalized and the different amplitudes with the same frequency are plotted together with increasing frequency from left to right panel in Fig. 9. Outliers were filtered in the same manner as in Fig. 8. Here, the mean and error bar is presented if a bin contains at least 30 tracked bubbles. A clear collapse of the different amplitudes can be seen for each frequency. The mean of the observed $u_b$ is located within the expected region for most vertical bins. The wavenumbers from the different amplitudes are averaged to show theoretical $u_w$ and expected $u_b$ for each frequency. Visually, the spreading of tracked horizontal velocities seems to decrease with increasing frequency. This observation is consistent with the decreasing relative standard deviation for higher velocities, which was the case for the dimensional graphs in Fig. 8.
Table 3: Statistics of observed $u_b$ for 1.4 Hz waves in the second vertical bin from the top. The fourth column from the left is the percentage error between observed $u_b$ and theoretical $u_w$ (from Eq. 7) and the fifth column is the percentage error between observed and expected $u_b$ (from Eq. 11).

<table>
<thead>
<tr>
<th>$a$ [mm]</th>
<th>Mean obs. $u_b$ [m/s]</th>
<th>$\sigma$ obs. $u_b$ [m/s]</th>
<th>PE $u_w$ [%]</th>
<th>PE $u_b$ [%]</th>
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<tr>
<td>7</td>
<td>0.038</td>
<td>0.008</td>
<td>-13.9</td>
<td>3.8</td>
</tr>
<tr>
<td>15</td>
<td>0.079</td>
<td>0.010</td>
<td>-11.7</td>
<td>6.5</td>
</tr>
<tr>
<td>23</td>
<td>0.115</td>
<td>0.013</td>
<td>-16.0</td>
<td>1.3</td>
</tr>
</tbody>
</table>

Figure 9: Horizontal non-dimensional velocity profiles divided into 10 equally spaced vertical bins below wave crests. Wave frequency: 1.4, 1.6 and 1.8 Hz from left to right panel respectively. Blue dots: observed horizontal bubble velocity $u_b$. Green error bars: mean and $2\sigma$ uncertainty of the observed $u_b$ within each bin (at least 30 observations must be present within the bin for the error bar to be displayed). Red line: $e^{ky}$. Orange shaded region: expected $u_b$ from estimated bubble size and Eq. (11). The observed mean $u_b$ lies within the expected region for most of the vertical bins.

4 Field deployment

The ROV-PV system was tested close to an ice floe in the North-West Barents Sea on April 26, 2019. The objectives of the field experiment were to test the setup under Arctic conditions and to investigate the flow around floating ice subjected to wave motion. The site was near Hopen Island, which is a part of the Svalbard Archipelago shown in Fig. 10, and the geographical coordinates were 76.18°N, 25.77°E (indicated by the red dot). The setup was lowered down from a beam by a pulley system into the water next to the floe, which had a diameter of approximately 20 m. Further details on the ice floe can be found in [39]. Conductivity–temperature–depth (CTD) casts were not performed during the present study, but CTD profiles from a nearby location at the same time of the year in 2008 were reported in [40]. They found that the water density was quite constant in the upper 10 m of the water column, and that stratification occurred 20-50 m below the surface. These results indicate that any potential changes to bubble dynamics over the FOV due to stratification should be small close to the surface. Figure 11 shows the ROV positioned in front of the grid (left panel) and the bubble plane seen from the ROV (right panel). The bubbles were illuminated by both ambient light and the ROV headlights.

All the selected images were first calibrated with a cubic coordinate transform to find the relation between pixels and real world coordinates [22]. Due to the dense bubble plane and the relatively long distance from the ROV to the bubbles, it proved difficult to track single bubbles between frames obtained next to the ice floe. Therefore, PIV was applied to obtain a 2D velocity field within the calibrated region of the image. The PIV processing was performed with subwindows of $48 \times 48$ pixels and 75% overlap. The optimal window size was found from a convergence analysis where the sensitivity on the window size was investigated. A search range of 1/3 of the subwindow was used. This configuration yielded approximately $38 \times 30$ 2D velocity vectors inside the bubble plane. Each velocity vector can be decomposed into $u_b$ and $v_b$, i.e. a horizontal and a vertical component, respectively.

It is necessary to compensate for the vertical buoyancy driven velocity component of the bubbles in order to find $v_w$. It was found that the bubble rise velocity was approximately twice as high in the laboratory as in the field. For this reason, the laboratory results were not used to compensate for the buoyant motion. Instead, an on-site calibration under relatively calm conditions was performed. From a reference image couple where the horizontal velocity component was very small, PIV analysis were applied to find the velocity field. It is assumed that $v_w$ was small at the time of the reference image couple, since the observed $u_w$ was small. The mean vertical velocity from the reference image couple $v_{b, field}$ was found by spatially averaging the vertical velocity component $v_b$ over the entire velocity field. At
this stage in the analysis, the bubble rise velocity was approximated with $v_{b,field}$. Thereafter, an image couple with visible stronger flow was chosen and PIV analysis were performed to obtain the velocity field. Then, $v_{b,field}$ from the reference image couple was subtracted from all $v_b$ in the image couple that was analyzed, i.e. $v_w = v_b - v_{b,field}$. The alternative approach would be to subtract the most frequently observed terminal velocity in the laboratory experiment, i.e. $v_w = v_b - v_{b,lab}$, where $v_{b,lab}$ is the average vertical velocity from the right panel of Fig. 4. However, this method yields quite unrealistic results because of the much higher bubble rise velocity observed in the laboratory, and is therefore not pursued.

