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The physical oceanography of the Gulf of Finland: a review

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The Gulf of Finland is today an actively-investigated sea area. Basin-wide studies are carried out in an international cooperation between all the coastal states of the gulf. Understanding of the basic physics of the gulf is vital in order to assess the state of the marine environment and to construct coupled hydrodynamic-ecological models for describing the response of the sea to human activities. This paper is a literature review of the physical features of the Gulf of Finland. Our main interest is the general circulation as well as the horizontal and vertical structure of water masses, but surface waves, the sea level, ice conditions and air-sea interaction are also briefly reviewed. Special attention has been paid to the eastern end of the gulf, which is an important mixing area and buffer zone for pollutants.

Introduction

The Gulf of Finland (hereafter denoted the GoF) is an elongated estuarine sea in which physical phenomena ranging from small-scale vortices up to large-scale circulation take place. It is a complicated hydrographic region, with saline water input from the Baltic Sea Proper in the west and with a large fresh water input from rivers mainly in the east.

The eastern GoF is an area of special interest. It receives the largest single fresh water input (River Neva) of the whole Baltic Sea (15% of the total Baltic river inflow), including the most severe loading. Recently, the Neva Bight has become an even more specific area due to the construction of a flood protection barrier (the "dam" or "damba") separating this basin to a large measure from the gulf. Moreover, a number of other coastal sub-basins exist in the eastern GoF (the Bay of Viborg, the Luga Bight, the Koporye Bight) with their special oceanographic features.

Scientific studies in the GoF began in the late 1890s by Admiral S. Makarov and by Professor Th. Homén (Witting 1910). Some extensive investigations were made as early as the first years of the 20th century. As a result of this, Witting (1909, 1910) described the general circulation and stratification conditions of the GoF. Up to the Second World War the GoF could be investigated from its southern to its northern coast (Palmén and Laurila 1938); after that, no joint research covering the whole cross-section of the GoF was possible until the 1990s. For several decades field experiments were mainly local, but now the physical oceanography of the gulf can once again be studied as an entity.

The limited possibilities for conducting joint large-scale research in the GoF over several decades mean that many important processes are still quite inadequately understood. However, the geographically limited size allows now the area to be covered by observations with a rather dense spatial resolution. Descriptions of some of the general features of the physics of the GoF can be found from several textbooks, where, however, the main focus has been on studies of the Baltic Sea Proper. The reader is referred to reviews of the Baltic Sea marine systems by Magaard and Rheinheimer (1974), Voipio (1981), Mälkki and Tamsalu (1985) and Fonselius (1996).

Although the number of published works is great, most of them are devoted to either local problems or very specialised topics. Many of the publications fall into the "grey literature" category, which may be hard to find. The results from the eastern GoF are mainly published in the Russian language. The multi-scientific joint project "The Gulf of Finland Year 1996" between Estonia, Finland and Russia inspired us to look more thoroughly at the physics of the GoF. Our motivation has been to bring into the light of day as much literature as we could find, even though we have not summarised the results of all the publications. The limited scientific discussion on the physics of the GoF has led to a situation in which some problems have been studied in detail, while other topics have not been dealt with very much at all. This is also reflected in the balance between the different sections. We have limited our scope here mainly to the circulation physics, distributions of salinity and temperature and related processes.

Basin dimensions and water masses

Geographic shape and topography

The GoF is a gulf in the north-eastern Baltic Sea. It lies between 59°11´N, 22°50´E and 60°46´N, 30°20´E (Fig. 1). In contrast with the other subbasins of the Baltic Sea, the GoF is a large estuarine basin having no sill to the Baltic Sea Proper. The line between the Hanko peninsula and the island of Osmussaar is often treated as the western boundary of the GoF. This, however, is more a convention than a real physical boundary. The relevant parameter values for the GoF are given in Appendix 1. Its volume of 1 103 km^3, is about 5% of the volume of the whole Baltic Sea. The drainage area is 420 990 km^2, i.e. about 20% of the total drainage area of the Baltic Sea (*see* Falkenmark and Mikulski 1975, Astok and Mälkki 1988).

