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Towed Spectra of Internal Waves in the Pycnocline of the Baltic Sea

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In this paper, the data on temperature and salinity measurements with a towed temperature and salinity profiler operated in scanning mode in the pycnocline between the Baltic (surface) waters and transformed inflow (deep) waters of the North Sea origin in the Gdansk Deep of the Baltic Sea are analyzed. The measurements were carried out on March 3, 2006. It is shown that the spectral level of vertical displacements is below the background Garrett–Munk spectrum. Such an effect is explained by the absence of tides in the hydrological regime of the Baltic Sea.

Internal waves in the ocean are generally induced by any perturbations of water particles in the water column in the presence of stratification. Wind and tides [9] are undoubtedly the strongest sources of internal waves. In 1972, Garrett and Munk contrived a universal spectral model of internal waves for the entire ocean [5, 6]. The model is based on the following basic assumptions. Fluctuations on the internal wave scale observed in the ocean are determined exclusively by internal waves and not by any other processes. The observed wave field is determined by a superposition of linear internal waves of different scales with random phases and amplitudes. The internal wave energy is continuously distributed by frequencies and wave numbers (continuum of modes) and is not related to any specific modes. The upward and downward fluxes of wave energy are equal, and the internal wave field is horizontally isotropic. The wave field in different regions of the ocean and at different time moments is approximately the same. Thus, a universal background wave field exists in the entire World Ocean. Perturbations appear over this background at different time moments and at different spatial points. The spectrum of dimensionless energy contrived by Garrett and Munk depends on the frequency (ω) and wave number (k_x, k_y, k_z).

The authors reduced the scope of the model to the middle part of the water column and excluded the upper and bottom layers, each a few hundred meters thick, in order not to contradict the model with the data of observations. They also excluded high and equatorial latitudes, as well as tidal frequencies, from consideration in the background model because significant input of energy to the internal wave range occurs at these latitudes and frequencies.

Usually, frequency spectra calculated from moored measurements are considered in the study of internal waves and in comparison of the results with the Garrett–Munk model [5, 6]. The authors of the model call them moored spectra. In our work, we shall consider rarely analyzed measurements with the instruments towed by a vessel. The spectra calculated from such measurements are called towed spectra, and they are compared with the so-called towed spectra of the Garrett–Munk model.

While contriving the model energy spectrum, Garrett and Munk deliberately excluded the upper ocean layer from the model because it seems that no universal spectrum exists for this layer owing to the strong influence of surface effects caused by the wind and surface waves. However, we can compare experimental spectra with their model presentation to estimate the energy of internal waves. Numerous attempts to obtain such spectra show that the energy of internal waves near the surface is significantly greater than the model level. Internal waves in the upper layer of the ocean differ strongly from those observed in the deep layers [7]. Many models were suggested to describe internal wave spectra in the upper ocean. For example, one of the approaches is the supposition that the energy of internal waves is the sum of the background energy and energy of the waves generated near the surface [10]. In the ice-covered sea, where the atmospheric influence is weak, the energy of internal waves is close to the model [8].

Tidal motion is an important factor in the ocean. Since tidal motion is absent in the Baltic Sea, the values of spectral densities of internal waves should be smaller

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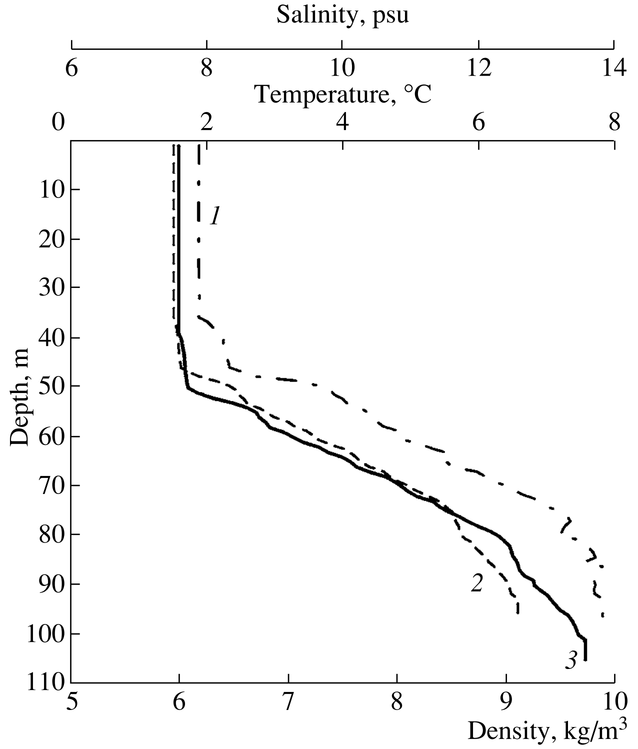