An example from an instantaneous velocity field obtained next to the ice floe is presented in Fig. 12. Measured bubble displacement in pixels between two frames are distributed into bins and displayed (blue) in the left ($u_b$) and right ($v_b$) panels. The probability distribution for the vertical water displacement obtained from the above mentioned approach, i.e. $v_w = v_{b,field}$ (orange) is shown in the right panel. The probability distribution for horizontal water displacement...
Figure 12: Example of an instantaneous velocity field obtained next to the ice floe. Probability distributions of the tracked bubble displacement (blue), and estimates for the water displacement in pixels between two frames from reference field experiments (orange) in the horizontal (left panel) and vertical (right panel) direction.

Figure 13: Vector plot of the same instantaneous flow field next to the ice floe as presented in Fig. 12. The vertical components are estimated from $v_b - \bar{v}_b,field$.

between two frames is shown in the left panel, where it is assumed that $u_b \approx u_w$. Similar water velocity magnitudes were observed in the horizontal and vertical direction.

A vector plot from the same image pair as shown in Fig. 12 is presented in Fig. 13. It is produced with the above mentioned approach, i.e. subtraction of the mean reference velocity in the vertical direction. The undisturbed ocean surface is set to $y = 0$ and the ice floe is located at a negative x-value. The direction of the water flow is towards the ice floe. In fact, it was observed from the ROV images that the horizontal velocity oscillated towards and away from the ice floe with periods in the order of 10 s. This motion is either direct wave motion or flow generated from the heaving ice floe.

5 Discussion

In order to determine the accuracy of the introduced ROV-PV system, potential sources of error must be identified and quantified. Fluid velocity is estimated from the motion of bubble tracers by identifying the bubble position with an
image processing technique. Two central questions that need to be addressed are how accurately the bubble position can be determined and how well a particle follows the fluid motion. According to [24], the position of particles spanning over one pixel is usually determined with pixel accuracy, while the uncertainty decreases for particles spanning over several pixels. In the wave tank experiments, the bubbles typically covered at least a couple of pixels. Therefore, one pixel is used as a conservative accuracy estimate of the bubble position, which corresponds to approximately 0.4 mm in real world coordinates. The bubble velocity is determined from a central differencing scheme, i.e. over three image frames. With a typical vertical velocity of 0.25 m/s, a bubble travels 0.025 m during three frames. This gives 1.6% relative error in velocity.

When it comes to the passivity of the particles or their capability of following the motion of the surrounding fluid, it has been shown that the horizontal bubble velocity lags behind the fluid velocity under periodic waves. This effect is attributed to the fluid acceleration and the bubble inertia due to the effect of added mass. For the 1.4 Hz waves, the discrepancies between observed bubble velocity and theoretical water velocity were 11.7–16%. The relative error in velocity caused by the bubble inertia increases with the bubble radius and is theoretically found to be 10.6–21.6% for the bubbles investigated under the 1.4 Hz waves. Hence, the error due to the bubble inertia is an order of magnitude larger than the error due to the image processing technique. The estimated horizontal slip velocity agrees with the observations. Note that the observed horizontal slip velocity yields Reynolds numbers of unity order of magnitude, which suggests that Stokes drag is a reasonable approximation.

A considerable spread in the observed $u_b$ under the wave crests can be seen from Figs. 8-9, where $\sigma \approx 0.01$ m/s. This spread is partly related to the varying bubble size and therefore varying slip velocity, as indicated by the orange shaded regions in the figures, but also due to secondary motion, typically oscillation of rising bubbles. The perforated hose was designed to produce as small bubbles as possible to obtain rectilinear trajectories. Although this was achieved to a certain extent, Fig. 4 shows maximum absolute horizontal velocity values of approximately 0.03 m/s for terminal velocity around 0.25 m/s ($Re \approx 300$) and zero minimum absolute horizontal velocity, i.e. a 0–12% relative error in velocity. This observation is in agreement with [34], who described onset of secondary motion from $Re = 200–1000$, where the typical horizontal component is 5% of the vertical component. Oscillations are most likely caused by vortex shedding. The magnitude of the horizontal velocity component of a rising bubble can be considered analogous to the oscillating part of surface waves, see e.g. Eq. (7). Its magnitude will depend on $y$-position and time. For bubbles of the same size, we can expect a uniform distribution in the horizontal velocity component due to oscillations. Since the bubble size varies, the distribution should be denser around the mean, which appears to be the case by visual inspections of Fig. 9.