The central GoF is quite deep (over 60 m) up to longitude 28°E. The south-eastern part is somewhat shallower and the easternmost part of the GoF is very shallow. The southern coast of the western GoF is rather steep, whereas the northern coast is shallower and more broken, with small islands. Some peninsulas, like Hanko and Porkkala, steer the currents on the Finnish coast; currents are also steered by the island of Naissaari and the ridge between it and the Estonian coast. The large and wide eastern basin gets narrower and shallower east of Narva. It is broken by two peninsulas on its southern coastline. The transition zone between longitudes 28°E and 29°E with decreasing width and depth is sometimes called "the Seskar basin". It plays an essential role in the transportation of water and substances and, because of the intensification of deposition processes there act as a kind of buffer. The easternmost part of the GoF, the Neva Bight, 22 km long and 14–15 km wide, is a very shallow area. The mean depth there is only some 5 m. The topographic features of the GoF are rich and they play an important role in the modification of the circulation (*see* Circulation).

General hydrographic features

The hydrography of the GoF is typical of estuaries. It is characterised by large horizontal and vertical variations. The whole gulf is a transition zone from fresh water to the waters of the Baltic Sea Proper. This holds true for the deep waters of the GoF, too, because there is no sill between the GoF and the Baltic Sea Proper. Thus, there are no topographically-isolated water masses in the GoF (Fig. 2).

