Relationships between coastal processes and properties of the nearshore sea bed dynamic layer*

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Abstract

The paper discusses the notion of a layer of sandy sediments overlying a substratum of cohesive deposits in the coastal zone. This layer of sand is generally more mobile and is therefore conventionally referred to as the dynamic layer. Its parameters are important to coastal lithodynamic and morphodynamic processes caused by waves and currents. On the other hand, the dynamic layer is formed by nearshore hydrodynamic impact. The variability of the features of the dynamic layer on the southern Baltic dune and cliff shores in Poland is analysed on the basis of selected geological data supported by local seismo-acoustic field investigations. It appears that the conventional notion of the dynamic layer makes sense only in specific geomorphologic conditions. In such cases, mostly related to cliff shores, theoretical modelling of sediment transport should take the properties of the dynamic layer into account.

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1. Introduction

It is assumed in the modelling of sediment transport and seashore evolution that the resources of sand in the coastal zone are unlimited. Actually, along most southern Baltic shores, the dynamic layer, i.e. the layer of potentially mobile sandy sediments overlying a substratum of other types of deposits, is not thought to stretch far out to sea. Moreover, the thickness of this layer can be expected to be small on many stretches of shoreline. According to some investigations (see e.g. Boldyrev 1991), the thickness of the dynamic layer at the upper end of the eroded cross-shore profiles (on the emerged part of the beach called the backshore) does not exceed 2 m. On shores of this kind, the dynamic layer thickness can decrease to zero even at a distance of a dozen or so metres from the shoreline.

The properties (geometrical parameters) and the total volume of the dynamic layer are indicators of the seashore condition with its vulnerability to beach erosion and potential to produce large morphological bed forms (e.g. bars). Results of field measurements show that the existence of underwater bars, as well as their state and number, are closely correlated to the character of a coast, including the amount of accumulated sediments that constitute the dynamic layer of the nearshore sea bed. It can be roughly assumed that the presence of bars is visual evidence for the existence of the dynamic layer. Analyses carried out to date also indicate that the greater the number of bars and the higher their stability, and the greater resources of material in the dynamic layer, the thicker it is and the farther out to sea it extends (see Pruszak et al. 1999).

In the above context, the dynamic layer of the sea bed is treated as a potentially active sandy layer that can be subject to dynamic changes without any constraints. The dynamic layer can be considered at various spatial and time scales, depending on the scientific discipline and the purpose of research.

Detailed investigations of sediment motion and sea bed changes at time scales of seconds/hours/days and spatial scales of centimetres/metres relate only to the surface part of the sandy sea bed dynamic layer, which in fact can often be much thicker. The investigated sea bed layer is defined as the active layer (at a certain assumed time scale) or the mixing layer (subject to instantaneous changes). The latter is frequently equated with the nearbed sediment motion layer known as the sheet flow layer, representing a moveable sea bed under intensive hydrodynamic conditions. The thickness of the layer so defined depends mainly on actual wave-current impact, sediment features and location in the coastal zone. The maximum sheet flow layer thickness, even at greater depths (h = 15 m), can exceed 4 cm during heavy storms with a return period of 100 years (see

Myrhaug & Holmedal 2007). The sea bed surface activation (mobilization) thickness A_d increases with the wave height H and period T. Studies done to date imply a linear dependence of the depth of sediment activation on wave height. The ratio $k = A_d/H$ lies in a wide range of 0.02–0.4 (see Kraus 1985, Sunamura & Kraus 1985, Sherman et al. 1994 and Ciavola et al. 1997). As demonstrated by the above investigations, the quantity k depends on local coastal morphodynamic conditions, mostly the sea bed slope and wave energy dissipation patterns. According to measurements by Kraus (1985) for a mildly sloping sea bottom (dissipative cross-shore profile) and breaking wave conditions represented by $H_b = 0.63 - 1.61$ m and $T = 4.9 - 10.2$ s, the parameter k amounted to only 0.027. The value of k increases with increasing sea bed slope and can be ten times larger, i.e. $k = 0.27$, for a reflective seashore on which plunging wave breakers predominate (see Ciavola et al. 1997). For both of the above extreme, opposite cases, there is a distinct correlation between wave height/period and mixing depth. The relevant figures, based on numerous investigations conducted at various sites, can be found in Ciavola et al. (1997).

Available results of investigations also show that the mixing depth in the surf zone is a weakly increasing function of sediment size for a breaking wave height of < 1.5 m (see Ciavola et al. 1997 and Saini et al. 2009). Investigations carried out by the latter authors confirmed the strong dependence of the parameter k on the cross-shore profile shape and its minor dependence on sediment features. Quite unexpectedly, however, k has been found to oscillate within a small range from 0.22 to 0.26 for a wide variety of sediments (from sand to pebbles) in both stormy and non-stormy conditions.

From the geomorphological point of view, Boldyrev (1991) distinguished three major types of beach/dune shores displaying features of the dynamic layer:

- Erosive shores, with a considerable deficiency of sandy sediments, the absence of foredunes or the presence of narrow and low-crested foredunes, a narrow beach zone at the backshore (maximum 20–25 m[1]), a foreshore with no bars or 1–2 bars at most and a 0.4–1 m thick dynamic layer at the shoreline. This dynamic layer disappears near the shoreline, often at depths of no more than 3–4 m.

- Shores, along which a 'transitional' movement of sandy sediment takes place, with distinctly shaped wide and high-crested foredunes, a wide

[1]There is frequently no beach at the backshore; if there is one, it is only a few metres wide.

beach at the backshore (35–50 m), a shoreface characterized by 3–4 bars and a dynamic layer 2–5 m thick, stretching seawards to the so-called depth of closure[2].

