

Potential and Limitations of a Commercial Broadband Echo Sounder for Remote Observations of Turbulent Mixing

JULIA MUCHOWSKI^{1a,c}, LARS UMLAUF^b, LARS ARNEBORG^c, PETER HOLTERMANN^b, ELIZABETH WEIDNER^{a,d}, CHRISTOPH HUMBORG^{a,e} AND CHRISTIAN STRANNE^{a,e}

^a Department of Geological Sciences, Stockholm University, Stockholm, Sweden

^b Institute for Baltic Sea Research, Warnemünde, Germany

^c Department of Research and Development, Swedish Meteorological and Hydrological Institute, Gothenburg, Sweden

^d Department of Earth Science, University of New Hampshire, Durham, New Hampshire

^e Baltic Sea Center, Stockholm, Sweden

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ABSTRACT: Stratified oceanic turbulence is strongly intermittent in time and space, and therefore generally under-resolved by currently available in situ observational approaches. A promising tool to at least partly overcome this constraint are broadband acoustic observations of turbulent microstructure that have the potential to provide mixing parameters at orders of magnitude higher resolution compared to conventional approaches. Here, we discuss the applicability, limitations, and measurement uncertainties of this approach for some prototypical turbulent flows (stratified shear layers, turbulent flow across a sill), based on a comparison of broadband acoustic observations and data from a free-falling turbulence microstructure profiler. We find that broadband acoustics are able to provide a quantitative description of turbulence energy dissipation in stratified shear layers (correlation coefficient $r = 0.84$) if the stratification parameters required by the method are carefully preprocessed. Essential components of our suggested preprocessing algorithm are 1) a vertical low-pass filtering of temperature and salinity profiles at a scale slightly larger than the Ozmidov length scale of turbulence and 2) an automated elimination of weakly stratified layers according to a gradient threshold criterion. We also show that in weakly stratified conditions, the acoustic approach may yield acceptable results if representative averaged vertical temperature and salinity gradients rather than local gradients are used. Our findings provide a step toward routine turbulence measurements in the upper ocean from moving vessels by combining broadband acoustics with in situ CTD profiles.

KEYWORDS: Turbulence; Mixing; Acoustic measurements/effects; In situ oceanic observations; Remote sensing; Diapycnal mixing

1. Introduction

Vertical turbulent mixing is the dominant underlying process determining the vertical fluxes of heat, salt, and other dissolved substances in the ocean (Wunsch and Ferrari 2004), fjords and estuaries (Inall and Gillibrand 2010), and lakes (Wüest and Lorke 2003). By influencing the vertical distribution of heat and dissolved substances, vertical turbulent mixing plays a key role for marine ecosystems (Sanford 1997). On a global scale, the distribution of heat and salt governs the thermohaline circulation of the world oceans (Rahmstorf 2006), while at the basin-scale vertical turbulent mixing is a key player in many processes. Examples include sharp frontal regions in the surface layer (Peng et al. 2020; D'Asaro et al. 2011), localized shear instabilities inside density interfaces

(Smyth and Moum 2012; van Haren and Gostiaux 2010), and turbulent bottom boundary layers (Davis and Monismith 2011; Becherer and Umlauf 2011; Lappe and Umlauf 2016). Regions with steep seafloor bathymetry have been identified as mixing “hotspots” (Polzin et al. 1997; Wunsch and Ferrari 2004; Nash and Moum 2002; Arneborg and Liljebladh 2009; Arneborg et al. 2004), as hydraulically controlled flow over steep seafloor bathymetry can lead to strong turbulence and thereby cause highly elevated vertical mixing (Farmer and Armi 1999; Smyth and Moum 2001; Klymak and Gregg 2004; Lamb 2004; Abe and Nakamura 2013; Arneborg et al. 2017). In these marine and limnic systems vertical turbulent mixing is often difficult to quantify from local measurements due to the fact that turbulence is often localized and strongly intermittent.

The amount of vertical mixing in the water column can be estimated from observations of dissipation rates of temperature variance, salinity variance (Osborn and Cox 1972), or turbulent kinetic energy (TKE) in combination with density stratification data (Osborn 1980). Established direct methods, such as turbulent microstructure observations with airfoil shear probes and fast response thermistors mounted on vertical profilers, moorings, and ocean gliders, capture only one-dimensional profiles through the water column at a certain time or a time series at a certain location. Similarly, turbulent dissipation rates and fluxes obtained from state-of-the-art acoustic Doppler velocimeters

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Corresponding author: Julia Muchowski, julia.muchowski@geo.su.se

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(ADVs), based on inertial subrange scaling and eddy covariance approaches, provide excellent temporal resolution, but such measurements cover only a few points in space even with the most advanced systems of this type (Davis and Monismith 2011). High-resolution time series of the dissipation rate provided by acoustic turbulence measurements from ADCPs, based on structure-function and inertial-subrange approaches, have partly overcome these problems (Wiles et al. 2006; Lorke and Wüest 2005). Yet these methods can presently only be applied for high-frequency ADCPs with very limited range, and to our knowledge have not been used for vessel-mounted ADCPs (Lucas et al. 2014). Moreover, they only provide reliable data in highly energetic turbulence regimes, often failing to capture the stratified interior region of marine systems; therefore, it is likely that the majority of bathymetry-related hotspots for turbulent mixing still remains undiscovered by currently available measurements.

In models, the representation of stratified flow over small-scale steep bathymetry is challenging as it requires a nonhydrostatic description and very high resolution, and still the results are sensitive to the subgrid-scale parameterization of mixing (Berntsen et al. 2009). Regional models are generally of much too coarse resolution to directly resolve the flow and the mixing it causes, and therefore, turbulence needs to be parameterized. However, a generally applicable parameterization does not exist today.

Active acoustic systems have been increasingly used to remotely estimate levels of turbulent mixing at higher spatial and temporal resolution than traditional methods. Stratified turbulence scatters sound from density and compressibility perturbations in turbulent eddies at the Bragg wave vector (Goodman and Forbes 1990). A well-known acoustic scattering model of random, small perturbations in medium density and compressibility (e.g., Kraichnan 1953; Tatarski 1961) has been used by several groups to estimate dissipation rates (e.g., Seim et al. 1995). Lavery et al. (2003) expanded the model to oceanic applications by incorporating scattering as a function of temperature and salinity profiles. Through inversion of the acoustic scattering model, dissipation rates have been remotely measured in oceanic conditions by a number of groups: Ross and Lueck (2003, 2005) using narrowband systems (100, 306 kHz), and more recently Lavery et al. (2013) using broadband systems (160–600 kHz). Broadband systems use chirped pulses that sweep over a range of frequencies and, through pulse compression processing, provide higher signal to noise ratios and increased range resolution, $O(1\text{--}10)$ cm compared to narrowband systems (Lavery et al. 2003; Stanton and Chu 2008; Turin 1960). Frequencies used for broadband systems for oceanographic applications range from about 10 kHz to 1 MHz, congruent with the range of length scales at which turbulent microstructure is expected to be detectable (Lavery et al. 2013).

A standardized method to estimate turbulent vertical mixing from acoustic backscatter in different environments has not yet been established. Previous studies demonstrate good agreements between acoustic and direct microstructure observations, but are limited to specific environments with high dissipation rates [$O(10^{-4}\text{--}10^{-5})$ W kg $^{-1}$] and strong stratification

(limited to estuaries or the halocline region) (Ross and Lueck 2003, 2005; Lavery et al. 2013). Additionally, these studies only apply the acoustic model on acoustic observations and in situ measurements of dissipation rates and background stratification parameters that are temporally and spatially closely connected and lay in the range of <10 m from the transducer. The frequencies of the acoustic systems used in those studies range between 120 and 600 kHz.

In this study, we present acoustic broadband data acquired using a hull-mounted Simrad EK80 broadband (45–90 kHz) echo sounder, collected from a sill region in the northern Baltic Sea with complex salinity and temperature stratification and dissipation rates on the order of $O(10^{-9}\text{--}10^{-6})$ W kg $^{-1}$. The acoustic observations are validated with nearly coincident in situ turbulence measurements from a free-falling microstructure profiler. We find that using a broadband echo sounder with lower frequencies than previously applied enables us to detect turbulent microstructure with a hull-mounted system down to the maximum water depth in the study region of ~ 200 m. The range of our acoustic observations is therefore about an order of magnitude larger than in previous efforts. Another potential advantage of the lower frequencies used in this study is that the acoustic observations become less sensitive to small particles (e.g., suspended particles, plankton).

Here we discuss different cases of mixing (in strongly and weakly stratified parts of the water column), evaluate the applicability of the acoustic model as a function of background temperature and salinity gradients, and evaluate the potential and the limitations of our commercially available broadband echo sounder. This manuscript is a step toward establishing a robust method to quantify turbulent diapycnal mixing from acoustic observations.

2. Study area and sampling

The dataset was collected during a cruise with Stockholm University's Research Vessel (R/V) *Electra* on 21–26 February 2019 in the Sea of Åland (Fig. 1). Acoustic broadband data were repeatedly collected along a transect across a sill in the southern part of the study region (Fig. 1), combined with velocity data from a vessel-mounted acoustic Doppler current profiler (ADCP) and turbulence microstructure data from a free-falling microstructure profiler (MSS). The MSS was operated in a “tow-yo” mode from the aft deck of the ship. The approximately 1.2-km-long transect was sampled 34 times while cruising with a speed of 0.5–1.5 kt (1 kt ≈ 0.51 m s $^{-1}$) against wind and waves, thereby collecting 168 MSS profiles with a horizontal spacing of about 150–300 m.

We use absolute salinity and buoyancy frequency in the analysis according to the international TEOS-10 standard for seawater (Millero et al. 2008; Feistel et al. 2010) and in situ temperature as opposed to conservative temperature because those are the water properties that affect the propagation of the acoustic wave. Absolute salinity, in situ temperature, and buoyancy frequency profiles from all MSS casts conducted upstream of the sill (Fig. 2) show a fairly homogeneous cold and fresh surface mixed layer (at 0–20 m depth) above a halocline (at 20–40 m depth). Peak stratification of the halocline

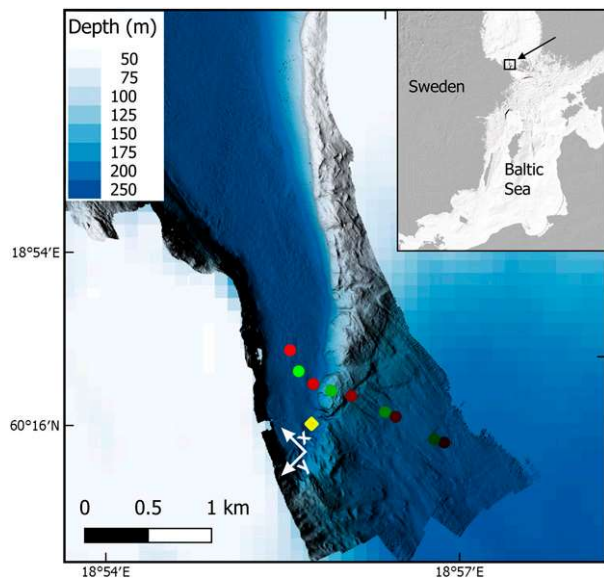


FIG. 1. Bathymetry of the study area in the Åland Sea. Dots indicate positions of microstructure profiler measurements collected during two selected transects (red and green) as shown in Fig. 4, and the yellow square indicates the position of a bottom-mounted ADCP. The deep-water flow direction is from southeast across the sill (the cross-sill and along-sill coordinates, x and y , shown in the figure are obtained by a 135° rotation with respect to the east and north directions). The detailed multibeam bathymetry data acquired by R/V *Electra* were granted public release by the Swedish Maritime Administration (release 17-03187).

reaches $N^2 \approx 2 \times 10^{-4} \text{ s}^{-2}$ at approximately 25 m depth. The layers below the halocline ($\sim 50\text{--}200$ m) are characterized by stable salinity stratification and a weakly destabilizing temperature gradient, combining into an N^2 typically slightly less than 10^{-5} s^{-2} . The deepest $\sim 20\text{--}30$ m of the vertical profiles indicate a nearly homogeneous bottom boundary layer.