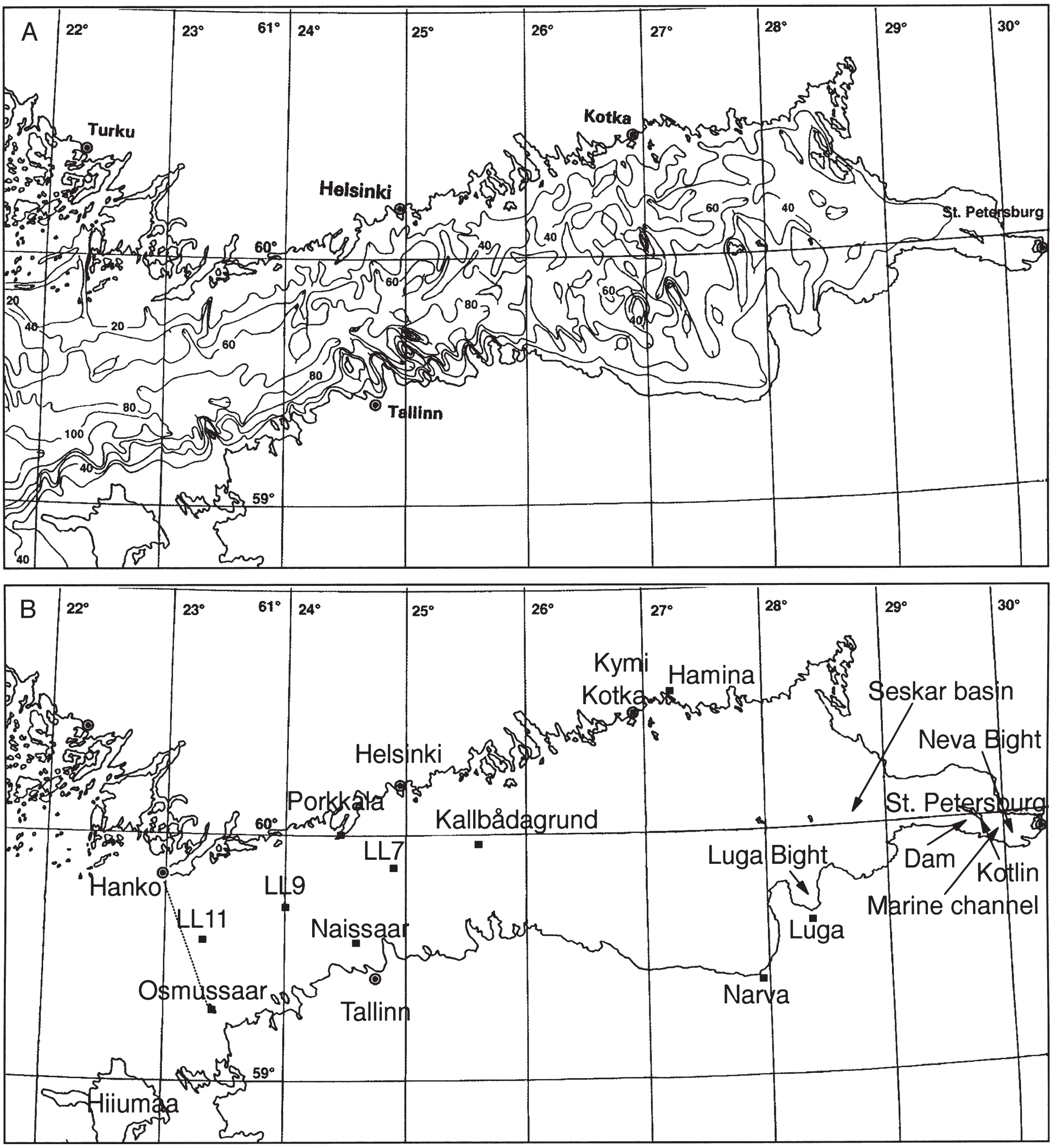

Fig. 1. — A: The bottom topography of the Gulf of Finland (from Fonselius 1996). — B: relevant geographic locations in the Gulf of Finland.

Most of the fresh water input enters the GoF in the easternmost part of the basin. The annual runoff of the river Neva varies considerably from year to year (data exist from the year 1859, *see* also Bergström and Carlsson 1993), ranging from 42 $km^3 y^{-1}$ (observed in 1940) to 115 $km^3 y^{-1}$ (in 1924). The mean value of the total annual river runoff into the GoF is 114 $km^3 y^{-1}$, which is about one quarter of the total fresh water input to the whole Baltic Sea. This shows how diverse the water masses of the GoF are. The river Neva has a mean discharge of 75.5 $km^3 y^{-1}$ (2 400 $m^3 s^{-1}$). The rivers Kymi and Narva, on either side of the eastern GoF, contribute to the fresh water supply with an annual mean runoff of 9.5–12.5 $km^3 y^{-1}$ (300–400 $m^3 s^{-1}$) each, and the river Luga with 3 $km^3 y^{-1}$ (100 $m^3 s^{-1}$) (Mikulski 1970).

The large fresh water input is an important

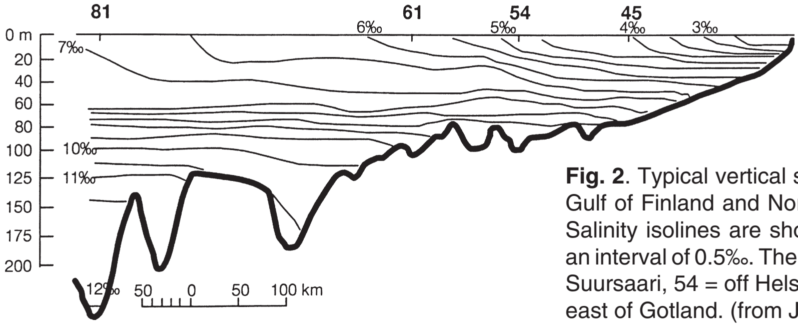

Fig. 2. Typical vertical section of salinity through the Gulf of Finland and Northern Baltic Sea in summer. Salinity isolines are shown as continuous lines with an interval of 0.5‰. The numbered locations are: 45 = Suursaari, 54 = off Helsinki, 61 = off Hanko and 81 = east of Gotland. (from Jurva 1952a).