- Accumulative shores, with extensively developed dune systems and very wide sandy backshores (60–120 m), multi-bar cross-shore profiles (with at least 4 distinct large-scale bed forms) and a dynamic layer more than 5 m thick.

Without doubt, the dynamic layer is also observed on cliff shores. Further, the notion of the dynamic layer takes on a particular significance on the shoreface of a cliff, whether active or dead. The presence of sandy (Holocene) sediments at the toe of a cliff (built of deposits older than the Holocene) makes the nearshore zone shallower and causes wave energy to dissipate as a result of breaking and bottom friction at greater distances from the shoreline. In such a situation, the cliff slope is not threatened by marine erosion and a stable beach can exist in front of the cliff, which increases the shore's value as a tourist amenity and makes it useful for recreation and coastal water sports. Most frequently, however, cliff shores have very narrow beaches at their toes or do not have beaches at all. The example of a dynamic layer in front of a cliff at Gdynia-Oksywie (Poland – KM 90.9)[3] (see Figure 1 for the location of the site) is shown in Figure 2, after Frankowski et al. (2009).

Knowledge of the features of the dynamic layer, a most important aspect of coastal geomorphology, is crucial not only for scientific investigations of nearshore lithodynamic processes but in the planning of many coastal engineering ventures as well. Knowledge of the local parameters of the coastal dynamic layer appears to be necessary with regard to artificial shore nourishment and the design of coastal protection structures. Among these structures, groynes deserve particular attention because their accumulative effectiveness (and thus also practicability) depends on the supply of sediment to the shore section requiring protection. The features of the dynamic layer are one of major indicators of this supply.

Identifying the thickness and offshore range of the dynamic/active layer also plays an important role in the optimization of solutions for laying cables and pipelines at the sea-land interface. These objects should be dug into

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[2]The depth of closure (the depth 'closing' the cross-shore transect) determines the maximum depth at which the sea bed remains unchangeable. In southern Baltic Sea conditions, it is assumed equal to 15–20 m at the long-term scale and 7–8 m at the medium-term scale.

[3]Location in the longshore coordinate system used by the Maritime Offices in Poland (KM 0.0 stands for the Polish–Russian border).

Figure 1. Location of the study sites

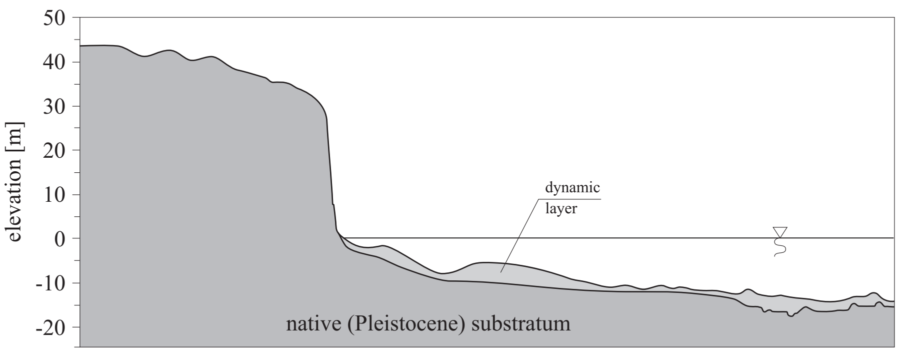

Figure 2. Dynamic layer at the cliff toe at Gdynia-Oksywie, Poland (KM 90.9), after Frankowski et al. (2009)

the nearshore sea bed sufficiently deep to resist long-term hydrodynamic (wave-current) forcing. In order to carry out a proper design process, one ought to know not only the erosive or accumulative tendencies in long-term coastal evolution but also the parameters of the nearshore layer of sandy sediments, which are the most vulnerable to scouring by nearbed wave-induced oscillatory flows and wave-driven steady currents.

The importance of the above issue, together with the availability of new measuring instruments, has become an inspiration and encouragement to carry out new fundamental studies on the characteristics of the dynamic layer and to determine their links to background morphodynamic processes taking place in the conditions of the dissipative, multi-bar, sandy southern Baltic shore (at the IBW PAN Coastal Research Station, Lubiatowo). Some archival data have been used as supporting research material. The field surveys of the dynamic layer were conducted in the southern Baltic coastal zone with the use of the StrataBox (SyQwest Inc. USA). Additional measurements for testing the equipment and improving the interpretation of the recorded signals were carried out in the Vistula Lagoon.

2. Variability of dynamic layer features: the Polish experience

As mentioned above, the notion of a dynamic layer exists in a number of disciplines, e.g. in coastal engineering, oceanography and geology. According to coastal engineers (see Mielczarski 2006), the dynamic layer in a non-tidal sea is defined as a layer of nearshore sediments spreading seawards to the depth where the sea bottom is affected by extreme waves and currents. For geologists (see Subotowicz 1996), the dynamic layer is a 'temporary layer, predominantly sandy, deposited on older formations as a result of the action of waves and currents'. In both of the above definitions, the driving forces of sea bed dynamics (waves and currents) play an important role. The influence of these hydrodynamic factors, through the mechanism of bed shear stresses, set the grains of seabed sediments in motion, thereby displacing them, resulting in the evolution of the seabed and the sea shore. Two questions arise: 1) To what extent and at what spatio-temporal scales are the dynamic layer parameters formed by coastal hydrodynamic and lithodynamic processes? 2) How do the sandy sediment resources accumulated in the dynamic layer (and the distribution of the sediment volumes on the cross-shore profile) influence actual sediment transport rates, the local sediment budget and sea bed changes?