The identified uncertainties and data spread in the wave tank experiment can be summed to give an estimate of the total relative error of the measured velocity. To narrow it down to a specific case, the 1.4 Hz waves are considered close to the surface. The most dominating source of error is introduced by the slip velocity, which is theoretically estimated to be 10.6–21.6%. The second largest error is due to the oscillating bubble motion. As the relative error due to oscillating bubble motion, which is mentioned in the previous paragraph, is based on the absolute horizontal velocity, this term can contribute in both directions and is estimated to be -12–12%, i.e. it contributes to data scattering. The smallest relative error arises from the tracking algorithm and is conservatively estimated to be 1.6%. When all contributions are summarized, the theoretically estimated relative error is 0.2–35.2%, which corresponds quite well with the observations in Fig. 9. This is of course a large data spread in the context of laboratory measurements, but acceptable in most field applications.

A possible issue with the methodology is whether the rise of bubbles could induce an undesired upward motion in the water itself, and thus be an intrusive technique. However, experimental studies of rising bubbles in vertical pipes, where $v_w$ and $v_b$ have been measured to find $v_{slip}$, suggests that this is not the case. Previous works show that $v_{slip}$ decrease with the local void fraction with approximately 20-37% when the local void fraction is increased from zero to 10% [42, 43]. This decrease is independent of $v_w$ and has been attributed to the hindrance effect due to neighboring bubbles in the surroundings acting as obstacles. Since the reported decrease in $v_{slip}$ is mainly caused by the decrease in $v_b$, the potential increase in $v_w$ should be small. The local void fraction was estimated to be less than 2% in the terminal velocity experiment presented in Section 2.3 by calculating the ratio of the bubble area on the total area of the FOV.

There is a large uncertainty associated with measurements of the vertical water velocity component since the bubbles are naturally buoyant. Bubbles rising in calm water and vertical columns directly below wave crests have been investigated in laboratory experiments. In both cases, the vertical water velocity is known to be zero, so that the vertical bubble velocity equals the vertical slip velocity. However, such controlled environment is seldom obtained in the field. In the present study, on-site observations in the field were used to approximate and remove the buoyancy driven vertical velocity, because the bubble rise velocity measured in the laboratory was much higher. A possible explanation for this deviation is the temperature difference, which may have changed the flexibility of the perforated hose. For example, the diameter of the holes where the bubbles escape may decrease in lower temperature, or they could be partly clogged by
ice. The reduction in rise velocity could also be due to a different ratio in terminal velocity versus radius because of the lower temperature. As an example, Haberman and Morton reported a 1/3 reduction in terminal velocity for the same bubble size in distilled water when the temperature was reduced from room temperature to 6°C [29].

The current ROV-PV system with bubbles as tracing particles is suitable for measuring flows where the horizontal velocity component is of interest. Examples of such situations are waves propagating under an ice cover and horizontal shear flow in the boundary layer beneath an ice cover. For accurate measurements of the vertical velocity component, it is necessary to generate uniform bubbles in size. In this way, the terminal velocity due to the buoyancy can be found on-site or in advance if the laboratory conditions (temperature, salinity and pressure) are similar, and assumed within reasonable accuracy to be constant for all bubbles. The vertical fluid velocity can then be found from Eq. (2) in the case of steady flow, or from a combination of Eq. (2)-(3) in the case of accelerated flow. Further effort must be spent in order to improve the perforated hose to overcome this challenge in future studies. At the same time, focus should be directed towards keeping the construction simple in use and robust for field conditions. Hydrogen bubbles generated from hydrolysis could be an option as these tend to be small (diameter around 0.1 mm), although [44] reported a distribution in bubble size due to coalescence phenomena. On the other hand, bubbles cannot be identified by the particle tracking software if they are too small.

6 Conclusions

In this paper, a new method for measuring 2D velocity fields in the upper ocean by utilizing image processing technology and air bubbles as tracing particles has been presented. The novelty of this method is the combination of bubbles and an ROV, which gives a simple and lightweight setup which is suitable for field measurements in the polar regions. The ROV-PV system has been demonstrated to measure the flow in the vicinity of an ice floe during an Arctic field campaign. There is a need for detailed ocean kinematics in the marginal ice zone, as a lot of important atmosphere-ice-ocean energy transfer processes occur here and relatively few in situ observations of 2D velocity fields exist. The introduced technique could be suitable for visualization and quantification of many interesting flow phenomena in the Arctic, for example wave propagation underneath various ice layers or turbulent structures induced by colliding ice floes. Detailed observations within this field are important for an improved understanding of the underlying physics and the design and validation of numerical sea ice models.

Detailed laboratory experiments have been carried out to quantify the relation between diameter and terminal velocity of the generated bubbles. Although many approximations and models exist for this relation, the present results demonstrate the importance of a thorough investigation of bubble properties since substantial deviations to models may occur. The ROV-PV system has been utilized to measure horizontal velocities under periodic water waves with an accuracy in the order of 10%. The deviations are mainly attributed to the slip velocity caused by the bubble inertia due to the added mass, and is important to consider when measuring accelerated flow. In addition, there is a spread in the data expressed through a relative standard deviation in the order of 10%, due to the bubble size distribution and the oscillatory motion of rising bubbles. More precisely, the relative error with respect to the horizontal water velocity is estimated to be 0.2–35.2%, which agrees with the observations. Future studies should focus on generating smaller and more uniform bubbles, which could decrease the spread in observed velocities and improve the reliability in measurements of the vertical velocity component.