factor causing water movements. It is of special importance in spring, when the rivers have their annual maximum runoff and, on the other hand, the winds and atmospheric pressure gradients are weak. At that time of the year the fresh water flows outwards from the GoF in the surface layer. In autumn, the fresh water input is smaller and strong south-westerly winds predominate. There is then on average an inflow of water along the southern coast and an outflow of fresher water along the northern coast. This holds true as a long-term average, not necessarily as an instantaneous situation (*see* Circulation). The salty water from the Baltic Sea Proper tends to penetrate into the GoF along the bottom. The approximate amount of water exchange across the line Hanko–Osmussaar is 600 km^3 out of the GoF and 480 km^3 into the GoF per year. The annual water exchange is thus more than half of the total water volume of the GoF (Witting 1910). Some more recent studies (Aitsam and Astok 1972, Astok and Mälkki 1988) have given smaller values, indicating that the renewal time of the GoF water is about three years. However, those authors remark that there is actually no sense in quantitatively estimating the water exchange between the GoF and the Baltic Sea Proper, because the GoF is not topographically isolated from the main Baltic Sea. The horizontal water exchange is therefore a continuous process at all depths.

The GoF, like the whole Baltic Sea, has a positive fresh water balance, determined mainly by the salty water input, by the river runoff and by the output of relatively fresh water from the GoF. Taking annual mean over the whole Baltic Sea area, the evaporation and precipitation were considered to balance each other (*see* e.g., Ehlin 1981). However, the balance was found to strongly depend on the place and time. In the north, like in the GoF, the precipitation exceeds evaporation on average (*see* Appendix 1), but with pronounced seasonal and interannual variations. According to the later studies, the annual mean of precipitation seems to be clearly larger than the evaporation (*see* HELCOM 1986, Omstedt *et al.* 1997), but still with pronounced seasonal and interannual variations.

In the eastern GoF, the discharge of the river Neva is greatest in May–July and has a minimum in winter. This variability is also seen in the annual changes in surface salinity. Launiainen and Koljonen (1981) found a clear correlation between the maximum river runoff and the minimum surface salinity in the GoF. There also exists a negative correlation between the river Neva discharge and deep water salinity at Utö (slightly west of the entrance of the GoF). Launiainen and Koljonen (1981) explained this by considering the dynamic balance of the water masses in the Baltic Sea; the isohalines and pycnoclines are inclined. Changes in the amount of fresh water runoff cause dynamic changes in the system. At a proper site, this may also be observed below the halocline as a negative correlation between river runoff and salinity.

The salinity has both horizontal and vertical gradients all along the gulf except in the Neva Bight. The salinity increases from east to west and from north to south. The surface salinity varies from 5‰–7‰ in the western GoF to about 0‰–3‰ in the east. The bottom salinity in the western GoF can typically reach values of 8‰–9‰. After the saline water input in January 1993, even higher values, up to nearly 10‰, were observed.

The long-term salinity variations in the Baltic Sea are connected to the changes in fresh water runoff and to the varying water exchange between the Baltic Sea and the North Sea via the Danish Sounds. This has also been reflected in the salin-