Fig. 1. Vertical profiles of (*1*) temperature, (*2*) salinity, and (*3*) density in the Gdansk Deep of the Baltic Sea in March 2006.

than in the ocean. However, the close location of measurements to the surface should increase the internal wave energy because the influence of wind generation is strong.

The results of measurements of internal wave spectra in a tidal sea (Sea of Japan) and in a tide-free sea (Black Sea) are presented in [1, 2, 3]. The main result of their research in the tide-free sea is the low level of spectral densities, which are lower than the model Garrett–Munk spectrum. Sometimes, intensification of the internal wave field in the tide-free sea leads to an increase in the spectra to the model level. The slope of the spectra is close to the model, while insignificant peaks on the spectra are related to the frequency of waveguide widening. In a tidal sea, internal wave spectrum in the upper layer always exceeds the model spectrum.

In the Baltic Sea, a sharp interface is located at a depth of approximately 50–70 m between the surface freshened waters of the Baltic origin and transformed deep waters, which inflow periodically from the North Sea. The upper approximately 40-m-thick layer is practically mixed in winter by storms and winter convection. During intense inflows of warm saline waters from the North Sea, the interface is very sharp. In the course of time, the sharp pycnocline is mixed by internal waves. Figure 1 shows characteristic changes in temperature, salinity, and density with depth in March 2006. An increase in temperature and salinity with depth is a specific feature of the parameters measured in this region.

Using the model Garrett–Munk spectrum, one can write the following form of the 1D spatial model spectrum and the amplitudes of vertical displacements recorded during towed measurements:

$$F_\zeta(k) = 0.167\frac{f}{N}k^{-2}, \qquad (1)$$

where $F_\zeta(k)$ is the towed spectrum of the amplitudes of vertical displacements (m^3), f is the Coriolis parameter, N is the Brunt–Väisälä frequency, and k is the wave number.

Such a simple expression for the spectrum is valid if the wave number is in the range of values close to 1 cycle/m. In other ranges, the relation for the spectrum is more complex. Desaubies [4] suggested the following analytical relation:

$$F_\zeta(k) = \left(\frac{2}{\pi}\right)^3 rt\frac{f}{N}\left(\ln\frac{N}{f} - \frac{N^2-f^2}{2N^2} - \right)k^{-2}, \qquad (2)$$

in which one can vary parameters $r = E\hat{b}^2\hat{N}$ and $t = \frac{j}{2\hat{b}\hat{N}}$. $E = 2\pi 10^{-5}$ is dimensionless energy per square unit; $b \approx 1300$ m is stratification depth in the ocean; N = 3 cycles/h is frequency scale; and j is number of modes.

During measurements in March 2006, the Mark III NBIS profiler was towed by the ship with a speed of approximately 5 knots (2.5 m/s) in a scanning regime between depths of 50 and 80 m. This means that the depth of the instrument was controlled so that it was continuously lowered and raised between the given limiting depths. The horizontal distance between the start and end points of the cycle was approximately 30–40 m. Processing of the measurements included the compilation of an array grid with a 10-m horizontal interval and 20-cm vertical interval (grd-file). The depth of different isotherms was calculated from this file. The 6 and 5.5°C isotherms did not go beyond the upper and lower limits of the instrument motion. Thus, the spectra were calculated from the depths of these isotherms.

Figure 2 demonstrates spectral densities of vertical displacements of the two isotherms. The values of the spectral densities are smaller than the model Garrett–Munk spectrum. The 5.5°C isotherm located closer to the surface than the 6°C isotherm shows stronger fluctuation related to internal waves because it is subjected to stronger influence of energy pumping from the surface.