In the following, we will only discuss data from a 21-h period on 25–26 February 2019, during which currents were comparatively stationary and consistently directed from southeast toward northwest across the sill (Fig. 3), allowing us to unambiguously define the lee side of the sill with a turbulent wake that will be studied in detail below (note that the tides are negligible in this part of the Baltic Sea). During this 21-h period, 20 cross-sill transects with in total 103 microstructure casts with the MSS profiler were obtained.

3. Instrumentation

Broadband acoustic records were combined with nearly coinciding in situ measurements from a free-falling turbulence microstructure profiler. ADCP data were continuously collected using R/V *Electra*'s hull-mounted system and an upward-looking ADCP moored at the bottom on the northwestern side of the sill (Fig. 1).

a. Broadband acoustics

A Simrad ES70-7C split beam transducer (Kongsberg, Norway) with a 7° circular beamwidth was mounted in the hull of R/V

Electra in the front half of the ship behind the ice knife (no sea ice was encountered during our study). A 4.1 ms pulse with transmit power of 750 W was generated with a Simrad EK80 wideband transceiver (WBT). A ping rate of 1 Hz was used throughout the study and the frequency range of the chirp pulse was 45–90 kHz with a center frequency of $f_0 = 70$ kHz, with an acoustic wavenumber of

$$k = 2\pi f_0 / c \approx 310 \text{ m}^{-1}$$

using a mean sound speed of $c = 1420 \text{ m s}^{-1}$. The acoustic wavenumbers range from $k \approx [200, 400] \text{ m}^{-1}$, corresponding the wavelengths of $\lambda \approx [1.6, 3.2] \text{ cm}$ for $f = [45, 90] \text{ kHz}$. Combined with pulse compression processing, the bandwidth of 45 kHz leads to a vertical range resolution of about 1.5 cm.

The system was calibrated in the study area with a 38.1-mm tungsten carbide sphere, following the procedure described by Demer et al. (2015). EK80 data were continuously collected in the study area during the field campaign. Position and heave of the ship were measured with a Seapath 330 + RTK GPS unit (accuracy < 1 m) and an MRU5 + motion sensor (heave accuracy 2 cm), integrated in R/V *Electra*. The acoustic backscatter signal was match filtered and corrected for absorption and spherical spreading, using MATLAB code written and provided by Kongsberg Maritime Norway (L. Andersen 2017, personal communication). After compensating for the ships draft and heave, the range of the acoustic signal was calculated using the temperature and salinity profiles that were collected closest in time.

For all calculations, we use calibrated acoustic backscatter strength per volume S_v in dB re $1 \mu\text{Pa}$, thus assuming that turbulent microstructure covers the entire volume of an acoustic match-filtered sample and is homogenous within this ensounded volume.

b. Turbulence microstructure measurements

Dissipation rates of turbulent kinetic energy (ϵ) were estimated from shear microstructure profiles obtained with a free falling MSS-90L microstructure profiler from Sea and Sun Technology (SST; Germany). The falling speed of the MSS was adjusted to approximately 0.7 ms^{-1} . A sensor protection cage allows the MSS to hit the seafloor, and thus sample turbulence and CTD data down to approximately 0.1 m above the sediment. In addition to two PNS06 airfoil probes, the MSS-90L is equipped with an internal shear sensor, precision CTD sensors and an FP07 fast thermistor. All sensors were sampled at 1024 Hz, digitized with 16 bit resolution, and transferred online to a computer on the ship. After despiking, all data were averaged to 256 Hz resolution for noise reduction, and temperature and conductivity data were corrected for different sensor response times. Vertical shear spectra were calculated from half-overlapping 256-sample Hanning windows at a resolution of 0.35 m (each window corresponding to 1 s segment length in time or approximately 0.7 m depth bins, depending on the sinking speed). Dissipation rates were obtained by integrating the vertical shear spectra, using Taylor's frozen turbulence hypothesis and assuming local isotropy in the dissipative subrange. Frequency bands in which sensor

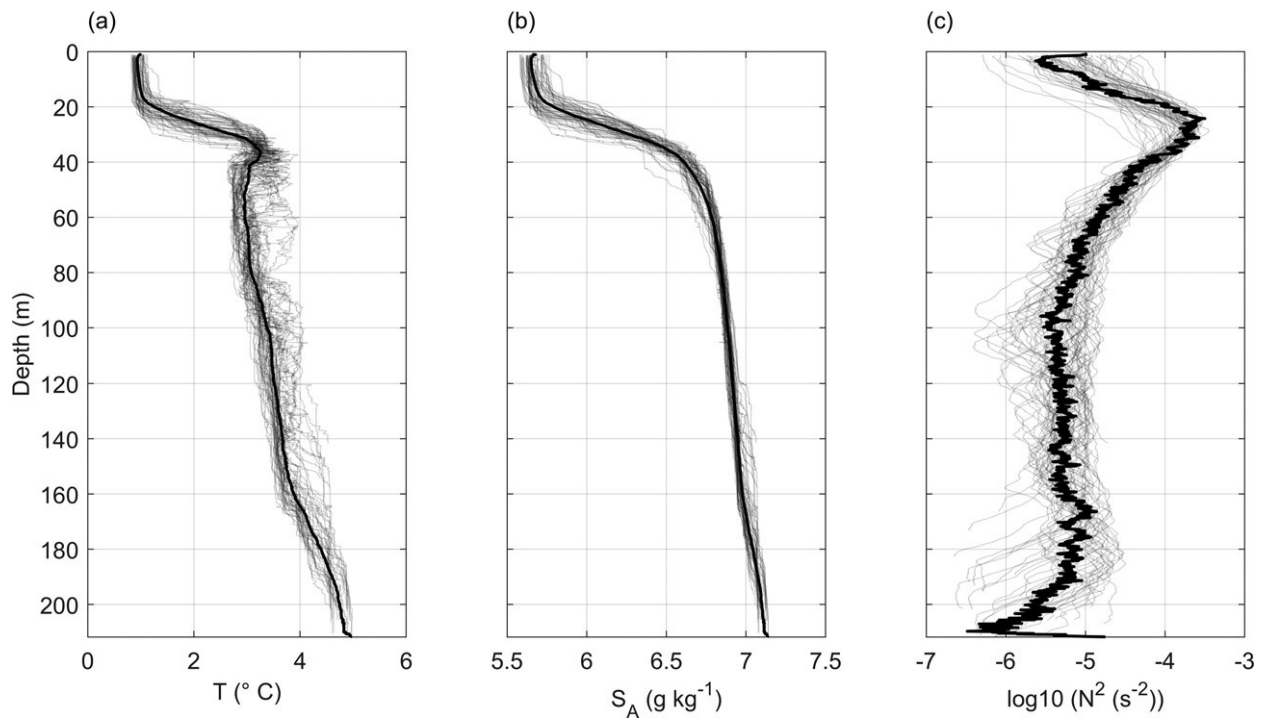


FIG. 2. Vertical variability of (a) in situ temperature, (b) absolute salinity, and (c) buoyancy frequency (squared) upstream of the sill (see Fig. 1) from MSS measurements conducted between 1830 UTC 25 Feb and 1530 UTC 26 Feb 2019 (corresponding to the period shown in Fig. 3). Individual profiles and their mean values are shown in gray and black color, respectively.

vibrations play a role were carefully removed. The upper wavenumber for integration was found iteratively as a function of the Kolmogorov wavenumber with a correction for lost variance due to unresolved scales (see, e.g., Moum et al. 1995). Dissipation rates estimated from both shear probes were finally

averaged into bins of 0.5 m thickness for further processing. This combination of raw data averaging interval, segment length for the spectral analysis, and vertical bin size for the further analysis was found to be the best compromise between statistical significance and our wish to resolve the sharp vertical

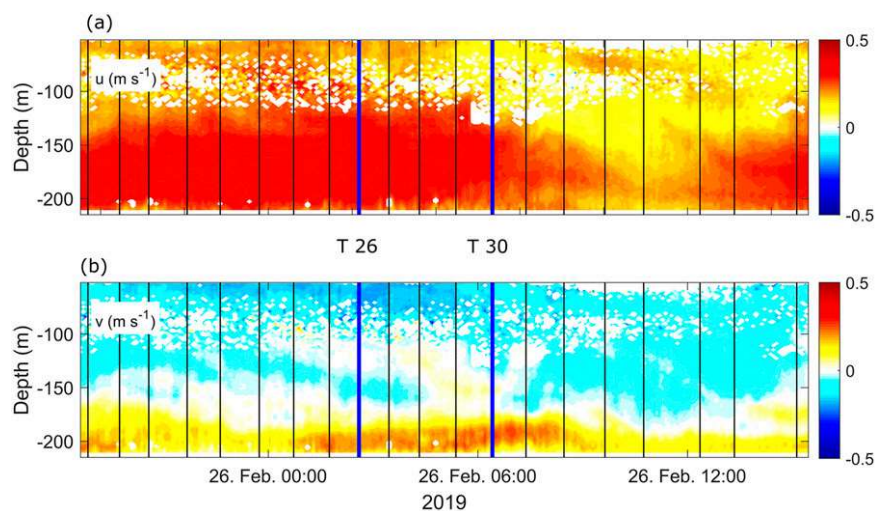


FIG. 3. Variability of (a) cross-sill and (b) along-sill velocities during the 21-h period of continuous sampling along the cross-sill transect (coordinate directions and the position of the moored ADCP are shown in Fig. 1). Black lines mark times of each transect (approximately corresponding to the time when the peak of the sill was crossed), and thick blue lines highlight the two selected transects (26 and 30) shown in Fig. 4.

gradients in the dissipation rates that are typical for our dataset. Sensitivity tests with different segment lengths and bin sizes showed that the results are not sensitive with respect to these parameters. The internal shear sensor is identical to the two external shear sensors and measures sensor vibrations from which a “pseudodissipation rate” was calculated which the profiler would measure in nonturbulent water solely due to vibrations. Data segments with pseudodissipation rates above the noise floor, which are mostly due to cable tension, were manually discarded.

We compensated for the spatial and temporal offset between the MSS casts and the acoustic backscatter observations by taking into account the lateral advection of the profiler with the currents as described in more detail in [appendix A](#). As shear-microstructure measurements are believed to be the most reliable and generally applicable technique to estimate energy dissipation in oceanic turbulence (e.g., [Lueck et al. 2002](#)), results from this approach will be considered as reference for comparison with our acoustic turbulence measurements.

c. Ship ADCP and mooring ADCP

R/V *Electra*'s 600 kHz Workhorse ADCP (Teledyne RDI, United States) was used to continuously collect data in the study area during the field campaign. Additionally, an upward looking 300 kHz Workhorse ADCP (Teledyne RDI) was deployed at 215 m depth on 22 February 2019 and recovered on 27 February 2019 at 60°16'20.21"N, 18°55'48.25"E ([Fig. 1](#)). The water column was sampled every second in 2 m bins. The raw data were postprocessed with an IOW in-house software package by quality checking and omitting data that showed absolute error or vertical velocities above 0.1 ms⁻¹, backscatter amplitude counts below 40, or a beam correlation below 32 counts. Finally, the ship ADCP data were averaged to 5 s, and the mooring ADCP data to 1 min, intervals for noise reduction.