Part of the answer to the first question can be found in the numerous results of experimental and theoretical investigations of coastal evolution, see e.g. Ostrowski (2004) and Pruszak (1998). The theoretical description and mathematical modelling of the sea shore and sea bed dynamics at various spatio-temporal scales, validated by laboratory and field experiments, are assumed to be research and engineering tools ensuring a satisfactorily accurate solution to most coastal morphodynamic problems, especially those related to the prediction of coastal erosion processes. The theoretical cross-shore profile and shoreline evolution models assume, however, that the nearshore resources of sandy sediments are unlimited, which is not always true. In many seas around the world, there is little or no sand on coastal sea beds (on rocky shores, for example). In such cases, the computational results may become more reliable if the modeller imposes a local apparent strengthening of coastal elements. For instance, in a one-line model, based on long-term (e.g. annual) sediment transport calculations[4], certain shore segments can be defined as unchangeable, i.e. built over by man-made coastal structures or resistant to erosion because of their geological composition.

The answer to the second question (strictly related to the first one) is not so easy to find. Although the dynamic layer is governed by coastal waves and currents, it is not completely understood how the sediment transport rate depends on geological factors, i.e. on parameters of the dynamic layer such as its local thickness. Sediment concentrations in the water column high above the sea bed even in storm conditions are very small, having values not exceeding a few grams per litre[5] (see Kaczmarek 1999). The concentration of sand grains is larger in the nearbed suspension layer (the so-called transitional or contact load layer) and in the bedload layer (the moveable sea bed layer), but the theoretically estimated total thickness of the contact load and bedload layers is no more than 2–3 cm (see Kaczmarek 1999). The results of field surveys carried out using radio-isotope tracers by Pruszak & Zeidler (1995) indicate that the thickness of the nearbed moveable sediments under extreme storm conditions is equal to $A_d = 4$–6 cm. Such quantities, very close to the sheet flow layer thickness reported by Myrhaug & Holmedal (2007), have been observed for a breaking wave height $H_b \approx 0.8$–1.2 m (at water depth $h \approx 1.5$–2.0 m), which yields the parameter k equal to about 0.05. This value, obtained for the non-tidal southern Baltic coastal zone, is slightly bigger than its counterpart obtained for a tidal oceanic coast (0.027) by Kraus (1985) and Sunamura & Kraus (1985).

[4]The one-line model simulates shoreline displacement as a function of the spatial variability of the net longshore sediment transport rates.

[5]Sediment concentrations may be greater at wave breaking locations.

The sheet flow layer thickness is sometimes wrongly considered to be equivalent to the mixing layer thickness. At the time scale of a storm, the mixing layer is many times thicker than the sheet flow layer observed instantaneously at any moment during the storm. In this context, the mixing layer can be equated with the dynamic layer representative of the individual storm. A one-storm dynamic layer can have a significant thickness, reflecting the entire history of sea bed change occurring during the storm. This history can encompass a distinct sea bed evolution, including migration of underwater bars. It is worth noting that as a consequence, local sediment transport rates depend on the shape of the sea bottom, which is the upper limit of the dynamic layer.

In view of the above findings, one can imagine that relatively small sediment resources in the dynamic layer can 'saturate' the water flow with sand grains in a short time scale (a matter of minutes). It is doubtful, however, whether the small sediment resources in the dynamic layer can feed the water flow satisfactorily and maintain the sandy 'saturation' for a longer time, exceeding the wave period, i.e. at scales of minutes, hours and days. Further, one may ask what influence local sand resources exert on coastal evolution along adjacent shore sections in the long term – over months and years.

As already mentioned, the dynamic layer's parameters are governed by the coupled impact of waves and currents, causing sediment motion in the coastal zone. In non-tidal seas, including the Baltic, the most spectacular geomorphologic effects are related to longshore sediment transport. This is so intensive that, according to some researchers (see e.g. Pruszak 2003), it gives rise to the longshore movement of sand with a net rate of more than $100\,000$ m^3 year^{-1}. It is assumed in theoretical calculations that the amount of sediment set in motion depends only on hydrodynamic forcing and sea bed grain diameters. The analysis of Racinowski & Baraniecki (1990) shows, however, that computationally obtained longshore sediment transport rates reflect only longshore transport ability and should be interpreted as the 'maximum mass or volume of sand that can be displaced along the shore in given coastal hydrodynamic conditions'. It has also been pointed out by Mielczarski (2006) that the longshore sediment transport rate, determined conventionally on the basis of the longshore component of wave energy, is actually the 'transport ability of wave motion', the real usefulness of which depends on the amount of sandy sediments accumulated in the nearshore dynamic layer.

The southern Baltic coast is dominated by beaches and dunes: consisting mostly of Holocene sands, they make up about 80% of the Polish shoreline. Locally, there is also peat and mud on the sandy shores, usually in the form of interbeddings under the beach or dune surface. Cliff shores, making up the remaining 20% of the Polish coast, are basically built of Pleistocene formations, mainly till and silt, but also sand, gravel and pebbles. Small amounts of Holocene sands can be found at the toes of the cliffs (see Figure 2).

The Polish shore in the eastern part of the Gulf of Gdańsk, from the Polish–Russian border to the Vistula river mouth, is an example of a beach-dune coast. Here, stable accumulative shores, with wide beaches and high dunes, are predominant. In particular, intensive sand accumulation has taken place on the shores of the Vistula Spit. This is most probably due to the convergence of two oppositely directed longshore sediment fluxes. Recently, joint Polish–Russian investigations have been carried out with the aim of identifying this convergence region. Extensive studies have shown that the convergence point for the hydrodynamic conditions of the mean statistical year is located near the base of the Vistula Spit[6]. It is worth noting that an artificial channel across the Vistula Spit is planned at the nearby village of Skowronki 3 km to the east (KM 23.3) (see Figure 1 for the location of this study site). A simplified geological transect of the coastal zone at Skowronki is shown in Figure 3.