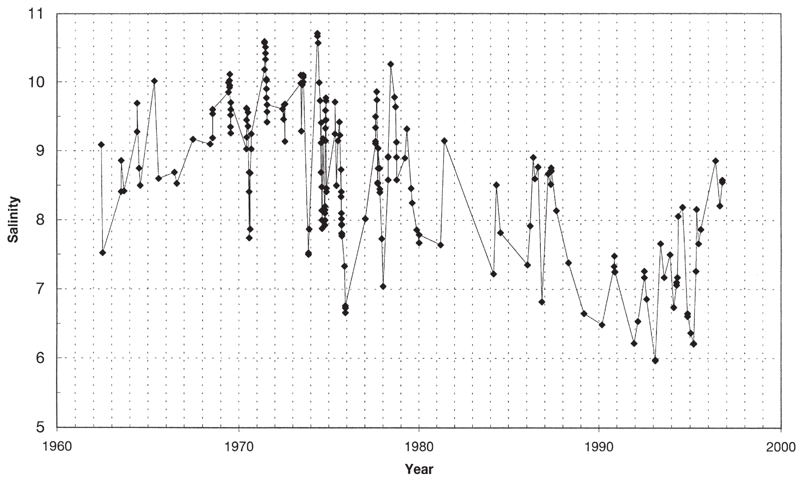

Fig. 3. Long-term variations in deep water salinity in the central Gulf of Finland (station LL7 between Helsinki and Tallinn).

ity of the GoF (Fig. 3). Several authors have concluded that the salinity increased in the Baltic Sea from the 1930s up to 1970s by about 0.5‰ (Hela 1966, Fonselius 1969, Perttilä *et al.* 1980, Pitkänen and Malin 1980, Launiainen and Koljonen 1981). After the major saline inflow in 1976, a 17-year-long stagnation period prevailed in the Baltic Sea, during which the salinity slowly decreased (Astok *et al.* 1990, Suursaar 1992, Alenius 1988). The next major saline water inflow took place in January 1993, interrupting the stagnation of the central Baltic deep water (*see* e.g., Matthäus and Lass 1995, Jakobsen 1995). The pulse increased salinity in the southern Baltic Proper, and an increase in the salinity of the GoF could be detected some 18 months after the pulse.

The salinity plays an important role in the buoyancy-driven baroclinic currents in the Baltic Sea in contrast to the oceans, where temperature differences make a larger contribution to buoyancy (*see* e.g., Mälkki and Tamsalu 1985). This is the reason why the horizontal and vertical distribution of salinity are of special interest in this paper.

The annual variations in the sea-surface temperature (SST) are large in the GoF. The maximum SST occurs in late July–early August, with the average maximum SST varying between 15 °C in the west to 17 °C in the east. The highest values are over 20 °C near the coasts. Discussions on the temperature distributions can be found in e.g., Witting (1936), Jurva (1952a), Mälkki and Tamsalu (1985) and Haapala and Alenius (1994) as well as in the forthcoming sections.

Other factors that affect the hydrography of the GoF are air-sea-interaction (wind stress, heat and vapour exchange), bottom topography, coastal effects and the Coriolis-effect. The role of these processes in the hydrography of the GoF will be discussed in more detail in the next sections.

Vertical stratification

Seasonal variations dominate the temperature variance in the GoF, as in the whole Baltic Sea. The time evolution of the stratification of temperature and salinity differ from each other clearly (Fig. 4). In winter, the upper layer down to the halocline at a depth of about 60 m is almost isothermal (Alenius and Leppäranta 1982, Haapala and Alenius 1994). The freezing point of the surface water varies, depending on the salinity, from –0.17 °C in the east to –0.33 °C in the west.

The surface temperature begins to rise in early April, when the ice has melted and the incoming solar radiation increases, changing the radiation balance from negative to positive at the sea-surface. In its first stage the heating proceeds faster in the easternmost areas of the GoF, and thus during the period May–June the temperature of the upper layer shows an increase from west to east. The rise of the surface temperature towards the maximum density temperature, i.e. about 2.5 °C in the western and 3.5 °C in the eastern GoF, makes the surface water heavier and causes the onset of vertical convection. When the surface temperature becomes higher than the maximum density temperature, the seasonal thermocline begins to form everywhere over the GoF, except in the Neva Bight, where mixing keeps the water