4. Acoustic model for backscatter from turbulent microstructure

We used a theoretical model for acoustic backscatter from turbulent microstructure in the viscous-convective subrange ([Seim et al. 1995](#); [Goodman and Forbes 1990](#); [Lavery et al. 2003](#)), as described in detail in [appendixes B and C](#). The model describes acoustic backscatter σ_v as a function of the spatial wavenumber K , the dissipation rates of turbulent kinetic energy ε , temperature variance χ_T , and salinity variance χ_S :

$$\sigma_v = \frac{qK}{32} \sqrt{\frac{\nu}{\varepsilon}} [A^2 \chi_T f(\hat{K}_T) + B^2 \chi_S f(\hat{K}_S) + 2AB \chi_{TS} f(\hat{K}_{TS})]. \quad (1)$$

The parameter $\chi_{TS} = \sqrt{\chi_T \chi_S}$ represents the covariance of the gradient spectrum of T and S as suggested by [Ross et al. \(2004\)](#) in their Eq. (4.1). Further, ν is the water viscosity, and $q = 3.7$ is a spectral model parameter ([Oakey 1982](#)). The parameters A and B account for changes in sound speed and density due to changes in temperature T and salinity S , respectively, as described in more detail in [appendix B](#). The function f

accounts for the exponential decay of scalar variance due to molecular diffusion at high wavenumbers ([appendix B](#)):

$$f(\hat{K}_{T,S,TS}) = \exp(-\hat{K}_{T,S,TS}^2/2), \quad (2)$$

which is a function of the nondimensional wavenumber

$$\hat{K}_{T,S,TS}^2 = qK^2/k_{BT,S}^2 \quad (3)$$

defined based on the Batchelor wavenumbers

$$k_{BT,S}^2 = \sqrt{\varepsilon/[\nu\kappa_{T,S}^2]} \quad \text{and} \quad k_{BTS}^2 = \sqrt{\varepsilon/\left[\nu\left(\frac{\kappa_T + \kappa_S}{2}\right)^2\right]}. \quad (4)$$

Here, the molecular diffusivities of salt and heat are denoted as $\kappa_{T,S}$. For our instrument configuration (transducer and receiver are collocated), the Bragg wavenumber, where maximum backscatter is observed, is exactly twice the acoustic wavenumber k , and we will therefore replace K by $2k$ in (1) and all derived expressions.

As the flow structures determining the molecular smoothing rates χ_T and χ_S of temperature and salinity variance are notoriously difficult to observe in energetic turbulent flows due to their extremely small scales, we use the [Osborn and Cox \(1972\)](#) and [Osborn \(1980\)](#) relations to express $\chi_{T,S}$ as functions of the dissipation rate and the background temperature and salinity gradients (using in situ temperature T and absolute salinity S_A):

$$\chi_{T,S} = \frac{2\gamma\varepsilon}{N^2} \left(\frac{dT_A}{dz} \right)^2. \quad (5)$$

where $\gamma = -G/\varepsilon$ is the flux coefficient, comparing the vertical buoyancy flux G to the dissipation rate. The overbar in (5) denotes an “appropriate” vertical averaging operator. With this model for χ_T and χ_S , our approach is formally limited to stably stratified shear flows away from boundaries, which excludes the surface and bottom boundary layers. This condition is satisfied in regions where the Ozmidov length $L_O = \sqrt{\varepsilon/N^3}$ is smaller than the distance to the boundary, implying that the size of vertical overturns is controlled by stratification rather than by the distance to the boundary. Despite evidence for significant variability in γ from previous studies ([Moum 1996](#); [Garanaik and Venayagamoorthy 2019](#)), we use a constant $\gamma = 0.2$ here, following the recent recommendation of [Gregg et al. \(2018\)](#). We compared the values of χ_T computed from (5) with direct observations from our FP07 fast thermistor data (based on fits to theoretical scalar spectra), which showed good agreement in the pycnocline region that will be the main focus of this study.

By inserting (2)–(5) in (1), $f(\hat{K}_{T,S,TS})$ and $\chi_{T,S}$ can be replaced and (1) can be rewritten as

$$\sigma_v = \frac{q\gamma k \sqrt{\nu\varepsilon}}{8N^2} \left[A \left(\frac{dT}{dz} \right) e^{-2qk^2\kappa_T/\sqrt{\nu\varepsilon}} + B \left(\frac{dS_A}{dz} \right) e^{-2qk^2\kappa_S/\sqrt{\nu\varepsilon}} \right]^2, \quad (6)$$

corresponding to Eq. (4) in [Ross and Lueck \(2005\)](#). We use the Bragg scattering relation $K = 2k$ (see above) and evaluate

the acoustic wavenumber k at the center frequency f_0 of the broadband signal (see [section 3a](#)), and we calculate the molecular viscosity ν based on the nearest MSS profile according to [Sharqawy et al. \(2010\)](#).

As (6) has no analytical solution, we follow the suggestion of [Ross et al. \(2004\)](#) and solve numerically for ε . When neglecting the exponential terms (see [appendix C](#)), we can obtain ε by rearranging (6):

$$\varepsilon = \sigma_v^2 \frac{(8N^2)^2}{(q\gamma k)^2 \nu} \left[A \left(\frac{dT}{dz} \right) + B \left(\frac{dS_A}{dz} \right) \right]^{-4}. \quad (7)$$

To relate the dissipation rates inferred from acoustic backscatter to those of the in situ MSS measurements, we calculate the horizontal displacement of the profiler using the ADCP data as described in [appendix A](#). We then average over ± 5 vertical acoustic profiles (pings) around the path of the profiler, smooth the time average over depth with a Gaussian filter using a 50-point window (equivalent to approximately 0.5 m) for noise reduction and then use this as σ_v in Eqs. (6) and (7).

5. Results

In the following, we compare the acoustically inferred dissipation rates derived with the method described above with our in situ turbulence microstructure measurements in order to assess the reliability and limitations of the acoustic approach, and to investigate its sensitivity with respect to implementation details.

[Figure 4](#) shows typical measurements of acoustic backscatter collected along two selected cross-sill transects (indicated in blue in [Fig. 3](#)), and the corresponding positions of the MSS profiles (marked on map in [Fig. 1](#)). During these measurements (and all other transects discussed in the following), the currents across the sill were directed toward northwest (i.e., from left to right in [Fig. 4a](#)) with typical flow speeds of $0.3\text{--}0.4 \text{ m s}^{-1}$ at the sill level ([Fig. 3](#)). On the lee side of the sill, the backscatter patterns in the lower part of the water column ([Fig. 4a](#)) show a stratified hydraulic flow with a pronounced downstream relaxation regime and particularly strong backscatter signals in the lowest 100–150 m of the water column. As the focus of this manuscript is on the technical aspects of acoustic turbulence observations, we will not analyze the physical processes in this region in any more detail here.

[Figure 4](#) also shows enlarged views of a number of subregions representing distinct dynamical regimes with specific backscatter characteristics. Especially interesting are the approximately 10-m-thick shear layer with pronounced Kelvin–Helmholtz billows in the stratified region below the surface mixed layer ([Fig. 4c](#)), and the approximately 100–150-m-thick turbulent wake region on the lee side of the sill ([Fig. 4e](#)). As both types of flow, Kelvin–Helmholtz instability and stratified flow across a sill, represent prototypical turbulent flow regimes that have been frequently studied with the help of acoustic methods (e.g., [Farmer and Armi 1999](#); [Moum et al. 2003](#); [Lavery et al. 2010](#)), we will especially focus on the

analysis of turbulence backscatter signals from these regions in the following section.

Before turning to this analysis, it is important to note that elevated backscatter signals in other regions of the transect are most likely *not* predominantly associated with turbulence microstructure. [Stranne et al. \(2017\)](#) demonstrated that stratification can contribute significantly to the acoustic backscatter signal. This seems so be the case in parts of the halocline with particularly strong stratification, especially when turbulent microstructure is weak ([Fig. 4b](#)). Below the halocline, in the depth range between 50 and 100 m much of the backscatter is dominated by large amounts of single-target scatterers (e.g., fish, zooplankton); see [Fig. S1](#) in the online supplementary material. Such single-target scatterers can be distinguished from other sources of backscatter due to their typical bowl-like shape ([Fig. 4d](#)), caused by the movement of the transducer over the scatterers. We excluded this depth range between 50 and 100 m from our analysis ([Fig. 4](#)). Note that single-target backscatter does affect also most of the other regions but does not dominate the signal there.

Acoustic backscatter from plankton was minimized by scheduling the expedition in February, i.e., before the start of the spring bloom in the northern Baltic proper. We cannot exclude the possibility of plankton contributing to the total backscatter strength, but the strong correspondence between turbulent patches in both the MSS casts and the backscatter signal (see below), and the complete suppression of the latter in nonturbulent regions, suggests that plankton is not a relevant factor in our dataset.

Acoustic backscatter from suspended sediment might contribute to the total backscatter in this region, especially in the lower part of the water column on the lee side of the sill where turbulence is large. Maximum near-bottom current speeds measured during this cruise were about 0.4 m s^{-1} ([Fig. 3](#)), which could lead to erosion of grains with a diameter of up to 1 mm according to the Hjulström curve ([Hjulström 1935](#)). Particles with 1 mm size are in the Rayleigh scattering regime of the ES70 transducer with a wavelength of approximately 2 cm at its center frequency. Based on our findings, a major part of the observed backscatter signal in this region is most likely due to turbulence microstructure (see [section 5b](#) below) but we cannot exclude the possibility of resuspended sediment contributing to the total backscatter strength.

a. Application of the full acoustic model:

Turbulent shear instabilities

MSS profile 175 (marked in red above the echogram in [Fig. 4a](#)) includes a classic case of an unstable shear layer characterized by Kelvin–Helmholtz billows inside a strong density interface (here: a halocline) bounding the surface mixed layer from below ([Fig. 4c](#)). We will focus in the following on the applicability and limitations of acoustic microstructure observations for the quantification of turbulence rather than on the physical analysis of this type of shear layers, which have been described in detail previously (e.g., [Geyer et al. 2010](#)).