Figure 3 implies the existence of very large amounts of sandy (Holocene) sediments accumulated in the coastal zone. The nearshore sandy layer (1) is ca 20 m thick and extends a long way offshore. It is worth noting that to some extent the Pleistocene substratum also consists of sandy sediments. These sediments and the Holocene sands may well be of similar grain sizes. Therefore, one should be aware of the fact that the results of any seismo-acoustic measurements for determining the thickness of the Holocene layer may be ambiguous. As pointed out by Frankowski et al. (2009), difficulties in the interpretation of seismo-acoustic field data, despite ongoing significant progress in surveying techniques and devices, incline (or rather force) geologists and engineers to apply also other, more direct, investigative methods, e.g. the collection and analysis of sediment core samples. This issue will be discussed in the next section of the paper.

The shores of the western part of the Gulf of Gdańsk are at most accumulative, with huge amounts of quartz sand in layers of a considerable thickness. At Sopot, for instance (see Mojski 1979), drillings carried

[6]Wind and wave conditions vary significantly from one year to another and the longshore sediment transport convergence point migrates within a large area along almost the entire Vistula Spit coast. It should be mentioned that the Vistula Spit has been nourished naturally, mostly by sediments derived from eroded shores in the north-eastern part of the Gulf of Gdańsk (Sambian Peninsula). The opposite longshore sediment flux, moving from the Vistula mouth region towards the Vistula Spit, is much less intensive.

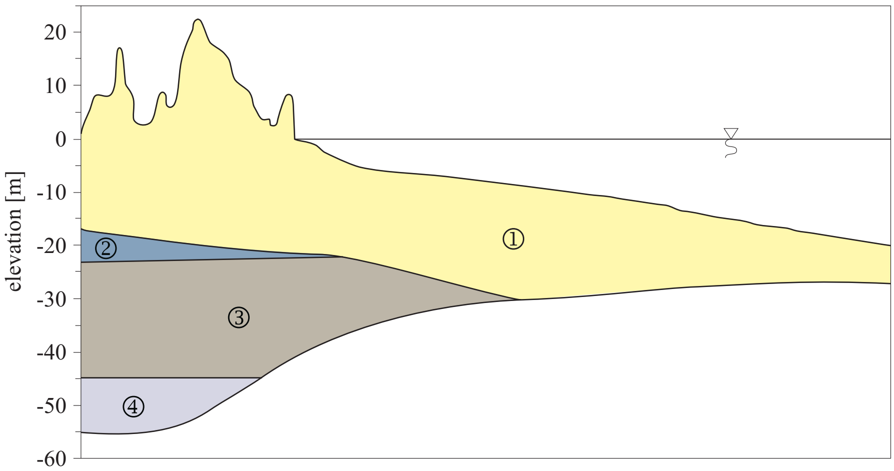

Figure 3. Geological cross-section across the Vistula Spit coast (KM 23.3), after Frankowski et al. (2009): 1 – aeolian and marine sand, 2 – clay, silt and sand of limnic and fluvial origin, swampy sediments (peat, gyttja and mud), 3 – sand and sand with gravel (fluvioglacial, fluvial and marine), 4 – clay, silt and sand of ice-dammed marginal lakes

out near the beach by the Polish Geological Institute revealed a 40 m thick surface layer of medium-grained sand with small amounts of silt admixtures. These sediments are Quaternary deposits, overlying older (Neogene) formations of various types (loam, sand). We have not managed to unearth any data which could distinguish the Pleistocene and Holocene layers in these Quaternary sands. This is a further argument warning of the ambiguity of geological survey results, possibly to be used in practical coastal engineering applications, and of the uncertainty of conclusions drawn from them.

The coast at Lubiatowo is a characteristic segment of the 'open' sea shore (see Figure 1 for its location), with a significant area of the coastal zone covered by aeolian deposits (beaches and dunes). According to Uścinowicz et al. (2007), beach-type and spit-type sands are found on the emerged part of the shore, the thickness of this layer being 3–5 m. On the shoreface, these sands extend back some 70–80 m from the shoreline, where they overlap marine sediments. The latter are represented mainly by fine sands in a layer whose thickness varies over a wide range from 1 to 7 m. Holocene sediments of various origin – fine sand with some organic matter (e.g. peat) – lie beneath the beach and dunes, down to 7–8 m below the mean sea level. The sediments underlying these consist mostly of Pleistocene glacial sand and gravel, as well as till. A simplified geological cross-section of the coastal zone at Lubiatowo is shown in Figure 4.

The vertical lines A–E in Figure 4 indicate the locations and depths of drillings. It should be assumed that the layers shown in Figure 4 are absolutely true only at these locations, whereas the remainder of the cross-section represents a hypothetical system of sediment layers. Most probably, seismo-acoustic methods were applied, particularly where the water was deeper (more than 5–6 m)[7]. The features of the sediment layers shown in Figure 4 demonstrate the existence of a boundary between the non-cohesive Holocene and Pleistocene sediments. This boundary may remain undetected in seismo-acoustic measurements (a separating layer of organic-bearing material has been found in drill cores on land only). It is extremely doubtful whether the notion of the coastal dynamic layer makes sense in the case of the geological cross-section shown in Figure 4 (as in the layout shown in Figure 3).