Thus, the results of investigations show that the spectra of vertical displacements of towed measurements in the Baltic Sea in the interface layer between the Baltic waters and transformed waters of the North Sea origin are approximately one order of magnitude

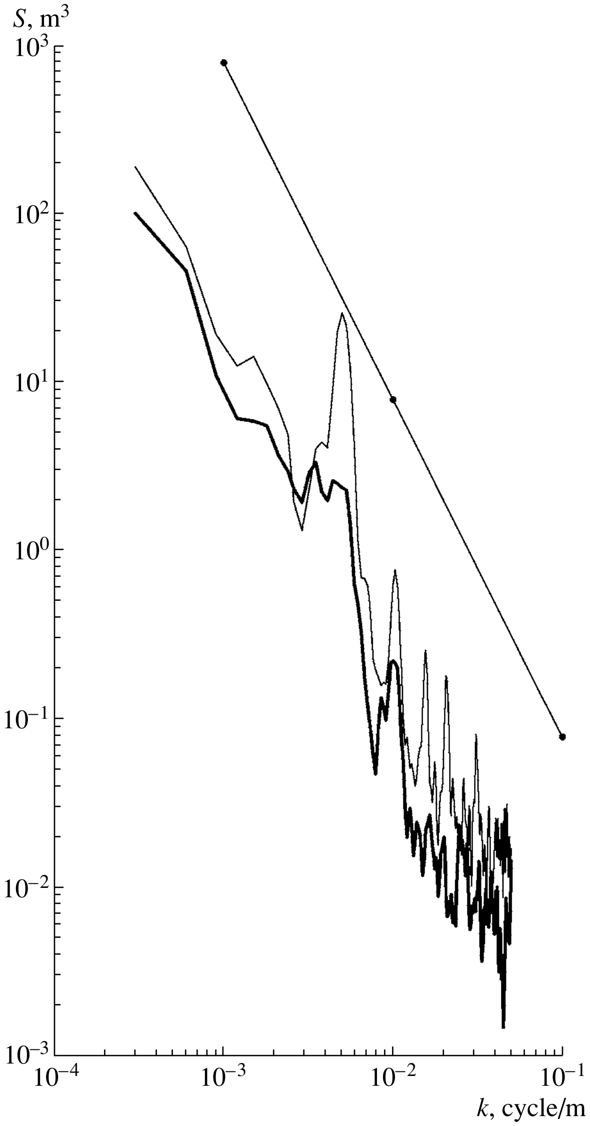

Fig. 2. Spectrum of the fluctuations of 6°C (thick curve) and 5.5°C (thin curve) isotherm depth. The Garrett–Munk spectrum is shown with a straight line.

lower than the background Garrett–Munk spectrum. Such an effect is explained by the absence of tides in the hydrological regime in the Baltic Sea, where no energy is transferred to the short-period wave range from decaying internal tides. The energy transported from the surface is not enough to maintain the background level of the Garrett–Munk spectrum in the interface layer at a depth of 50–70 m. Mixing of waters by internal waves in the pycnocline layer influences the spreading of pollution into the deep layers of bottom depressions.

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REFERENCES

1. V. A. Ivanov and A. N. Serebryanyi, Izv. Akad. Nauk, Fiz. Atmos. Okeana **21**, 648 (1985).
2. V. A. Ivanov and A. N. Serebryanyi, Izv. Akad. Nauk, Fiz. Atmos. Okeana **18**, 683 (1982).
3. A. N. Serebryanyi, Okeanologiya **25**, 744 (1985).
4. Y. Desaubies, J. Phys. Oceanogr. **6**, 976 (1976).
5. C. Garrett and W. Munk, Geophys. Fluid Dyn. **3**, 225 (1972).
6. C. Garrett and W. Munk, J. Geophys. Res. **80**, 291 (1975).
7. R. H. Kase and R. A. Clarke, Deep-Sea Res. **25**, 815 (1978).
8. M. D. Levine, Rev. Geophys. Space Phys. **21**, 1206 (1983).
9. W. H. Munk and C. Wunsch, Deep-Sea Res. **45**, 1977 (1998).
10. M. W. Rooth, M. G. Briscoe, and C. H. McComas, J. Phys. Oceanogr. **1**, 12 (1971).