Inferring dissipation rates based on (6) or (7) requires the determination of “representative” temperature and salinity

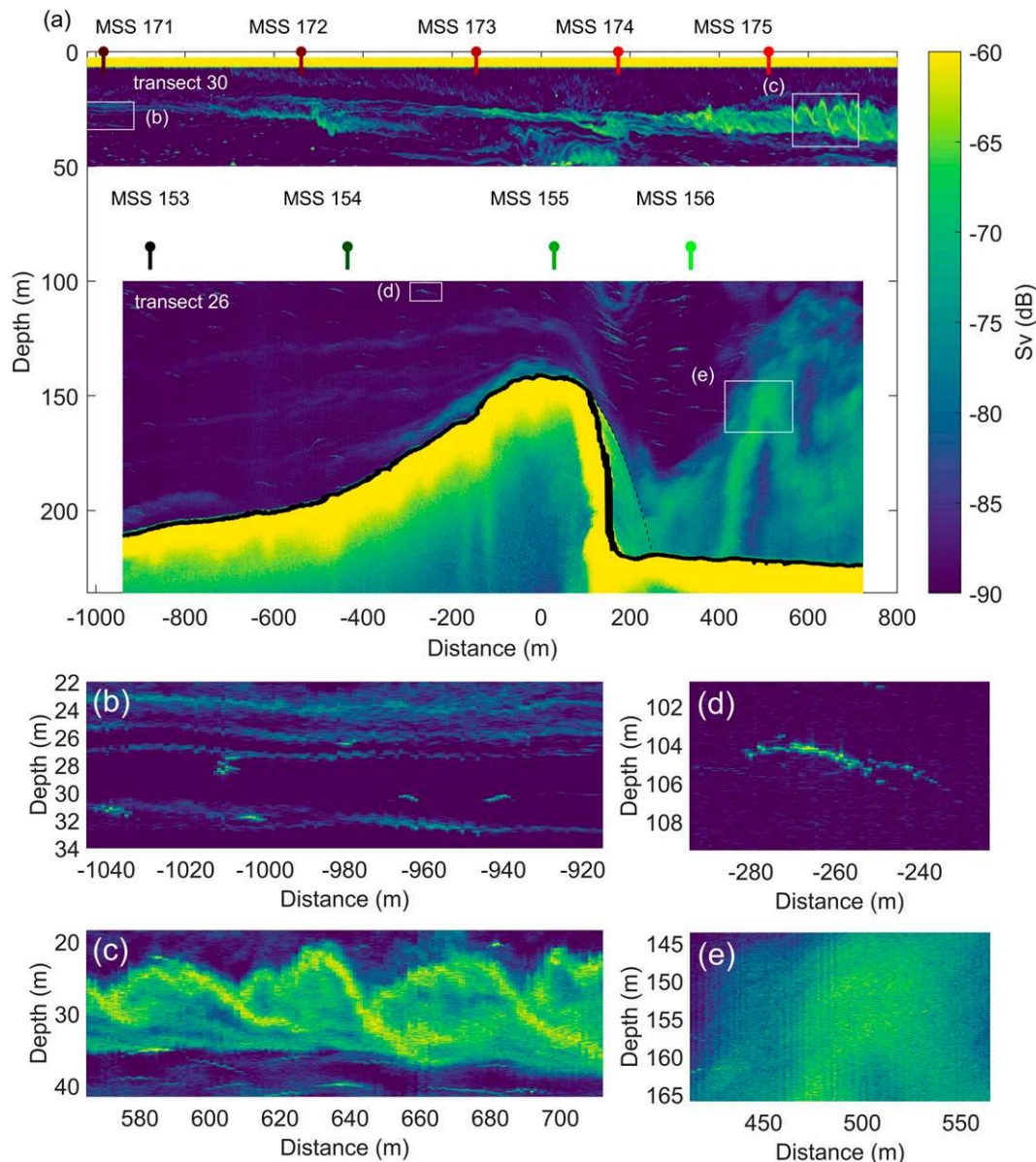


FIG. 4. (a) Acoustic backscatter (S_v) per ensonified volume in decibels (dB) as measured during transects (top) T30 and (bottom) T26 across the sill (transects marked in blue in Fig. 3). MSS profiles are indicated at the top of the panels in red and green (positions are shown as red and green dots in Fig. 1). General flow direction was from southeast to northwest (left to right). Seafloor bathymetry is marked by a black line and sidelobes from steep seafloor bathymetry are marked by a black dashed line. White boxes indicate insets showing (b) combination of stratification and turbulence microstructure, (c) turbulence microstructure associated with Kelvin-Helmholtz instabilities, (d) fish, and (e) turbulence microstructure in the wake of the sill.

gradients, which, as shown below, is a critical step in the analysis. In this example, we computed these gradients from 2 m vertical finite differencing after low-pass filtering the raw data (available at 0.2 m resolution, see above) with a 2 m box filter. This value was found to be a good compromise between vertical resolution and suppression of small-scale features. As explained in more detail below, the method is, however, sensitive with respect to the filter width, indicating that this parameter must be chosen with great care.

Figure 5b compares the acoustically inferred dissipation rates to those obtained from the MSS microstructure profiler inside an approximately 20-m-thick halocline region (region II in the following) just below the mixed layer.

At the mixed layer base (at approximately 18–22 m depth), where shear-generated turbulence is crucial for entrainment and mixed layer deepening, the acoustic model is in excellent agreement with the direct shear-microstructure observations. The acoustic model also reflects the strongly enhanced turbulence in

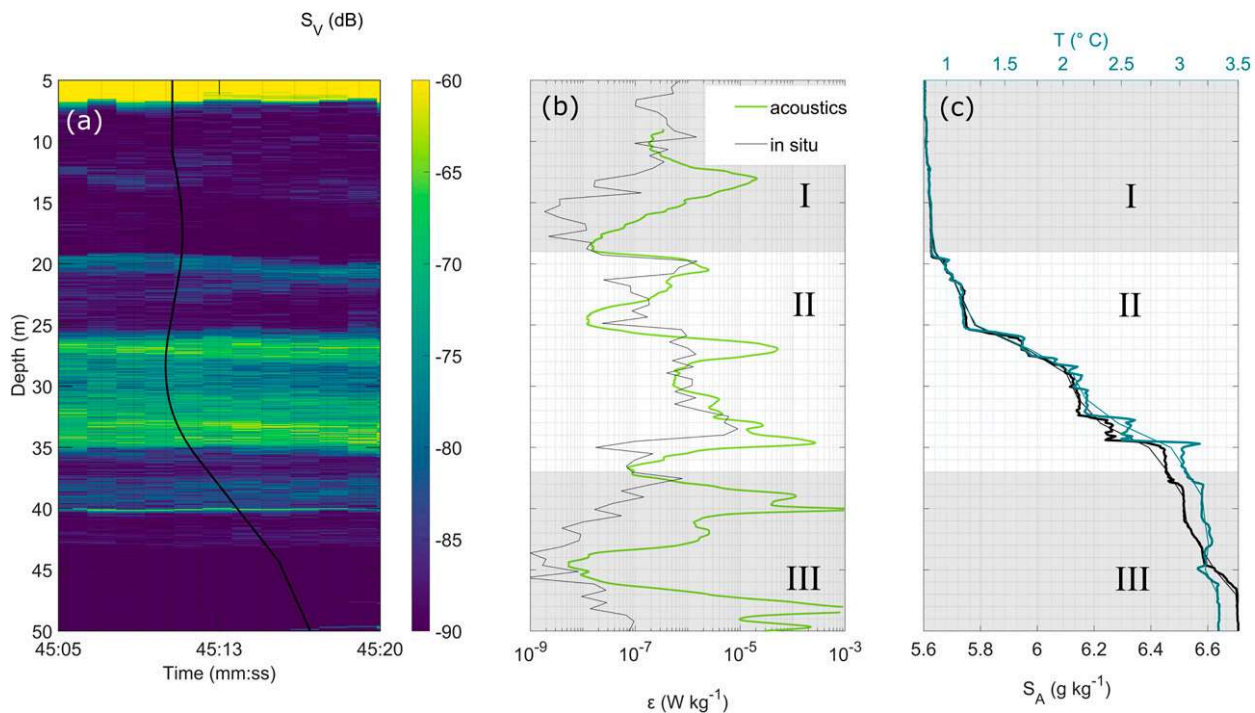


FIG. 5. Vertical variability of acoustic, hydrographic, and mixing parameters for MSS profile 175 in the vicinity of the Kelvin–Helmholtz instability shown in Fig. 4a: (a) 15 s time window of acoustic backscatter, where the black line shows the path of the microstructure profiler estimated from ADCP measurements (ship and mooring), (b) dissipation rate from the MSS profiler (black) and inferred from acoustic backscatter (green) based on (6), using the average of 11 pings surrounding the profiler path, (c) absolute salinity and in situ temperature from the MSS profiler (thin lines show the 2 m filtered data). Regions marked in gray in (b) and (c) are located outside the “halocline region” (region II, diagnosed from the gradient detection method described in the text), where the acoustic approach is likely to fail due to small temperature gradients.

the deeper Kelvin–Helmholtz billow region (24–36 m depth), but slightly overestimates the dissipation levels especially in the upper and lower flanks of the shear layers around 27 and 34 m depth, respectively. It is known (Geyer et al. 2010; Lavery et al. 2013) that these zones are characterized by enhanced vertical density gradients and secondary shear instabilities that combine into local maximal in temperature and salinity microstructure production and dissipation, χ_T and χ_S . According to (1), this is reflected in layers with locally enhanced acoustic backscatter. Such layers are clearly visible also in our backscatter data (Fig. 5a) at approximately 27 and 34 m depth, collocated with locally enhanced T and S gradients (Fig. 5c). These local backscatter peaks are not (27 m depth) or only weakly (34 m depth) reflected in enhanced dissipation rates, suggesting that the enhanced T/S gradients, appearing in the denominator of (7), at least partly compensate for the peak in σ_v in the numerator. This compensation effect is, however, not sufficiently realized in these peak regions. Figure 5b shows, e.g., that Eq. (7) translates the locally enhanced acoustic backscatter at 27 m into a local dissipation peak, which, however, is an artifact not seen in the turbulence microstructure data. A more detailed analysis shows that the strong T/S gradients in this region are too strongly smoothed by our 2 m low-pass filter, and thus too small to fully compensate for the enhanced local backscatter.

The influence of the filter width is studied in more detail in Fig. 6, comparing estimated and directly observed dissipation

rates and temperature gradients in the halocline region II for vertical filter widths that vary over an order of magnitude between 0.5 and 5 m.

The results using the largest filter width of 5 m (Figs. 6a,b) are largely consistent with the above analysis. While the temperature gradients at the upper and lower flanks of the shear layer are still marginally resolved at a resolution of 2 m (Fig. 6d), these features are no more visible at 5 m resolution (Fig. 6b). Similarly, also the collocated gradient regions in salinity (not shown) are almost completely smoothed out at this resolution. As a result, the artificial dissipation peaks in these regions are even more pronounced (Fig. 6a).

This problem cannot be overcome by reducing the filter width. When decreasing the filter width to 0.5 m (Figs. 6e,f), sharp T/S gradients are resolved more accurately, however, at the expense of strongly increased noise in the estimated dissipation data. Moreover, numerous zero crossings in the temperature gradients at this resolution lead to singularities in the acoustically inferred dissipation rates. As the slight misalignment between the acoustic and the in situ measurements becomes increasingly relevant at higher resolution, part of the noise may also be attributed to an offset between the gradients (from the MSS profiler) and the backscatter signal (from the EK80 echo sounder).