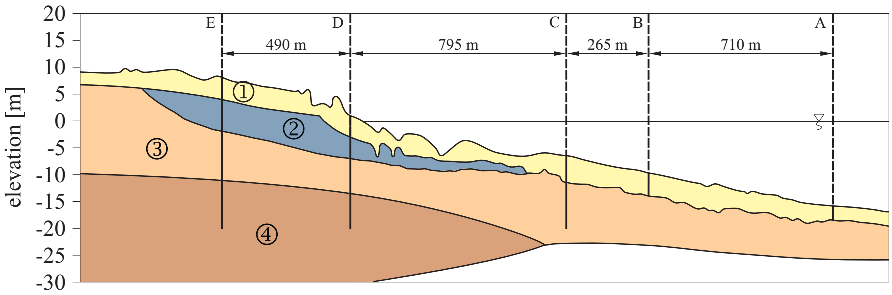

Figure 4. Geological cross-section of the coast at Lubiatowo at KM 163.73, after data from Uścinowicz et al. (2007): 1 – aeolian and marine sand, 2 – organic-bearing sediments, 3 – glacial sand and gravel, 4 – till

Long-term surveys of morphodynamic processes on the multi-bar dissipative shore near Lubiatowo show that the characteristics of sea bed deposits are subject to changes in time and space, both in the cross-shore and the longshore directions. These changes are caused by large-scale coastal evolution resulting from the motion of huge volumes of sandy material, visible as moving bars and the quasi-periodically varying positions of the bars.

[7]In the shallow-water nearshore zone, hydro-acoustic and seismic surveys encounter numerous problems. Here, only boats with a small draught can be used. Such a boat is sensitive to wave motion, which hinders measurements because of its instability (heaving, rolling, pitching etc.). Although advanced boat motion compensating systems have recently become available, nearshore investigations are really only possible in calm conditions, and such conditions are rare. Thus the opportunities to collect data close to the shoreline are limited.

3. Field measurements

The most reliable data on the geological structure of the coastal zone are provided by analysis of core samples taken from the sea bed. Although the accuracy of a geological cross-section depends on the number of drillings, even a large number of drill cores do not provide complete information on spatial changes in the sediment layers. Geophysical surveys providing a continuous record of both sea bed surface and sub-bottom layers are essential. Such measurements are possible owing to the specific properties of the aquatic environment, such as good propagation of mechanical waves – ultrasounds and seismo-acoustic signals. Ultrasonic methods are applied in investigations of the sea bed surface shape, whereas seismo-acoustic methods are used to survey the sea bed substratum layers.

Seismo-acoustic methods are based on the emission of a sound signal and analysis of the echo reflected from the individual layers making up the sea bed. Interpretation of seismo-acoustic measurements involves determining the reflection limits in the records, distinguishing uniform acoustic units and relating these to geological (litho-genetic) classifications. The aim of preliminary analysis of seismo-acoustic material is to select places where core samples of sediment ought to be collected. According to recommendations in Frankowski et al. (2009), the ultimate interpretation of seismo-acoustic data, leading to their conversion into geological cross-sections, should be preceded by drillings and analysis of the drill core samples, as well as verification of the findings of geophysical surveys other than acoustic measurements. During the interpretation and processing of the seismo-acoustic data, geological-engineering cross-sections are drawn showing the boundaries between the sediments and the thicknesses of the individual layers.

Devices used in seismo-acoustic surveys, known as sub-bottom profiling devices, are constructed in the same way as bathymetric echo-sounders, but they work at lower frequencies, most often not higher than a dozen or so kHz. They also have a higher emitted signal energy in comparison to hydrographic and navigable echo sounders.

Geophysical vessels have their seismo-acoustic equipment incorporated permanently in the hull. Smaller craft use towed or side-mounted submerged devices. Because these consume a relatively large amount of power, the supply to the sub-bottom profilers requires 230 V wiring, which is available on bigger vessels only. The StrataBox, produced by SyQwest Inc. (USA), is one of the few devices powered by 10–30 V DC. Having been purchased recently by the Institute of Hydro-Engineering of the Polish Academy of Sciences (IBW PAN), this equipment works with an acoustic frequency of 10 kHz and ensures penetration down to 40 m below the bottom for a sea bed built of cohesive deposits. For sandy sediments, the penetration range is no more than a few metres, but the transducer is light enough for it to be mounted on the side of a small boat. The power supply is 12 V or 24 V (DC). According to the specification sheet, the StrataBox can operate at maximum depth of 150 m; the minimum depth depends on the type of sediment on the sea bed surface. In addition, the user manual recommends that the distance between the transducer (its lower submerged surface) and the sea bottom should not exceed 2.5 m. The surveys described in the present study have proved this minimum distance to be slightly smaller, namely 1.8–2.0 m.

Measurements carried out in May 2009 near the IBW PAN Coastal Research Station (CRS) at Lubiatowo focused on surveying the structure of the non-cohesive sea bottom. It was known from analysis of surficial sea bed samples taken previously at Lubiatowo that the sea bottom consists mostly of fine sand with a median grain diameter of d_{50} = 0.20–0.25 mm; locally it is coarser – $d_{50} \approx 0.4$ mm. The objective of the seismo-acoustic survey using the StrataBox was to determine the thickness and offshore range of the dynamic layer as conventionally defined (Boldyrev 1991, Subotowicz 1996). When the measurements were planned and carried out, we were not in possession of the geological data in Figure 4; we had expected to identify the location of the seaward edge of the sandy (Holocene) sediments lying on a native (Pleistocene) substratum of different composition.

The measurements near CRS Lubiatowo were carried out using a motor boat with a length of 5 m and a draught of 0.3 m. The boat's position was determined using GPS Magellan. The StrataBox signals were recorded by the application of software StrataBox ver. 3.0.6.2, enabling simultaneous registration of the seismo-acoustic data and the geographical coordinates of the points surveyed. Figure 5 shows a photograph of the boat and the StrataBox transducer (before being lowered into water).