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References


Paper IV

Iceberg stability during towing in a wave field

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Iceberg stability during towing in a wave field

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ABSTRACT

Due to their large mass and small aspect ratio, icebergs pose a threat to boats and offshore structures. Small icebergs and bergy bits can cause harm to platform hulls and are more difficult to discover remotely. As icebergs are dynamic mediums, the study of icebergs in relation to safe human operations requires the rigorous analysis of the ice-ocean interaction, in particular with waves and currents. In this paper, we present iceberg towing experiments and analyze iceberg stability from GPS tracks and inertial motion unit data. The towline tension as well as the boat motion relative to the iceberg was measured. Different scenarios were investigated by changing the towing strategy with regards to towing speed, direction (straight or curved trajectory) and acceleration. Large amplitude roll oscillations with period of approximately 30 s were observed immediately after the load dropped and the iceberg returned to a stable static position. In two of the cases, the iceberg flipped over partly or entirely after some towing time. From the load cell, we observed oscillations in the system with periods of approximately 6 s, which were attributed to the rope elastic properties and the iceberg response. The load oscillations increased when the towing direction was against the waves as opposed to perpendicular to the waves.

KEY WORDS: Ice-ocean interaction; Iceberg towing; Iceberg stability; Waves.

INTRODUCTION

The decline in the Arctic ice cover that has been observed over the past decades has allowed for more human activities in the region, such as shipping and exploitation of natural resources (Smith and Stephenson, 2013; Feltham, 2015). Even though the ice cover is decreasing, icebergs are observed in almost all Arctic seas (Abramov and Tunik, 1996). Drifting icebergs poses a threat to floating or fixed structures and their existence influences the concept of offshore development. Due to their large mass, icebergs can apply considerable pressure stress on platform hulls and cause mechanical failure. The technical term for small icebergs like the ones investigated in this study is bergy bit (< 1000 tons), but they will be referred to as icebergs in the text. Small icebergs can still influence damage on constructions, and they are more difficult to observe remotely, as pointed out by Marchenko et al. (2020).

When there is a risk of collision between icebergs and platforms, it is necessary to deflect its drifting course to ensure safe human operations in polar offshore regions. Iceberg towing is a well known technique and many experimental tests and studies have been performed, also in the resent
years, e.g. Kornishin et al. (2019) and Efimov et al. (2019). Marchenko and Gudoshnikov (2005) identified several requirements for the towing operation to be successful, e.g. that the ship should have sufficient power to significantly change the iceberg trajectory, and the towline needs to resist the water and wave induced drag forces on the iceberg. Due to their small aspect ratio, especially spherical icebergs are prone to capsizing when external forces are applied. Such an event can cause harm to people and equipment during the operation. Knowledge of iceberg stability during towing is therefore of importance to reduce the risk of accidents (Marchenko, 2006).

Our aim with this study is to investigate iceberg stability during towing to improve the safety of such an operation. Unsuccessful towing events are often caused by towline slippage or iceberg overturning (Crocker et al., 1998). Therefore, we have applied sinking line towing, where the towline is partly submerged around the iceberg with the depth regulated with added buoyancy. This method reduces the probability of towline slippage. In addition, the towline force is applied closer to the iceberg center of rotation, which reduces the overturning moment and the risk of iceberg rollover (Crocker et al., 1998). Up until now, most studies on iceberg drift and towing have installed only GPS trackers on the icebergs, which give information about drift, but not on the iceberg dynamics. In this study, towing experiments with small icebergs are presented. The novelty of the investigation is the installation of Inertial Motion Units (IMUs) on the icebergs, which have allowed for observations of three-axes acceleration and rotation. Detailed surveillance of such motion, in addition to knowledge about the iceberg sub-surface geometry, which was obtained with a Remotely Operated Vehicle (ROV), gives insight in the hydrodynamic stability of the ice structures. Met-ocean parameters and the applied towing force were measured. The massive instrumentation allowed for the investigation of the effect of incoming waves and ocean current.

DATA AND METHODS

Towing experiments were carried out in the Tempelfjord near Longyearbyen on Svalbard in September 2020. Tunabreen glacier extends out in the fjord and frequent calving events makes it an ideal site for working with small icebergs and bergy bits. Figure 1 shows the location and a Sentinel-2 satellite image of the fjord and the glacier from September 22, 12:27 UTC.

![Figure 1. Location of the experiments. a) Map of the Svalbard Archipelago with the Tempelfjord indicated (TopoSvalbard, 2021). b) Satellite image of the Tempelfjord and the Tunabreen glacier.](image)

Experimental setup

A 10.5 m long, 500 Hp Polarcircle 1050 boat with 7 tons total mass was used in the experiments. It was decided to use a fast boat in order to save time on the daily 50 km trip from Longyearbyen to
Tempelfjord. Open source IMUs and GPS trackers (Rabault et al. 2017) were mounted on the boat roof and on the iceberg with ice screws to measure their absolute and relative motion. In addition, a pair of Garmin Astro GPS trackers were used as redundancy. The keel depth $h_k$ was determined and the sub-surface structure was investigated with an ROV. It was attempted to perform a video scan with the ROV to generate a 3D model of the sub-surface geometry, but the visibility was too poor due to the high concentration of glacier sediments. Wave motion was measured with an IMU and ocean current with an Acoustic Doppler Current Profiler (ADCP), which were mounted on buoys and moored to an anchor in the vicinity of the towing experiments.