A physically motivated explanation for the good agreement at 2 m resolution is the following. The largest vertical scale of

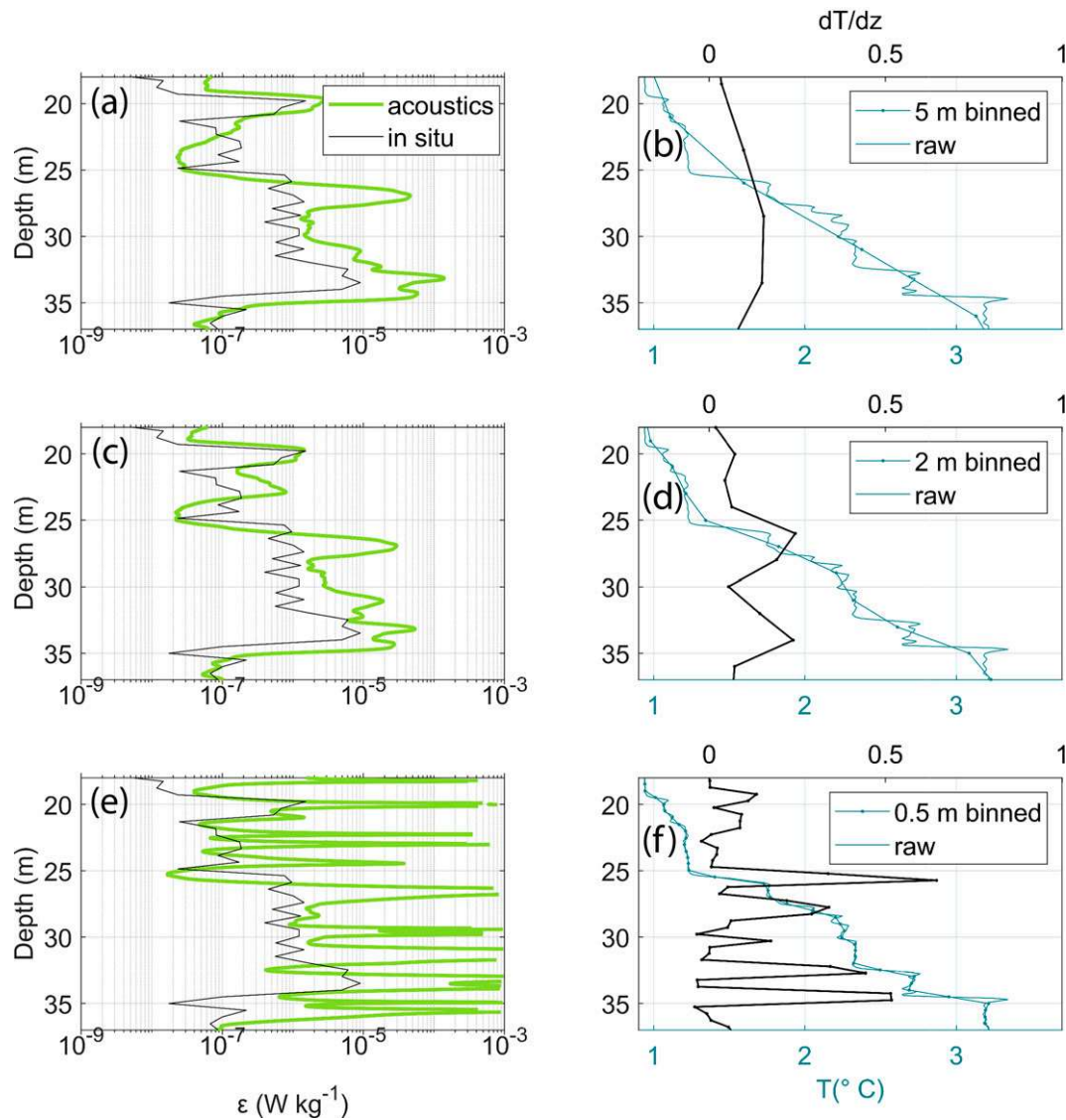


FIG. 6. Comparison of the acoustic model in (6) and MSS measurements for the halocline region II (see Fig. 5) for different vertical filter sizes. Shown are (left) dissipation rates and (right) temperature in turquoise and its gradient in black for (a),(b) 5, (c),(d) 2, and (e),(f) 0.5 m filtering.

turbulent motions is commonly estimated with the help of the Ozmidov scale, $L_O = \sqrt{\varepsilon/N^3}$. The maximum values for this length scale are on the order of 1 m for the depth range shown in Fig. 6. Stratification data obtained at scales smaller than or comparable to L_O are therefore likely to contain turbulent motions, which may explain the strong fluctuations visible in Figs. 6e and 6f. Vice versa, gradients computed at scales much larger than L_O are underestimated and, according to (6) and (7), dissipation rates overestimated, consistent with the findings above. The 2 m filter size used in Figs. 6c and 6d is therefore a physically justified compromise between resolution and the requirement to eliminate the signatures of turbulent motions in the computed gradients.

Figure 5 also shows that, different from the good agreement in region II, the model performance is much less satisfying in

the surface mixed layer (region I). In this nearly well-mixed region, T/S gradients are small (Fig. 5c), and therefore the acoustic backscatter signal (Fig. 5a) is weak due to the lack of perturbations in temperature and salinity in the ensonified volume, despite the relatively large dissipation rates measured by the microstructure profiler (Fig. 5b). Even in cases where the Osborn (1980) relation would hold and (5) would be applicable, this combination of small backscatter in the numerator and small T/S gradients in the denominator of (7) results in large uncertainties in the estimated dissipation rate from the acoustic model (Fig. 5b). This problem cannot be overcome in any systematic way by modifying the vertical filter width, suggesting that the acoustic approach is not applicable in nearly well-mixed water. A similar problem can occur also in other regions where either the T or S gradient, or both, are

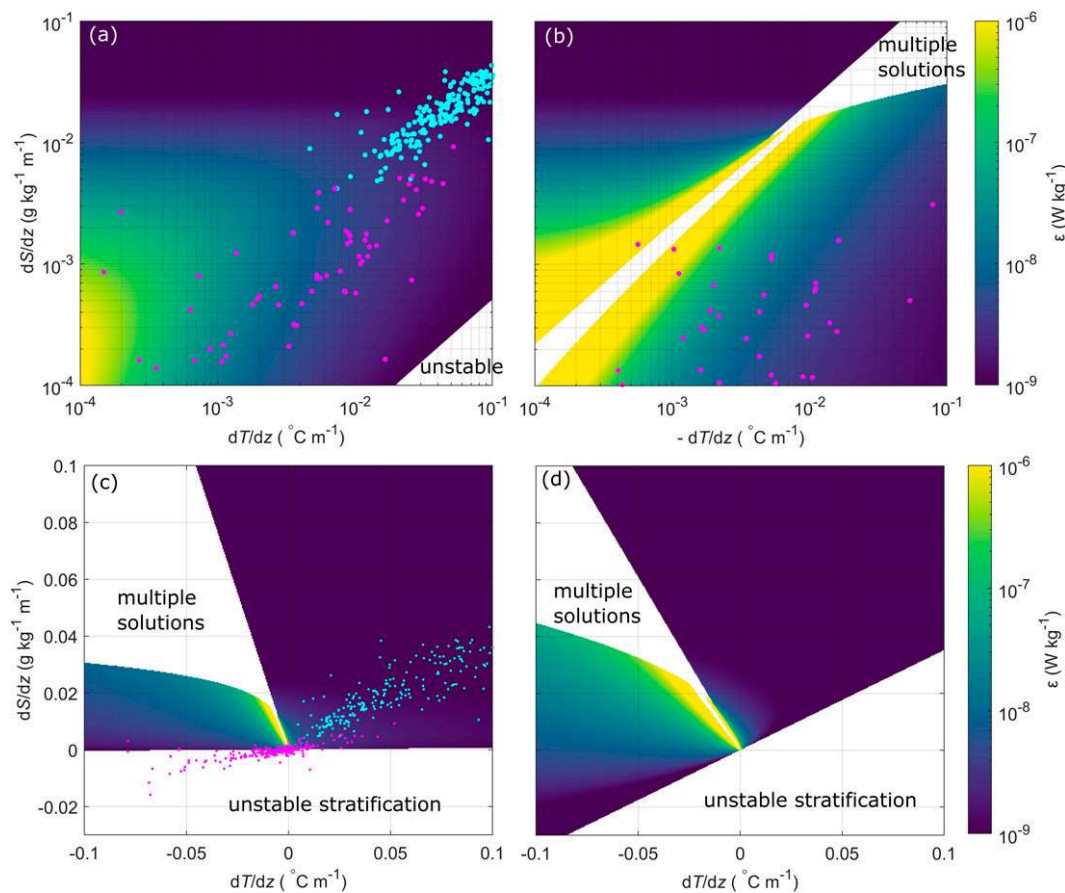


FIG. 7. Applicability and detection level of the full acoustic model, assuming a detection threshold of -80 dB for the acoustic system. Color code shows the minimum dissipation rate that can be reliably detected at that combination of T , S gradients. Logarithmic range of (a) positive T and positive S gradients, white corner marks regime of unstably stratified water, (b) negative T and positive S gradients. (c) Linear range of positive and negative T , S gradients. Cyan dots represent data points in the halocline region as used in Fig. 8, and magenta dots represent data points in the lower part of the water column as used in Figs. 9a, 9b, and 10a. All data points are based on 2 m binned T , S gradients. (a)–(c) Under conditions found in the study area during the campaign ($T = 3^{\circ}\text{C}$, $S_A = 6 \text{ g kg}^{-1}$) and (d) example for typical ocean conditions of $T = 20^{\circ}\text{C}$ and $S_A = 35 \text{ g kg}^{-1}$.

small. In region III, just below the halocline (see Fig. 5b), this situation is encountered in the weakly stratified layers of a few meters' thickness at approximately 40 and 47 m depth, respectively, where both T and S gradients based on the 2-m filtered data become vanishingly small (Fig. 5c). Similar to the surface layer, acoustically estimated dissipation rates in these regions become unrealistically large, and (6) and (7) may predict artificial dissipation peaks in regions where the backscatter is homogenous and close to the noise level (an example is the dissipation peak around 47 m depth).

The above findings suggest that the acoustic model can only be applied in certain temperature and salinity gradient regimes. To investigate this further, we defined the needed backscatter strength to reliably distinguish a signal from background noise to $S_v = -80$ dB, which is approximately 10 dB above the noise floor of our system in the conditions of the measuring campaign. Figure 7 shows how much dissipation is theoretically needed to cause a backscatter signal of -80 dB

over a range of temperature and salinity gradients, based on Eq. (6), in the conditions of our study and at the acoustic wavenumber of the center frequency of our echo sounder. It is straightforward to apply this analysis also for other datasets by adjusting the parameters in (6) accordingly. For negative temperature gradients in combination with certain salinity gradients, the acoustic model has multiple solutions for a given backscatter strength σ_v due to the nonlinearity of (6). Examples of nonmonotonic solutions of backscatter strength σ_v as a function of dissipation rates ε are shown in the supplementary material (Fig. S6). The acoustic model (6) cannot be reliably applied in regimes with multiple solutions as well as for combinations of temperature and salinity gradients that lead to unstable stratification (marked in white in Fig. 7). Small gradients in temperature and salinity lead as expected to a high detection limit of dissipation rates and a strong sensitivity of the method, therefore only large dissipation rates can be reliably quantified in this regime (Figs. 7a,b show a

range of gradients on a logarithmic scale for positive and negative temperature gradients, respectively; Fig. 7c shows the entire range of gradients on a linear scale). The detection limit and sensitivity of the method depend, through the thermodynamic parameters A and B (defined in section 4), on the temperature, salinity, and pressure of the study environment. While Figs. 7a–c show the applicability of the acoustic model for the environmental parameters in our study, Fig. 7d represents an example of typical ocean conditions.

Based on the above findings, we describe in the following an automatic detection algorithm that eliminates negative as well as small temperature gradients and thus allows for a straightforward application of the acoustic approach in the halocline region. To study the robustness and reliability, this algorithm will be applied to all available microstructure profiles (MSS114–MSS216) conducted during the cruise on transects across the sill.

The algorithm basically identifies the region of the halocline, which is in all profiles at a range between approximately 10 and 45 m depth. Based on the measured temperature, salinity, and buoyancy frequency profiles (Figs. 2a,c), the (inverse) temperature gradient has no significant effect on N^2 due to the small ratio of the expansion coefficients ($\alpha/\beta \approx 1/200$, for $S_A = 6 \text{ g kg}^{-1}$, $T = 3^\circ\text{C}$, $p = 50 \text{ dbar}$), implying that stratification is largely determined by the stable salinity gradient. Conversely, the temperature gradient is the dominating factor in the denominator of the acoustic model in (7). This somewhat surprising finding follows from the fact that, for the thermodynamic parameters encountered in our study, the factor A multiplying the temperature gradient in (7) is approximately 2 times larger than the factor B of the salinity gradient (see Table A1). Considering the fourth power in the dominator of (7) implies that the weighting factor of the temperature gradient is about a factor of $(A/B)^4 \approx 16$ larger compared to that of the salinity gradient. From this, and the larger relative magnitude of the temperature gradient compared to the salinity gradient (see the axes scales in Fig. 5c), it follows that temperature variations almost exclusively determine the denominator in (7), and thus the sensitivity of the method in regions with small gradients.