During the two-day long survey (19–20 May 2009) tens of files with seismo-acoustic signals were recorded. The aim of these measurements was to test the equipment and tune parameters (e.g. setting the optimal signal gain). The actual profiling survey was carried out on 20 May, in a direction approximately perpendicular to the shoreline, from the depth of about 13 m (starting point of the profile – $54°49.561'$N, $17°49.823'$E) to the nearshore shallow water region (end of the profile – $54°48.867'$N, $17°50.322'$E). The measured bathymetric cross-shore profile was found to have the same shape as the sea bottom transect shown in Figure 4. In the area where bars occur (at depths less than 8 m), where considerable changes in the sea bed take place not just at the scale of years but at the scales of months and weeks, the measured depths were slightly different than the ones in Figure 4. The

Figure 5. Seismo-acoustic measurements using the side-mounted StrataBox (the transducer pole, attached to the boat's side via an articulated joint, before being lowered to water) near CRS Lubiatowo in May 2009

maximum discrepancies between the sea bottom ordinates measured in May 2009 and those plotted in Figure 4 are 2 m.

The results at long distances from the shoreline, at water depths exceeding 10 m, indicated the presence of homogeneous sandy sediments in the sea bed. More interesting results were found closer to the shoreline. Excerpts of the StrataBox seismo-acoustic record of the surveyed profile are shown in Figures 6–8.

The record at 9 m depth (Figure 6) shows the boundary between two types of sediments. The data from drill core B (cf. Figure 4) suggest that the device has detected a local structure of the sea bed, consisting of a 3 m thick layer of marine sands above glacial sands. The measurements carried out in the vicinity of the gently-sloping outer bar at a distance of about 750 m from the shoreline (Figure 7) reveal the presence of weakly shaped boundaries between sands of various kinds and various origin. The echo reflected from the boundary at the −11.0 m ordinate may imply the existence of a distinct interface between the marine and glacial sands (see the drill core C in Figure 4).

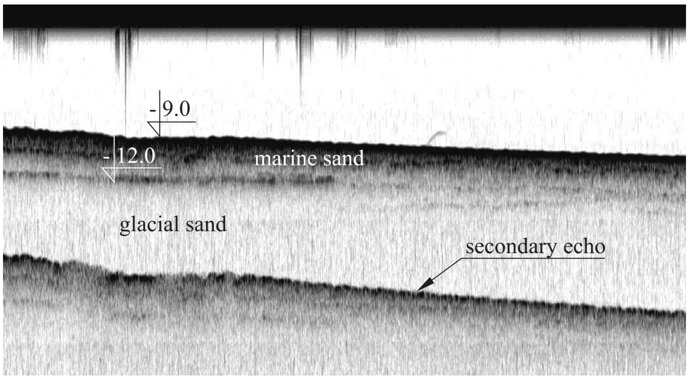

Figure 6. Sub-bottom profile surveyed with the StrataBox at Lubiatowo at a depth of ca 9 m (about 1000 m from the shoreline), near borehole B of Figure 4

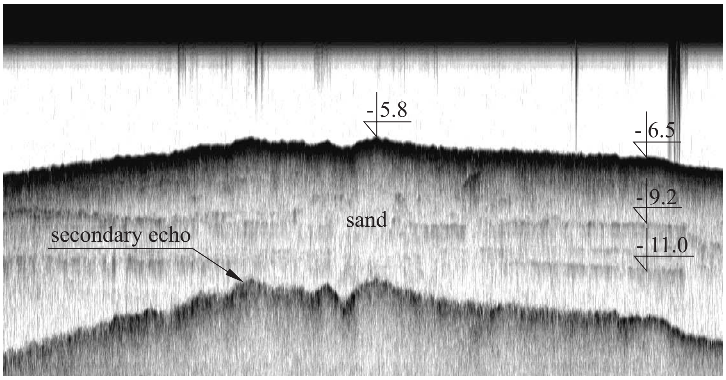

Figure 7. Sub-bottom profile surveyed with the StrataBox at Lubiatowo at a depth of ca 6 m (about 750 m from the shoreline), near borehole C of Figure 4

The profiling survey carried out in a deep trough between the bars located about 300 m from the shoreline (Figure 8) revealed layers which, on the basis of the data of Figure 4, may correspond to organic-bearing sediments (peat, sandy peat, mud, etc.). Measured in May 2009, the depth of water in the middle of the above-mentioned trough between the bars was 8.8 m, while the maximum depth in this region on the strength of Figure 4 was equal[8] to about 6 m. Moreover, Figure 4 shows the superficial layer of sand on the sea bed with a thickness of 1.5 m, overlying organic-bearing sediments. One can thus assume that erosion of the sea bed sandy layer has taken place at this site, thereby exposing the organic-bearing sediments.

[8]The field surveys reported by Uścinowicz et al. (2007) were carried out in the 1990s.