After the installation of instruments on the iceberg and the buoys, the towing setup was arranged. A 12 mm thick polyester sinking rope was applied. The 92 m long rope was deployed around the iceberg from a small rubber boat while the large boat stayed at a fixed distance to the iceberg. Floaters on 1.5 m long straps were attached to the rope with 2-3 m spacing, in order to keep the towline submerged under the surface and prevent slippage during the operation. The rope ends were connected to a 5 tons load cell from Strainstall (type 12160-3). A 9 m long piece of double rope was attached to each side of the boat stern and connected to the load cell in the center. The joining point was kept floating with a buoy. The setup is illustrated in Fig. 2.

Three successful towing events were carried out and will be referred to as T1-T3. Experimental details are summarized in Table 1. At the end of T3, the iceberg rolled over and the IMU attached to it was lost, hence the missing parameters in Table 1. The iceberg area $S$ at the water line was estimated from a series of photos of the iceberg and an object of known dimensions from different angles. In T2, the photos were used to produce a 3D model of the iceberg above-surface geometry (not presented), in order to determine the area with better accuracy. The mean free board $h_{fb}$ was estimated from visual observations. The period of natural oscillation in heave $T_h$ was determined from the heave spectra from the IMU placed on the iceberg with the Welch method, described under “Data processing”.

Table 1. Experimental details with emphasis on iceberg properties.

<table>
<thead>
<tr>
<th>Exp.</th>
<th>Date</th>
<th>Time (UTC)</th>
<th>$S$ [m$^2$]</th>
<th>$h_k$ [m]</th>
<th>$h_{fb}$ [m]</th>
<th>$T_h$ [s]</th>
<th>$m_i$ [tons]</th>
</tr>
</thead>
<tbody>
<tr>
<td>T1</td>
<td>22</td>
<td>14:20 ~ 15:30</td>
<td>31</td>
<td>8</td>
<td>1.6</td>
<td>7.32</td>
<td>424</td>
</tr>
<tr>
<td>T2</td>
<td>24</td>
<td>11:40 ~ 12:20</td>
<td>28</td>
<td>5</td>
<td>0.8</td>
<td>4.69</td>
<td>158</td>
</tr>
<tr>
<td>T3</td>
<td>25</td>
<td>11:40 ~ 12:00</td>
<td>29</td>
<td>4.5</td>
<td>1.0</td>
<td></td>
<td>145</td>
</tr>
</tbody>
</table>
Iceberg mass $m_I$ was estimated from $S$ and $T_h$ from the momentum equation in the vertical direction. Following Marchenko et al. (2020), inertial force is balanced by buoyancy and gravity forces

$$m_I \frac{d^2 z}{dt^2} = \rho_w g (V_w - S z) - m_I g,$$

where $z = z(t)$ is the vertical displacement of the iceberg (positive upwards) relative to the calm water surface at $z = 0$, $\rho_w$ is the water density, $V_w$ is the submerged iceberg volume in hydrostatic equilibrium and $g$ is the acceleration of gravity. In hydrostatic equilibrium, $\rho_w V_w = m_I$ and from the solution of Eq. 1, the iceberg mass is

$$m_I = \frac{\rho_w g S T_h^2}{4 \pi^2}.$$

In T3 when $T_h$ data were lost, $m_I$ was estimated from the approximated iceberg volume $S(h_{fb} + h_k)$ and a typical iceberg density of 910 kg/m$^3$ (Robe, 1980).

Different towing strategies were applied in the experiments, the boat heading was either straight or curved and the motor power was either constant or slowly increased. The different trajectories from the IMU GPS are presented in Fig. 3, where the boat position (blue) and the iceberg position during free drift before towing (gray) and towing (orange) are indicated. The starting time of the towing is indicated and 10 min intervals are marked with dots. Even though the instruments were lost when the iceberg tipped in T3, the Garmin Astro GPS tracker transferred coordinates via radio communication. The iceberg trajectory presented in Fig. 3c is therefore from the Garmin instrument. A comparison of the GPS tracks in T1-T2 (not presented) show good agreement between the two different instruments.

Due to the elastic properties of the towline, oscillations can be expected in the iceberg-rope-boat system. Marchenko and Eik (2012) investigated the stability of steady towing, i.e. at constant towing speed, propulsion and water speed, by evaluating small fluctuations in the vicinity of the steady solution of the system’s momentum equations in the axial direction. They estimated the period of the system oscillations $T_s$ and their Eq. (17) can be written as

$$T_s = 2 \pi \sqrt{\frac{m_B}{\int F_t \, dx}},$$

Figure 3. Boat and iceberg trajectory during towing and free drift. The starting time of the towing (in UTC) is indicated and 10 min intervals are marked with dots. a) T1. b) T2. c) T3.
where $m_B$ is the boat mass, $F_t$ is the rope tension, $X$ is the distance from the boat to the iceberg and the derivative is evaluated during steady towing. Equation 3 neglects the added mass of the boat $m_{B,a}$ in the axial direction ($m_{B,a}/m_B \ll 1$) and is valid when the boat mass is much smaller than the iceberg mass ($m_B/m_I \ll 1$), which was the case in all the present experiments.