To eliminate such regions with prohibitively small temperature gradients, we used a temperature gradient threshold of $0.02^\circ\text{C m}^{-1}$. This value was determined through iterative testing and the results were found to be rather insensitive between 0.001 and $0.06^\circ\text{C m}^{-1}$. The threshold criterion was implemented as follows. Based on the 2 m filtered data, we computed vertical temperature gradients from 2 m finite differencing, starting at the surface. If the gradient was above the threshold in four subsequent 2 m bins, the first of these bins was marked as the beginning of the halocline region in which the acoustic model in (6) was applied. If three subsequent bins showed gradients smaller than the threshold inside this region, or if one single gradient was negative, the first of these bins was marked as the lower end of the halocline region. Results were not strongly sensitive with respect to the number of bins.

This algorithm was applied to all 103 MSS profiles obtained during the 20 cross-sill transects marked in Fig. 3. These profiles

also included the profile shown in Fig. 5, where the selected halocline region is marked as region II. Two further examples are provided as Figs. S2 and S3 in the supplementary material. Figure S2 includes a well-defined halocline region with smooth and monotonic temperature and salinity profiles (Fig. S2c), and a compact turbulent shear layer with enhanced backscatter and energy dissipation (Figs. S2a,b). The detection method reliably identified this region, where the acoustic model in (6) also provided a good representation of the directly observed dissipation rates. This is contrasted by the more diffusive backscatter patterns shown in Fig. S3a, and the more complex temperature and salinity profiles with multiple inversions shown in Fig. S3c. Figure S3b, however, shows that also in this more complex situation our algorithm reliably detects the (small) subregion in which the directly measured and acoustically estimated dissipation rates are in good agreement.

It is therefore not surprising that the scatterplot in Fig. 8, including data from the automatically detected regions of all 103 MSS profiles, shows a correlation (Pearson correlation coefficient $r = 0.84$ calculated from logarithmic data) between the acoustic approach and the directly observed dissipation rates. This is one of our main results, suggesting that energy dissipation in the stratified region below the mixed layer can be reliably estimated from acoustic backscatter if the vertical temperature and salinity gradients are carefully preprocessed. Figure 8 also shows that the acoustically inferred dissipation rates are biased high by approximately half an order of magnitude, which may be attributed to a number of reasons including (i) an underestimation of the relevant temperature and salinity gradients due to vertical filtering, (ii) a contamination by biological (single target) scatterers like fish or plankton, (iii) backscatter from background stratification structure not influenced by turbulence, (iv) the larger volume of the acoustic beam compared to the sampling volume of the MSS (see discussion section), and (v) calibration errors that may have occurred despite the careful calibration procedure described in section 3a above.

b. Application of a simplified acoustic model: Turbulent wake on lee side of sill

As discussed above in the context of Fig. 4, the lee side of the sill is characterized by a vigorously turbulent wake reflected in strongly enhanced backscatter over the lowermost 100–150 m of the water column. Turbulence in this region shows overturns with vertical scales of $O(10\text{--}100) \text{ m}$ (Fig. S8), and turbulence parameters can vary by orders of magnitude on time scales of $O(1\text{--}10) \text{ min}$, as estimated from $N^2 \sim 6 \times 10^{-6} \text{ s}^{-2}$ and corresponding to a buoyancy period of $\sim 7 \text{ min}$. The mean buoyancy Reynolds number is $O(10^2)$ in the halocline (20–40 m depth) and $O(10^5)$ in the lower part of the water column on the lee side of the sill. Compared to the shear instabilities in the upper part of the water column, discussed in the previous section, two additional problems arise in this situation: (i) due to the large size, variability, and intermittency of the turbulent overturns in this region, it becomes difficult to define physically meaningful “representative” temperature and salinity gradients, required in the combined Osborn/Osborn–Cox model in (5) to relate the

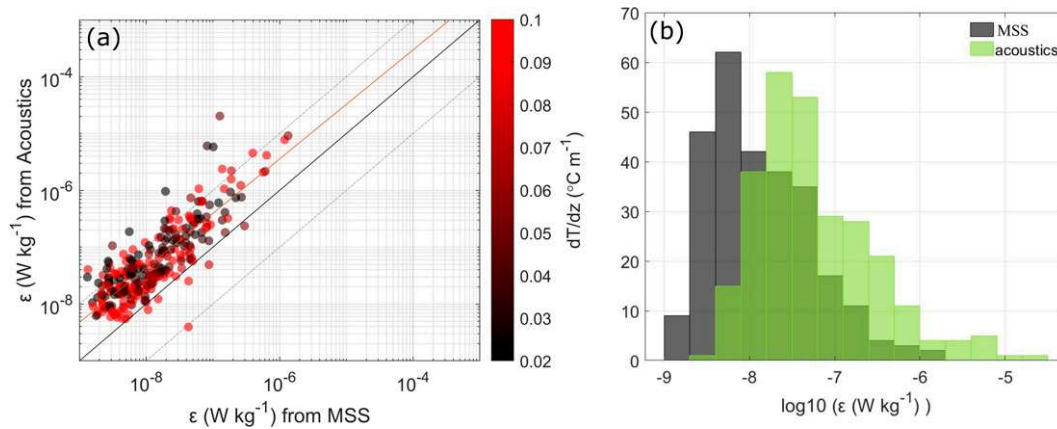


FIG. 8. (a) Comparison of dissipation rates inferred from the acoustic model in (6) and from direct MSS measurements for all MSS profiles (MSS114 to 216) obtained during the 20 cross-sill transects shown in Fig. 3. Each dot represents the geometric mean over a 5-m depth interval and is color-coded by the minimum temperature gradient in this interval, calculated from 2 m box-filtered temperature profiles. The acoustic model was only applied on regions of the profiles with stable temperature stratification, here defined by a temperature gradient larger than $0.02^{\circ}\text{C m}^{-1}$ which is aligned with the lower end of the color map representing the temperature gradients. The temperature and salinity gradients of the data points are represented in Figs. 7a and 7c (cyan dots). Red line shows the linear fit $\log_{10}(\varepsilon_{\text{acoustics}}) = 0.96 \log_{10}(\varepsilon_{\text{MSS}}) + 0.31 \log_{10}(\text{W kg}^{-1})$. The offset of $0.31 \log_{10}(\text{W kg}^{-1})$ is equivalent to a factor of 2.0 by which the acoustically inferred dissipation rates overestimate the in situ measurements in linear space. 87% of the acoustically inferred dissipation rates lay within one order of magnitude of the in situ measured dissipation rates. (b) Distribution of data points shown in (a), black bars represent in situ measured dissipation rates from the MSS, and green bars represent acoustically inferred dissipation rates.

observed energy dissipation rate to the scalar mixing rates χ_T and χ_S . Ozmidov scales in this region vary by three orders of magnitude in the range $L_O = O(1\text{--}100)$ m, making it difficult to define a relevant vertical filter width for the computation of the relevant T/S gradients; (ii) due to the larger water depth, the temporal/spatial offset between the acoustic and profiler-based turbulence measurements increases. In view of the quick evolution of stratification and mixing parameters in the wake region, it is likely that the patterns found in the acoustic data have dramatically changed when the microstructure profiler samples the same region a few minutes later. Furthermore, if the profiler is advected across the direction of the ship track, the spatial difference could lead to discrepancies due to the extremely locally confined turbulence observed here.

To analyze the performance of the acoustic method in this situation, we selected all available profiles in the wake region (19 in total) and disregarded those with an obvious mismatch between the acoustic backscatter and the profiler-based dissipation rates (6 profiles), example in supplementary material Fig. S5. This choice is, obviously, subjective, and serves the sole purpose of reducing the uncertainty introduced by the temporal/spatial difference between the MSS and acoustic measurements [point (ii) above] by ensuring that both instruments sampled the same turbulent patches. Figure 9 shows, however, that even after this preselection, no satisfying correlation between both types of measurements can be achieved for any vertical filter width (only two examples for 2 and 20 m filter width are shown, but all other choices we tried showed similar results). As mentioned under point (i) above, this lack of any significant predictive power of the backscatter model

in (6) can most likely be attributed to the failure of the Osborn/Osborn–Cox model in (5) in a flow situation that is much more complex than the shear bands investigated above. Additionally, a closer look at the temperature and salinity gradients in the lower part of the water column shows that a substantial amount of data points is in fact in a regime where the acoustic model cannot be expected to lead to reliable estimates due to the small and often negative temperature gradients (magenta dots in Figs. 7a–c represent data from selected leeside profiles as used in Figs. 9a, 9b, and 10a, here using 2 m binned T, S profiles to calculate the gradients).

Interestingly, we found that much better results can be obtained by combining all unknown parameters in (7) into a single constant:

$$\varepsilon = C \sigma_v^2 \quad (8)$$

This simplified acoustic model thus assumes that the turbulence dissipation rate is approximately proportional to the square of the backscatter strength. Figure 10a shows that this simplified model provides a strongly improved correlation between both types of measurements with a Pearson correlation coefficient of $r = 0.76$ for the selected profiles in the wake region and 74% of the acoustically inferred dissipation rates laying within one order of magnitude of the in situ measured values. We obtain $C = 7.4 \times 10^9 \text{ m}^4 \text{ s}^{-3}$ from a least squares fit of Eq. (8). The same value of C also provides a good representation of transect T26 (which we discuss in the context of Fig. 4 above), with a Pearson correlation coefficient of $r = 0.89$ and 94% of the acoustically inferred dissipation rates

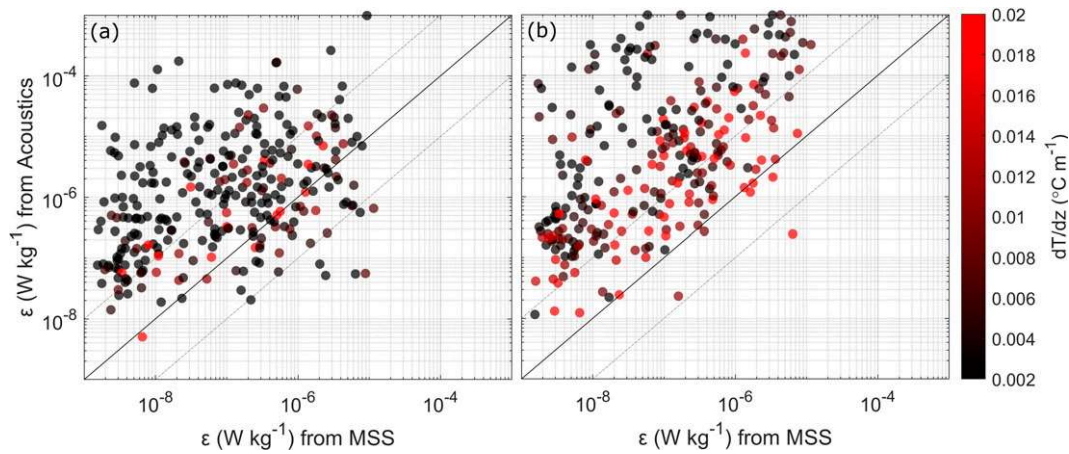


FIG. 9. As in Fig. 8, comparison of acoustically inferred and in situ measured dissipation rates where each dot represents the geometric mean over a 5 m depth interval and is color-coded by the minimum temperature gradient in this interval. Here for selected MSS profiles (138, 156, 160, 165, 169, 175, 180, 181, 187, 193, 200, 205, 211) and using (a) 2 and (b) 20 m filtered T, S profiles.

laying within one order of magnitude of the in situ measured values (Fig. 10b).