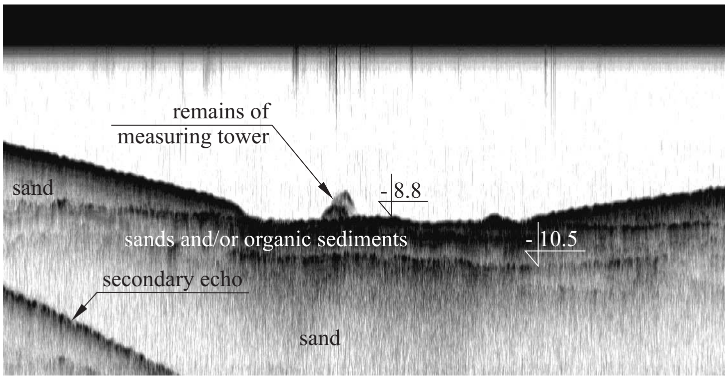

Figure 8. Sub-bottom profile surveyed with the StrataBox at Lubiatowo at a bar trough (about 300 m from the shoreline), between boreholes C and D of Figure 4; the distinct protrusion on the seabed represents the remains of the old measuring infrastructure at CRS Lubiatowo

However, because of the relatively small thickness of the organic-bearing layer (ca 1.5 m according to Figure 4), this material could also have been washed away, exposing the glacial sand located beneath.

In order to clarify the above doubts, the StrataBox device was tested under quite different conditions, namely in the Vistula Lagoon, the bottom of which consists mostly of muddy sediments. Carried out in August 2009, the measurements encompassed a few sites located in the south-western part of the Vistula Lagoon (see Figure 1). Part of the sub-bottom profile corresponding to the point with the coordinates $54°20.692'$N, $19°17.220'$E is presented by way of example in Figure 9.

The results of drillings commissioned by IBW PAN in autumn 2007 revealed the following layers of sediments at this site (from the surface downwards): highly plastic silty mud (thickness 1.2 m), highly plastic mud (thickness 1.8 m) and fine sand. The ordinates given in Figure 9 indicate that the attempt to interpret the seismo-acoustic signals did not fully correspond to the drill core data. The most important finding, however, is related to the picture of superficial muddy layers, visible in Figure 9, which differs considerably from the picture of sand, visible in both Figure 9 (the deeper sub-bottom layer in the Vistula Lagoon) and in Figures 6–8 (the sea bed at Lubiatowo). Thus, it can be concluded that the sea bed sediment limits in Figure 8 are the intersections between layers of various sandy sediments.

Nothing like the floor of the classically defined dynamic layer was detected in the seismo-acoustic data from Lubiatowo presented here, which implies that there are very large resources of sandy sediments on this shore

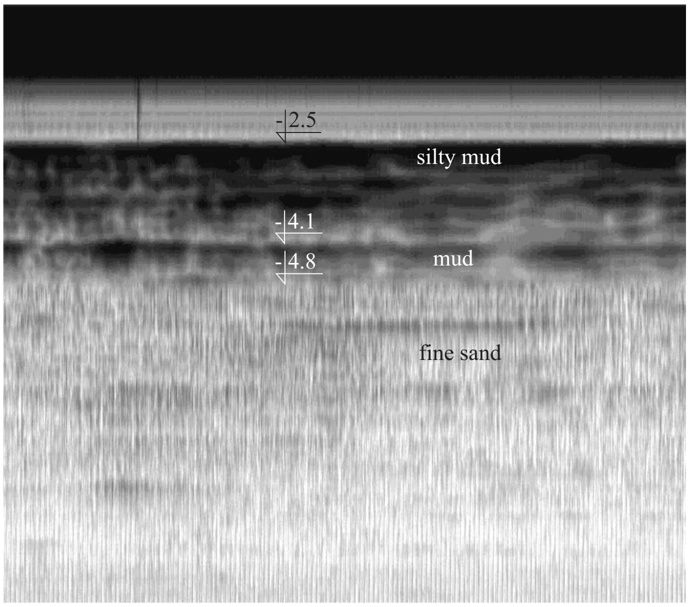

Figure 9. Sub-bottom profile surveyed with the StrataBox at a depth of ca 2.5 m in the north-western part of the Vistula Lagoon (54°20.692′N, 19°17.220′E)

segment. According to the typology proposed by Boldyrev (1991), the shore near Lubiatowo is accumulative.

4. Discussion and conclusions

The significance of the dynamic layer to the motion of water and sediment caused by waves and nearshore currents depends on the amount of sand in the coastal zone. Here, the geological origin of the sandy sediments is not important. The traditional notion of the dynamic layer is associated with a layer non-cohesive Holocene sediments overlying a Pleistocene substratum, on condition that this substratum is built of cohesive deposits, e.g. clay or silt. As pointed out by Subotowicz (2005), the geological cross-section of a dune-type seashore bears a slight resemblance to a cliff seashore. This likeness lies in the Holocene marine sand deposited at the toe of a dune or cliff. According to the classification of Subotowicz (2005), the emerged part of the cross-shore profile is formed of either Holocene aeolian sand (dune) or Pleistocene cohesive deposits (cliff). Subotowicz (2005) distinguishes concave and convex geodynamic shoreface classes for both dune-type and cliff shores. The concave shoreface has a dynamic layer with a large amount of sandy sediment (vulnerable to erosion), whereas the convex shoreface is characterized by a small amount of sand deposited on a Pleistocene substratum (resistant to erosion).

Simultaneous sub-bottom profiling and hydro-acoustic surveys in the multi-bar coastal zone of Lubiatowo, highly representative of the southern Baltic coast, reveal a correlation between the sediment resources (the dynamic layer thickness) and the existence of large sea bed forms. The presence of a distinct thick and permanent layer of sandy sediments is accompanied by a large number (4–5) of underwater bars that are stable even at very long (multi-year) time scales. Thus, the existence and condition of the bars can be assumed to be a visual indicator of the 'rich' dynamic layer. The stability of the shoreline position at various time scales is an additional indicator of dynamic layer permanence.

The mixing layer thickness A_b on the multi-bar dissipative shore at Lubiatowo yields the parameter k equal to about 0.05. This value lies relatively close to the results presented by Kraus (1985) and Sunamura & Kraus (1985), namely $k = 0.027$, obtained for the Pacific coast, which is characterized by different hydrodynamics and cross-shore profile shape.