**Data processing**

The sampling frequency was approximately 10 Hz for the IMUs and 1 Hz for the GPSs. Time series of speed, direction and distance between sensors were obtained from GPS positions. Signal smoothing of the GPS time series was performed with a third order polynomial Savitzky-Golay filter with a 41-point window size. Iceberg heave motion and ocean surface elevation were obtained from downward acceleration time series from the IMU placed on the iceberg and on the buoy, respectively. The time series were first re-sampled to obtain a constant sampling frequency of 10 Hz. Numerical integration of downward acceleration with respect to time was performed twice to obtain vertical displacement. After each integration step, a second order Butterworth bandpass filter with cutoff frequencies of 0.05 and 2 Hz was applied to remove any low frequency noise associated with the integration (Sutherland and Rabault, 2016). Displacement in the horizontal directions was obtained with the same approach.

Power Spectral Densities (PSDs) of displacement obtained from the double integrated accelerations, and of rotation obtained directly from the IMUs, were calculated from 30 min time series. The time series were subdivided into segments of 2048 data points with 50% overlap and the segments were Fourier transformed. The spectral estimates were ensemble averaged to decrease statistical uncertainties according to the Welch method, and a Hanning window was applied to each segment to reduce spectral leakage (Earle, 1996). The resulting PSD had approximately 25 degrees of freedom. Spectral 95% confidence intervals were estimated from the Chi-squared distribution (Earle, 1996).

The load cell output, which had a sampling frequency of approximately 10 Hz, was converted from voltage to kN from a calibration curve which was obtained by measuring the static load of three lifted objects with a known mass ranging 0.2-1.5 tons. Linear regression was applied to fit a straight line between the data points. A calibration was performed before and after the experiments, and the average slope was used.

**Oceanographic conditions**

Wave motion was measured in T2-3, but the IMU buoy was not deployed during T1. Significant wave height $H_S$ was found from $H_S = 4\sigma$, where $\sigma$ is the standard deviation of the 30 min ocean surface elevation time series. Non-directional wave spectra were estimated from the Welch method described in “Data processing”. Water velocity $U_W$ and direction $UD$ was measured with an RDI Sentinel 1200 (kHz) ADCP (0.3 m bin size) in T1 and with a Nortek Signature 1000 (kHz) ADCP in T2-3 (0.2 m bin size). Pings were ensemble averaged in 1 min intervals (The listed values are from the bin corresponding to 3 m below the surface). Wind speed $WS$ and direction $WD$ (relative to the boat) and air temperature $T_A$ was measured in 10 s intervals with a Young 81000 ultrasonic anemometer mounted 3 m above the ocean surface in T1 and T3. The listed $WD$ are absolute values obtained from the boat heading found from the GPS data. Conductivity–temperature–depth casts were performed with a SBE 19 V2 profiler to measure water density $\rho_W$, temperature $T_W$ (the listed values were measured 3 m below the surface) and salinity. The density profile presented in Fig. 4b show a strong stratification with a pycnocline around $h_0 = 10$ m below the surface. We use the going-to convention when describing direction, and the angle is defined as clockwise rotation from north. The observed oceanographic and meteorologic parameters are summarized in Table 2. It is assumed that the sea state and the wind conditions were constant over the duration of each towing
experiment (~1 h, listed in Table 1), hence are the ADCP and anemometer data time averaged over this period.

<table>
<thead>
<tr>
<th>Exp.</th>
<th>$H_S$ [m]</th>
<th>$U_W$ [m/s]</th>
<th>UD [$^\circ$]</th>
<th>$WS$ [m/s]</th>
<th>$WD$ [$^\circ$]</th>
<th>$T_A$ [$^\circ$C]</th>
<th>$\rho_W$ [kg/m$^3$]</th>
<th>$T_W$ [$^\circ$C]</th>
</tr>
</thead>
<tbody>
<tr>
<td>T1</td>
<td>0.12</td>
<td>168</td>
<td>4.0</td>
<td>148</td>
<td>2</td>
<td>1026</td>
<td>4.0</td>
<td></td>
</tr>
<tr>
<td>T2</td>
<td>0.02</td>
<td>0.02</td>
<td>302</td>
<td></td>
<td>1025</td>
<td>3.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>T3</td>
<td>0.12</td>
<td>0.08</td>
<td>188</td>
<td>3.8</td>
<td>68</td>
<td>1025</td>
<td>3.7</td>
<td></td>
</tr>
</tbody>
</table>

Wave measurements were not performed in T1, but the wave conditions were visually observed to be calm on this day, probably due to the fact that the wind was coming from south-west where the fetch was small. No considerable waves were measured in T2 and the wind conditions were visually observed to be calm on this day. Some wave activity was observed and measured in T3. The wind was coming from north-east on this day, which allowed the waves to build up in the longitudinal direction of the fjord. Deeming from the wave spectrum presented in Fig. 4a, low frequency swell coming in the larger Isfjord were dominating the wind waves.

Figure 4. Oceanographic conditions. a) Wave spectrum with 95% confidence intervals during T3. Low frequency swell can be observed around 0.05 Hz and high frequency wind waves are visible around 0.8 Hz. b) Density profiles from September 23 in Tempelfjord, where the upper layer depth $h_0 = 11$ m.