For the high-energy leeside profile MSS156 on transect T26, excellent agreement is found (Fig. 11). Two further examples are shown in the supplementary material: one where the method works fairly well (MSS138, Fig. S4) and one that was discarded because of the obvious mismatch (MSS148, Fig. S5). For comparison, we also apply the simplified model (8) to the halocline region studied in section 5a above, using all profiles that are included in Fig. 8. The simplified model yields slightly less accurate results (Fig. S7). Thus, taking the local T/S stratification into account improves the predictive skill of the acoustic model in the halocline while in the weakly stratified wake region below the halocline, the simplified model leads to better results.

The physical reasons for the applicability of the simplified model in (8) in the wake region are not entirely clear. In view of

the relatively constant upstream conditions (Figs. 2 and 3), it may be speculated that, despite the extreme intermittency visible in the individual transects, the averaged flow properties behind the sill are nearly stationary and can therefore be approximated by constant bulk parameters. To investigate this point in more detail, we processed the MSS measurements in the sill region as follows: we vertically low-pass filtered the T and S profiles as in the full acoustic model, then calculated T, S gradients and N^2 at the filter scale, and finally vertically averaged the results over the lower part of the water column (here 100 m to the seafloor). We then averaged these values over all profiles shown in Fig. 10a before calculating a representative bulk value for C from its definition in (7) and (8). For the selected profiles in Fig. 10a, we obtain $C = 4.8 \times 10^9 \text{ m}^4 \text{ s}^{-3}$ for 2 m filtered data and $C = 6.3 \times 10^{10} \text{ m}^4 \text{ s}^{-3}$ for 20 m filtered data, which bracket the value of $C = 7.4 \times 10^9 \text{ m}^4 \text{ s}^{-3}$ found above from least squares fitting. This supports our above interpretation.

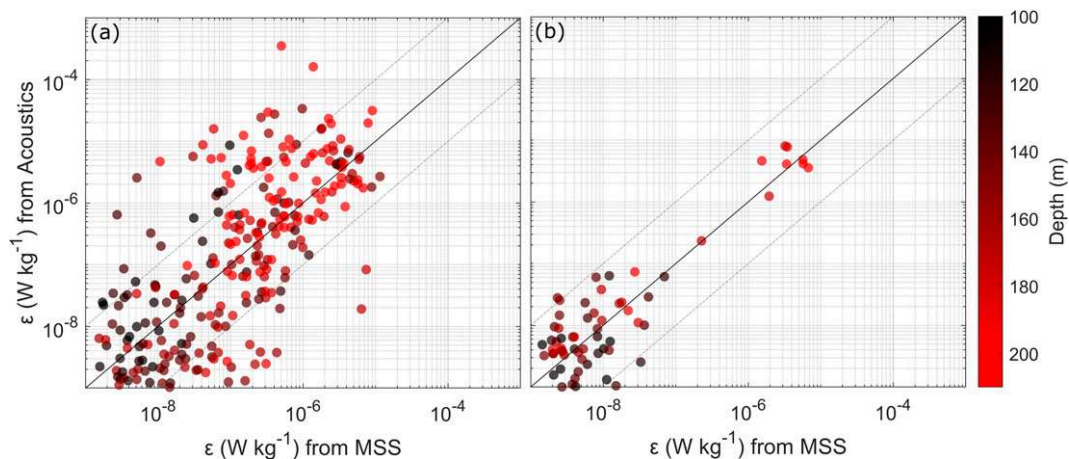


FIG. 10. As in Fig. 9, but for the simplified acoustic model (a) for all selected profiles in the wake region and (b) for all profiles taken along transect T26, MSS 153–156 (see Fig. 4).

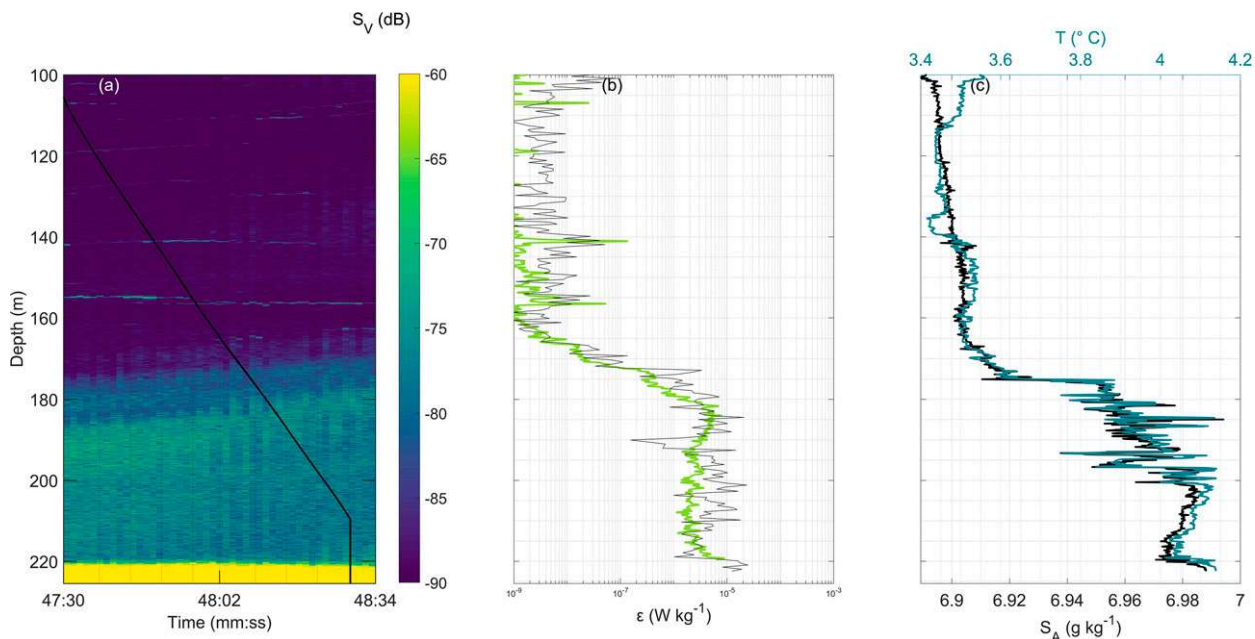


FIG. 11. As in Fig. 5, but for the lower part of the water column and MSS profile 156 (transect T26, Fig. 4) using the simplified acoustic model with constant $C = 7.4 \times 10^9 \text{ m}^4 \text{ s}^{-3}$ from fit over all selected leeside profiles (Fig. 10a).

6. Discussion and conclusions

Acoustically inferred and in situ measured dissipation rates lay within the same order of magnitude for the majority of samples in the entire water column (up to 220 m depth) when excluding the surface mixed layer and regions with high abundance of biological single target scatterers (see sections 5a and 5). We find best results when applying 1) the full acoustic model in the halocline and 2) a simplified model below the halocline.

- 1) In the halocline, most accurate and reliable results are achieved when (i) the T , S gradients in (7) are calculated based on the Ozmidov scale and (ii) the application of the full acoustic model is limited to stable stratification, with gradients above a certain threshold in the dominating parameter (T in our case). With this approach, the full acoustic model is applied in a general and automated manner on 103 MSS profiles, reaching a correlation coefficient of 0.84 between the logarithms of the acoustically inferred and the in situ measured dissipation rates. Thereby, significantly larger depth ranges are included in our study than previously shown (Ross and Lueck 2003, 2005; Lavery et al. 2013). Our findings indicate that it should be possible in the future to obtain routine turbulence measurements in stratified shear layers in the upper ocean by combining an EK80 with underway measurements of stratification parameters (e.g., underway CTD, ScanFish). In this case, the relevant length scale to filter T , S profiles would have to be determined in an iterative fashion: starting with a preliminary filter size of for instance 2 m as done previously (Lavery et al. 2013) to determine

- dissipation rates using (7), then computing the Ozmidov scale from those dissipation rates and using it as the filter size for the next iteration. Using the applicability study shown in Fig. 7, one can preselect temperature and salinity regimes in which the acoustic model leads to reliable estimates when using the temperature, salinity, and pressure of the study environment.
- 2) Below the halocline, the squared acoustic backscatter strength is often proportional to the in situ measured dissipation rates (within ± 1 order of magnitude). This simplified model is applied to a depth range from 100 m to the seafloor on (i) all 4 MSS profiles obtained during the example transect T26 shown in Fig. 4, and (ii) 13 MSS profiles from the energetic lee side of the sill in which the two instruments likely observe the same dynamics, reaching a correlation coefficient of 0.89, and 0.76, respectively. The correlation indicates that major parts of our acoustic observations below the halocline provide a direct estimate of dissipation rates. With certain limitations (see next paragraph), the acoustic data can be converted to a 2D image of dissipation rates along the ship track (Fig. 12) and thereby visualize the detailed structure and distribution of dissipation in the vicinity of a sill in the practically nontidal Baltic Sea. It is likely possible to make useful temporal and spatial integrations of acoustically derived dissipation rates but, because regions of backscatter from biological single target scatterers may result in significant errors, rigorous data cleaning and postprocessing would be needed in order to do so. While this is outside the scope of the present study, the potential in making such integrations could be investigated in future studies.

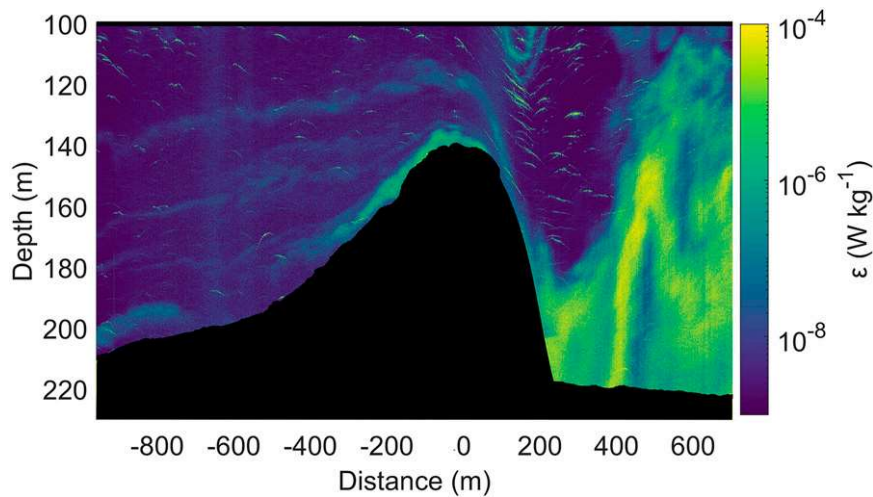


FIG. 12. Dissipation rates inferred from acoustic backscatter using the simplified model for transect T26, measured at 0217–0252 UTC 26 Feb 2019 (lower echogram shown in Fig. 4). Sea-floor and sidelobes of the sill (see Fig. 4) are marked in black. Note that the distinct, bow-shaped structures are caused by backscatter from biological single targets and need to be excluded. Furthermore, along the edges of turbulent regions only parts of the ensonified volume of the acoustic beam will capture turbulent microstructure and therefore the volume normalization of the acoustic data will likely cause an underestimation of dissipation rates and an overestimation of their spatial extent. Dark blue colored regions indicate that the observed dissipation rates reach the noise floor of the instruments.