The Polish coast consists predominantly of dune-type seashores where, in view of the available data (see e.g. the cross-sections in Frankowski et al. 2009), the Holocene aeolian and marine sand is most often deposited on the Pleistocene glacial sand. From the point of view represented by investigators of coastal hydrodynamic and lithodynamic processes, the classical definition of the dynamic layer has no sense in such conditions because the features of the superficial sea bed layer are very similar to the features of older sediments which lie beneath. Theoretical and experimental (laboratory and field) studies carried out to date show that two kinds of sand with rather similar grain sizes are almost equally vulnerable to erosion and subject to sedimentation in the same conditions. Only when significant differences in grain sizes appear (e.g. the median grain diameters d_{50} vary by an order of 0.1–0.2 mm) do the sediments behave quite differently under the same hydrodynamic impact. Therefore, in investigations of nearshore sediment motion and the evolution of most stretches of coastline in Poland, one can forget about limitations of sediment supply alleged to be due to the small thickness of the Holocene sediments.

The opposite situation holds true in the case of cliff shores. On most cliff shore segments in Poland, the deficiency of Holocene sediments just means a deficiency of sand. In research investigating cliff shores, models of sediment transport and shore evolution ought to be adapted to take account of the limited supply of sandy sediments. Hence, these models, hitherto capable of calculating the theoretical maximum sediment transport over the entire cross-shore profile, should be adapted to the actual conditions in which the dynamic layer does not extend far offshore. The computations carried out for real conditions of sediment supply on Polish cliff shores can be used to verify the state of the art with regard, for example, to net sediment transport rates along individual stretches of the Polish coast. Besides, as stated in the introduction, such computations would be helpful in the optimization of the anti-erosion protection of the Polish coast, the individual sections of which require different methods of protection owing to the spatially different parameters of the dynamic layer.

References

Boldyrev W. L., 1991, *Morphological and lithologic indicators of development regions of longshore sandy sediment fluxes*, Inż. Mor. Geotech., 2, 50–53, (in Polish).

Ciavola P., Tarborda R., Ferreira O., Dias J., 1997, *Field observations of sand-mixing depths on steep beaches*, Mar. Geol., 141 (1–4), 147–156.

Frankowski Z., Graniczny M., Juszkiewicz-Bednarczyk B., Kramarska R., Pruszak Z., Przezdziecki P., Szmytkiewicz M., Werno M., Zachowicz J., 2009, *Principles of documentation of geological-engineering foundation conditions for marine structures and coastal protection measures*, Polish Geol. Inst., Warsaw, 201 pp., (in Polish).

Kaczmarek L. M., 1999, *Moveable sea bed boundary layer and mechanics of sediment transport*, IBW PAN, Gdańsk, 209 pp.

Kraus N. C., 1985, *Field experiments on vertical mixing of sand in the surf zone*, J. Sediment. Petrol., 55 (1), 3–14.

Mielczarski A., 2006, *Remarks on notional models of sediment motion in the underwater part of the coastal zone*, Inż. Mor. Geotech., 1, 13–22, (in Polish).

Mojski J. E., 1979, *Detailed geological map of Poland on 1:50 000 scale*, Sheet – Gdańsk, Wyd. Geol., Warsaw, (in Polish).

Myrhaug D., Holmedal L. E., 2007, *Mobile layer thickness in sheet flow beneath random waves*, Coast. Eng., 54 (8), 577–585.

Ostrowski R., 2004, *Morphodynamics of a multi-bar coastal zone*, IBW PAN, Gdańsk, 163 pp.

Pruszak Z., 1998, *Dynamics of beach and seabed*, IBW PAN, Gdańsk, 463 pp., (in Polish).

Pruszak Z., 2003, *Marine basins. Outline of physical processes and environmental engineering*, IBW PAN, Gdańsk, 272 pp., (in Polish).

Pruszak Z., Różyński G., Szmytkiewicz M., Skaja M., 1999, *Quasi seasonal morphological shore evolution response to variable wave climate*, Proc. 4th Int. Symp. Coast. Eng. Sci. Coast. Sediment, ASCE, 1081–1093.

Pruszak Z., Zeidler R. B., 1995, *Sediment transport in various time scales*, Proc. 24th Int. Conf. Coast. Eng., Vol. 1, ASCE, 2513–2526.

Racinowski R., Baraniecki J., 1990, *Usefullness of lithologic indicators in description of the longshore sediment flux on the Polish Baltic shore*, Rozpr. Hydrotech., 51, 159–210, (in Polish).

Saini S., Jackson N., Nordstrom H., 2009, *Depth of activation on a mixed sediment beach*, Coast. Eng., 56 (7), 788–791.

Sherman D., Nordstrom K. F., Jackson N., Allen J., 1994, *Sediment mixing depth on low energy reflective beach*, J. Coastal Res., 10, 297–305.

Subotowicz W., 1996, *A need of new look at classification of the coastal zone*, Inż. Mor. Geotech., 5, 332–335, (in Polish).

Subotowicz W., 2005, *Lithodynamic determinants of shore protection in Poland*, Inż. Mor. Geotech., 3, 190–194.

Sunamura T., Kraus N. C., 1985, *Prediction of wave average mixing depth in the surf zone*, Mar. Geol., 62 (1), 1–12.

Uścinowicz Sz., Zachowicz J., Przezdziecki P., Zaleszkiewicz L., 2007, *Geodynamic map of the Polish south Baltic coastal zone*, Sheet – Lubiatowo, No. 34, Marine Geol. Dept., Polish Geol. Inst., Gdańsk, (in Polish).