**RESULTS**

Iceberg roll is defined as rotation about the IMU x-axis (randomly oriented in the horizontal plane) and yaw as rotation about the z-axis (pointing downwards). Figure 5 shows time series of iceberg roll, yaw, towing load, boat and iceberg speed and distance between boat and iceberg from T1. Three short acceleration tests were performed in the time span 14:20-14:27. A fourth acceleration test was carried out 14:30-14:34, where the towing load was increased up to 8.3 kN. The maximum towing speed was approximately 0.6 m/s. There were oscillations in the towing load with 6.1 s period, which can be seen in the inset plot of the rope tension during maximum load. The system oscillations were also present during slack conditions (when the rope was tight, but the load was small), although the amplitude was smaller. The oscillation period is further addressed in the towing load spectra in Fig. 8. Spectra from iceberg translation in the horizontal (surge/sway) and vertical (heave) direction (not presented) show oscillations with 12.8 and 7.3 s periods, respectively.
The acceleration tests shown in Fig. 5 induced an oscillating iceberg roll motion with periods of 34-51 s (from the roll spectra, not presented). Immediately after the third and fourth acceleration test, the amplitude of the roll oscillations grew quite large and was slowly damped. The iceberg yawed approximately 70° immediately after the fourth acceleration test. A slowly increasing roll motion was initiated by the third acceleration test. The roll angle increased from 3° initially to 15° at 15:20, when a sudden event, possibly a part falling off the iceberg or a shift in towline position, made the iceberg roll back to 1° with a slow damping, now with oscillation period of 26 s. A fifth acceleration test was carried out around 15:30, but the iceberg started to roll shortly afterwards, and the test had to be terminated 15:32 when the iceberg rolled around 90°. The IMU ended up floating on the surface and was barely accessible for retrieval.

Figure 5. T1 time series of iceberg roll (pink), iceberg yaw (red), towing load (green), boat (blue) and iceberg (orange) speed and distance between boat and iceberg (gray). The speeds are multiplied by 10 and shifted 5 units down and the distance and yaw angle are divided by 10 to increase the readability. The inset plot shows a close-up of the rope tension during maximum load. There was an interruption in the experiment ~14:40-15:30 with slack conditions, during which the load cell was switched off.

Figure 6. T2 time series. See figure text of Fig. 5 for further details.
Figure 6 shows the time series from T2. The first acceleration test was performed 11:42, during which the period of roll oscillation was 12.7 s (from the roll spectra, not presented). Thereafter, the iceberg was towed with constant motor power in a curved line in the time span 11:55-12:13, where the period of oscillation in iceberg roll decreased to 5.7 s. The maximum rate of change in iceberg course was approximately 32°/min. A second acceleration test was performed at 12:15, where the maximum towing load reached 9.0 kN, the towing load oscillation period was 6.3 s (shown in the inset plot) and the maximum boat and iceberg speed reached 0.75 m/s. The distance between the boat and the iceberg decreased from 53.3 m during maximum load to 51.3 m during slack conditions immediately after, i.e. the rope extended approximately 2 m or 4%. After the towing stopped, the iceberg rolled 6° and yawed 45° immediately after the load dropped.

Figure 7 shows the time series from T3 without the lost IMU data. Waves were present this day. The towing direction was towards the wind waves and perpendicular to the wind waves in inset plots a) and b), respectively, assuming that the wind waves were traveling in the same direction as the wind. The oscillations in towing load were more prominent in the first case, suggesting that the wave load affected the towing. Periods of oscillation were 4.5 s in both cases.

Figure 7. T3 time series. See figure text of Fig. 5 for further details. Inset plots are from instances when the boat was steaming a) against the wind waves and b) perpendicular to the wind waves.

Figure 8. Towing load spectra from 2 min time series corresponding to the inset plots in Figs. 5-7 with 95% confidence intervals (~6 degrees of freedom). a) T1. b) T2. c) T3, where towing direction against (gray) and perpendicular to (green) the wind waves are shown.

Power spectral densities were calculated from 2 min time series of the towing load with the Welch method described under “Data processing”, except that the segment length was 512 data points.
Figure 8a-c show the towing load spectra obtained from the time series presented in the inset plots of Figs. 5-7, and the local peak frequency corresponds to periods of 6.1, 6.3 and 4.5 s, respectively. The spectra in Fig. 8c confirms that the towing load oscillations were greater when the towing direction was against (green) than when it was perpendicular to (gray) the wind waves. The increase in system oscillation amplitude may be caused by the wave drag, which influences the iceberg equation of motion in the axial direction.

The resistance of the water to the motion of the iceberg is equal to and opposite directed as the force applied by the rope during steady towing. That is when the wave and wind drag forces are neglected. The water-iceberg form drag coefficient $C_{W,I}$ can be estimated as

$$C_{W,I} = \frac{F_t}{\rho W S_x (U_I - U_{W,x})^2},$$

where $S_x$ is the vertical cross-sectional area of the submerged part of the iceberg that is perpendicular to the axial direction, $U_I$ is the iceberg speed and $U_{W,x}$ is the water speed in the axial direction (Marchenko and Eik, 2012). Equation 4 is used to estimate $C_{W,I}$ in T1-2 when the wind and wave conditions were relatively calm, and their associated resistance forces can be neglected. $S_x$ is approximated as $hl$, where $l = (4S/\pi)^{0.5}$ is a representative iceberg length scale in the horizontal direction. The drag coefficient is evaluated at 14:33 in T1 and at 12:00 in T2, where the iceberg trajectory and speed were relatively constant, and the values 0.70 and 0.54 are obtained, respectively. These values agree with Robe (1980), who reported $0.5 < C_{W,I} < 1$.