Limitations and pitfalls of using acoustics to quantify turbulent diapycnal mixing include that 1) in order to apply the full acoustic model (6), the temperature and salinity gradients in the studied environment need to be in a regime in which the acoustic model leads to reliable results (Fig. 7). This can be particularly important for regions with temperature inversions. 2) The two instruments do not measure at the exact same time and location (see methods section). Due to the patchy and transient nature of turbulence in the study region, with dynamic flow conditions and locally fast changing salinity and temperature stratification, we expect to sometimes measure different dynamics with the two systems; 3) the lower limit of our analysis is set by the noise floor of the MSS profiler to dissipation rates of about $\varepsilon = 10^{-9} \text{ W kg}^{-1}$ which seems to approximately coincide with the noise floor of the acoustic system in this study (Figs. 9b and 11); 4) the sampling volume of the MSS profiler is several orders of magnitude smaller than that of the acoustic system as the measured acoustical signal represents an average over the volume of water ensonified by the acoustic beam. Averaging logarithmically distributed samples over different sample volumes leads to a statistical bias because larger sampling volumes increase the probability of capturing a locally high value which will then dominate the average over the entire volume (Seim et al. 1995). As dissipation rates in the study area can locally be four to five orders of magnitude above background levels, this bias can become important; 5) when normalizing for volume backscatter, we assume that the scattering source fills the entire volume of the acoustic beam. Especially at the edges of turbulent patches this assumption does not hold and therefore

the acoustic method likely underestimates the dissipation rates and horizontally overestimates the size of the patches (Diner 2001). The effects described in limitations 4 and 5 are particularly important for the lower part of the water column, as the volume of the acoustic beam increases with depth.

Furthermore, scattering sources not related to turbulent microstructure, such as suspended sediment, biological scatterers, and background stratification, can increase the acoustic backscatter signal as discussed in section 5. Their differentiation and quantification lies beyond the scope of this work for the following reasons. We lack ground-truth measurements (from, e.g., sediment cores, nets, images) to determine the grain size distribution and concentration of potentially resuspended sediments and the abundance and species distribution of biological scatterers during the measuring campaign. Our comparably narrow frequency bandwidth of the ES70 exacerbates distinguishing scatterers in frequency space—even if suspended sediments and biological scatterers would be known and modeled. The slight local and temporal misalignments mentioned previously prevent us from modeling the expected backscatter strength from background stratification in the halocline, as done in Ross and Lavery (2012), Stranne et al. (2017), and Weidner et al. (2020). The observed shear layer in this study seems to be too dynamic to estimate backscatter strength from background stratification in an automated manner with the collected in situ data.

Nevertheless, the overall good correlation between acoustically inferred and in situ measured dissipation rates indicates

that the presented method can be applied within the described limitations.

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Data availability statement. Data will be published and made accessible for downloading on the Bolin Centre Database website (<https://bolin.su.se/data>).

APPENDIX A

Alignment of Acoustic Backscatter with Microstructure Profiler Measurements

To compare our backscatter data to nearby MSS profiles, we assume that the turbulence we observe with the echosounder is stationary over the time it takes to collect data with the MSS (see discussion section). We then try to compensate for the spatial and temporal delay between the two measurements by choosing the part of the echogram that is closest to the MSS profiler measurement. One component in the delay calculations is related to the distance on board R/V *Electra* between the transducer mount position and the deployment position of the MSS profiler on the aft deck.

We further assume that the profiler does not sink straight downward (within one acoustic ping) but is instantaneously advected with the currents along the ship track, measured by the ship ADCP (about 10–40 m depth) and the moored ADCP (about 50–205 m depth) on the northern side of the sill. This implies the approximation that currents below 50 m are uniform on the measured transect. As we do not have acoustic data to the left and right of the ship track, we do not consider across track currents in this calculation. Using ADCP data, distance between the instruments, sinking velocity of the profiler (about 0.7 m s⁻¹), and ship speed during the MSS cast (0.7–1.5 kt), we estimate the delay between the profiler measurements and the acoustic backscatter

observations to align the two measurements (black line in Figs. 5, 10). Microstructure profiler data are then compared to an average of 11 s of acoustic data around the path of the profiler.

In addition to the potential error in the horizontal position of the profiler, there is a time difference between the in situ and acoustic measurements due to the finite MSS sinking velocity of about 0.7 m s⁻¹. This error increases with depth: close to the surface the time difference is about 25 s while at the maximum depth of 220 m in the study area, the time difference between the two measurements is approximately 5 min, assuming zero advection (i.e., that the profiler's path is vertical). At the bottom, the turbulence can therefore evolve for about 5 min after the acoustic observation until the same turbulent patch is measured in situ with the MSS profiler. Advection can decrease or increase this time difference depending on the current direction in relation to the direction of the vessel.

APPENDIX B

Acoustic Model for Backscatter from Microstructure

We follow the work by Ross et al. (2004), starting from their Eq. (2.1) that was originally derived by Lavery et al. (2003) based on Batchelor (1959) and Goodman and Forbes (1990):

$$\sigma_v = 2\pi k^4 [A^2 \Phi_T(K) + B^2 \Phi_S(K) + 2AB \Phi_{TS}(K)]. \quad (B1)$$

Here, σ_v denotes acoustic backscatter at the acoustic wavenumber k from isotropic turbulent microstructure of in situ temperature T and absolute salinity S_A , here for simplicity S . It is expressed as a function of the three-dimensional isotropic scalar spectra Φ_T , Φ_S of the T and S fluctuations, and their cospectrum Φ_{TS} , evaluated at the Bragg wavenumber $\mathbf{K}$. Acoustic backscatter from turbulent microstructure depends on the spectral subrange of turbulence in which the acoustic wave is scattered back to the transducer with the highest intensity. This is the case for turbulent microstructure at the Bragg wavenumber, because it causes constructive interference and thereby leads to increased backscatter intensity from this part of the spectrum. If turbulence is assumed to be isotropic, as in (B1), the dependency on the Bragg wavenumber vector reduces to a dependency on its magnitude, $K = |\mathbf{K}|$. Furthermore, for the special sensor configuration used in this study (transducer and receiver coincide), we have $K = 2k$ (see Lavery et al. 2003). The thermodynamics parameters A and B are defined as

$$A = a - \alpha = \frac{1}{c} \frac{\partial c}{\partial \Theta} \bigg|_{S_A, p} + \frac{1}{\rho} \frac{\partial \rho}{\partial \Theta} \bigg|_{S_A, p},$$

$$B = b + \beta = \frac{1}{c} \frac{\partial c}{\partial S_A} \bigg|_{\Theta, p} - \frac{1}{\rho} \frac{\partial \rho}{\partial S_A} \bigg|_{\Theta, p},$$

with the thermal expansion coefficient α , the haline contraction coefficient β , the fractional changes in sound speed,

TABLE A1. Variation of thermodynamic parameters A and B and viscosity ν within the water column in the study area.

	p (dbar)	Θ ($^{\circ}\text{C}$)	S_A (g kg^{-1})	A ($^{\circ}\text{C}^{-1}$)	B (kg g^{-1})	ν ($\text{m}^2 \text{s}$)
Surface	0	1	5	3.3×10^{-3}	1.7×10^{-3}	1.7×10^{-6}
Bottom	250	5	8	2.9×10^{-3}	1.7×10^{-3}	1.5×10^{-6}
Average	125	3	6.5	3.1×10^{-3}	1.7×10^{-3}	1.6×10^{-6}

a and b , from changes in conservative temperature Θ and absolute salinity S_A , respectively, the sound speed c and the density ρ . All thermodynamic parameters were computed according to the TEOS-10 standard (Millero et al. 2008; Feistel et al. 2010). The variations of A , B , and ν inside our study region are on the order of 10% as shown in Table A1.

The isotropic scalar spectra and cospectra $E_x(K)$ (with $x = T, S, TS$) are related to the three-dimensional spectra by (Tatarski 1961, p. 17)

$$\Phi_x(K) = \frac{E_x(K)}{4\pi K^2}. \quad (\text{B2})$$

Therefore, if we assume that the turbulence is isotropic at the observed length scales, we can insert (B2) into (B1), and express the volume backscatter σ_v as a function of the isotropic scalar energy spectra for temperature and salinity, $E_T(K)$ and $E_S(K)$, and their cospectrum $E_{TS}(K)$:

$$\sigma_v = \frac{k^4}{2K^2} [A^2 E_T(K) + B^2 E_S(K) + 2AB E_{TS}(K)]. \quad (\text{B3})$$

APPENDIX C

Spectral Shape Observed in This Study

Information about the observed regime of scalar turbulence can be obtained from the acoustic scales. The spatial cutoff wavenumber between the inertial-convective and the viscous subrange of scalar turbulence is given by (e.g., Seim et al. 1995)

$$K_{\text{cutoff}} = \frac{1}{2} \left(\frac{5}{12} \right)^{3/2} \left(\frac{\varepsilon}{\nu^3} \right)^{1/4}. \quad (\text{C1})$$

It is important to note that the cutoff wavenumber K_{cutoff} is the spatial wavenumber of the turbulent motions here, not the acoustic wavenumber.

From the highest dissipation rates of $\varepsilon = 10^{-5} \text{ W kg}^{-1}$ measured during our cruise and a viscosity of order $\nu = 10^{-6} \text{ m}^2 \text{s}^{-1}$, the cutoff wavelength is $K_{\text{cutoff}} = 240 \text{ m}^{-1}$. Therefore, the acoustic system on R/V *Electra* observed turbulence in the viscous-convective subrange and on smaller scales for the conditions encountered during the campaign. According to Ross and Lueck (2003), the temperature roll of starts at a length scale of $L \approx 2\pi(\nu\kappa_T^2/\varepsilon)^{1/4}$, which, for low dissipation rates of order $\varepsilon = 10^{-9} \text{ W kg}^{-1}$, yields $L = 1.5 \text{ cm}$, whereas for high dissipation rates of order $\varepsilon = 10^{-5} \text{ W kg}^{-1}$ the rolloff starts at $L = 1.5 \text{ mm}$. With a wavelength of about 2 cm at the acoustic center frequency, we see Bragg resonance effects at a

spatial scale of 1 cm . Therefore, we expect to see the diffusive rolloff of temperature variance only for low dissipation rates. The diffusive rolloff for salinity fluctuations occurs at even smaller spatial scales that are not relevant for the acoustic frequencies used in this study.

The isotropic scalar spectra $E_x(K)$ for $x = T, S, TS$ can therefore be expressed by the spectra for isotropic turbulence in the viscous-convective subrange:

$$E_x(K) = q\chi_x\varepsilon^{-1/2}K^{-1}\nu^{1/2}f_x(\hat{K}_x), \quad (\text{C2})$$

with the dissipation rates χ_{TS} of T and S variances, $\chi_{TS} = \sqrt{\chi_T\chi_S}$ for their covariance (Ross et al. 2004), the Bragg wavenumber $K = 2k$, the viscosity ν , and the model constant $q = 3.7$ (Oakey 1982). The function $f_x(\hat{K}_x)$ describes the exponential decay of T, S variance at high wavenumbers due to molecular diffusion:

$$f_x(\hat{K}_x) = e^{-\hat{K}_x}, \quad (\text{C3})$$

with $\hat{K}_x = \sqrt{2q}K/K_{Bx}$. Here $K_{Bx} = (\varepsilon/\nu\kappa_x^2)^{1/4}$ denotes the Batchelor wavenumber for temperature and salinity, respectively, with κ the coefficient of diffusivity. For the T, S cospectrum, we use $K_{BTS} = (\varepsilon/[\nu(\kappa_T + \kappa_S)/2]^2)^{1/4}$ as proposed in Ross et al. (2004). Inserting (C2) into (C1) leads to (1).

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