DISCUSSION

The observed oscillations in the towing load may be due to the towline elastic properties described in Eq. 3. The rate of change in $F_t$ with respect to $X$ can be estimated from the rope properties given by the manufacturer: 15% extension at maximum tension of 2.3 tons. Since the rope was deployed around the iceberg, the maximum tension should be doubled. With a rope length of 92/2 m, $dF_t/dX = 6540$ N/m, which corresponds to $T_s = 6.5$ s. Another approach is to estimate $dF_t/dX$ from the measured towing load and rope extension, i.e. 9.0 kN/2 m = 4500 N/m, which corresponds to $T_s = 7.8$ s. At least the first estimate corresponds well with the observed oscillating period of 4.5-6.3 s. Both estimates assume a linear relation between rope tension and extension, which is a simplification of reality. It should be emphasized that the measured rope extension, i.e. the change in distance between the boat and the iceberg, was calculated from the GPS positions, which has an accuracy in the order of 1 m. This uncertainty propagates to the velocities presented in Figs. 5-7.

Towing load oscillations could also be caused by iceberg motion. The observed periods of roll oscillation were > 30 s during T1, i.e. much greater than the towing load oscillations with periods of 6.1 s. However, during T2, periods of 5.7 s were observed in roll. Natural oscillations in heave with period 7.3 and 4.7 s during T1 and T2, respectively, are also likely causes for the towing load oscillations. In the case where the towline was tightly fixed around the iceberg, small vertical movements of the iceberg around an equilibrium state may have changed the length of the rope and altered the towing load.

Large oscillations in iceberg roll were observed immediately after the towing load dropped. It is likely that the force applied by the rope tilted the iceberg during towing, and when the towing load dropped, the iceberg returned to a stable static position. Only very symmetric floating bodies have a continuum of stable positions, e.g. spheres. Normal bodies with several faces, such as icebergs, have a finite number of stable static positions. Nonlinear oscillations induced by the towing may have influenced transitions between stable positions. This could explain the sudden roll observed 15:20 in T1 and 12:18 in T2. The temperature was above the freezing point which influenced systematic disconnection of ice features from the iceberg during the towing and consequently led to iceberg instability in the water. The non-spherical iceberg shape may prevent rotation when angular momentum is applied by the towing line, which could explain the observed oscillations in
the horizontal translation. Indeed, the yaw angle was relatively constant during towing, large changes occurred immediately after the load dropped.

The stratification shown in Fig. 4b is explained by ice melting. Internal waves may be generated at the pycnocline, which could impose a large additional drag force on the iceberg. This phenomenon is known as “dead water” in ship terminology and can reduce the ship speed substantially compared to normal conditions with equal propulsion. A strong internal wave and dead water resistance force can be produced when the ratio between the ship draught and the upper layer depth $h_0$ is close to 1 (Grue, 2018). At least in T1, $h/h_0$ was close to unity and the dead water may have substantially increased the resistance on the iceberg.

CONCLUSIONS

Three iceberg towing experiments were carried out on Svalbard in September 2020. The presented towing technique, which consisted of partly submerging the towing line around the iceberg, proved successful in the sense that towline slippage was avoided, and the iceberg trajectory was altered. However, the iceberg rolled approximately 90° in one situation and 180° in another. This illustrates that iceberg towing is not trivial and the importance of a well organized methodology with focus on safety. It is advantageous to have an impression of the iceberg sub-surface geometry in advance of the towing in order to be able to adjust the depth of the towline. This was obtained with an ROV in these experiments. The above-surface geometry was also documented, which enabled estimation of the iceberg mass (145-424 tons).

Boat and iceberg motion was measured with a three-axis accelerometer and gyroscope, and GPS trackers. Various met-ocean parameters were monitored, and the towline tension was measured with a load sensor. This massive instrumentation allowed for a detailed study on the dynamics of the iceberg and of the boat-rope-iceberg system. Slowly damped oscillations with period around 30 s were observed in iceberg roll, particularly right after the towing load dropped, which probably means that the iceberg returned to, or transitioned between stable static positions. Oscillations with period around 6 s were observed in the towing load. This phenomenon is attributed to either the rope elastic properties and/or the iceberg heave motion. Although the maximum significant wave height was only 12 cm, the amplitude of the towing load oscillation was significantly greater when the towing was performed against, compared with perpendicular to the wind waves, probably due to the extra wave drag on the iceberg in the axial direction. In addition to waves, upper ocean stratification could increase the iceberg drag according to the dead water phenomenon, and may be important to consider during towing operations, especially when the iceberg draft is comparable to the pycnocline depth, which was the case in one of the reported experiments.

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REFERENCES