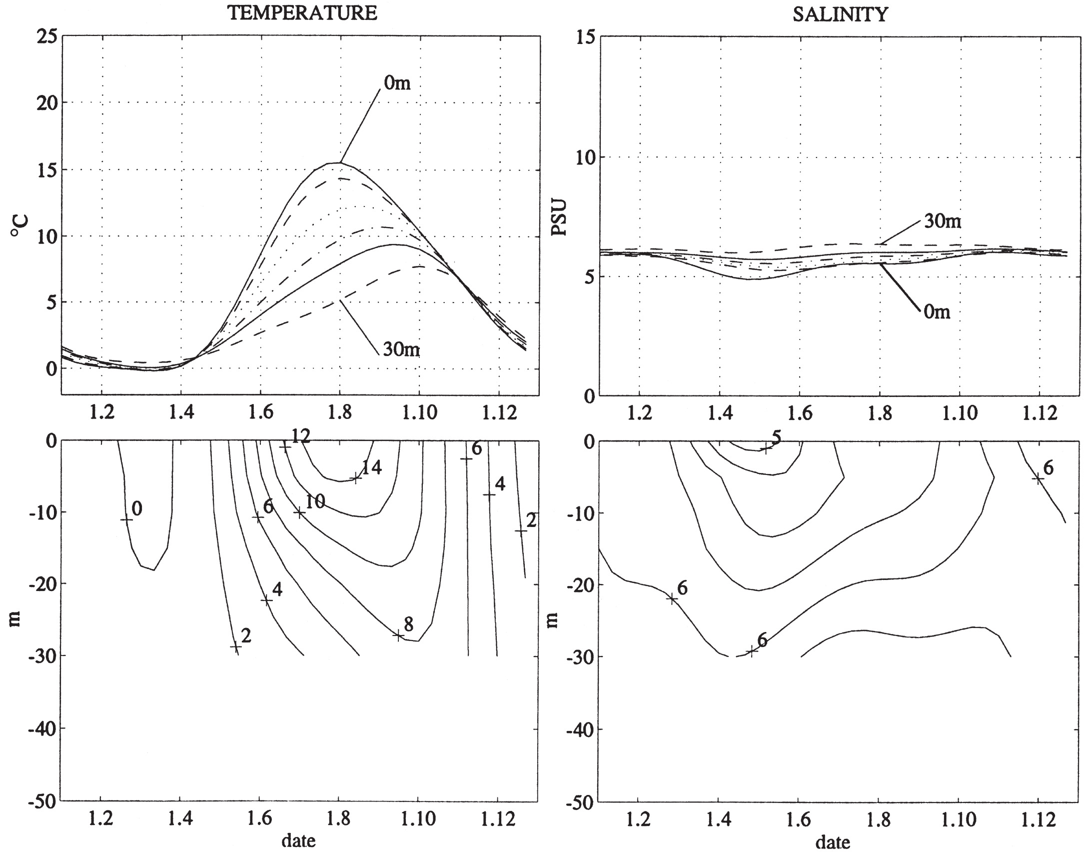

Fig. 4. Average seasonal variations (in years 1961–1990) of temperature and salinity at different depths at Harmaja (from Haapala and Alenius 1994). The values at different depths are shown as different line types (the depths are 0, 5, 10, 15, 20 and 30 m). The numbers with crosses in the isoline panels show the values of the isolines.

mass vertically homogeneous. In summer the upper mixed layer is thermally almost homogeneous because of continuous wind mixing; its thickness is then typically 10–20 m (Alenius and Leppäranta 1982). Underneath is the thermocline, in which the temperature drops from about 15–20 °C at the top to about 2–4 °C at the bottom of the thermocline. The pool of cold water below the thermocline is known as old winter water. The surface temperature usually reaches its maximum during late July–early August.

In late August, the radiation balance becomes negative. The sea-surface temperature drops, and vertical convection begins once again. Convection is dampened by the strong vertical salinity stratification. The well-mixed layer then deepens rapidly and in late October–early November the seasonal thermocline vanishes. The bottom temperature remains somewhat higher than that of the water masses above. Mechanical mixing due to stormy winds also plays an important role in the destruction of the thermocline. Convection proceeds until the temperature of maximum density is reached.

The temperature stratification is indirectly coupled with salinity conditions. In the eastern GoF and in the eastern part of the central GoF, there is no permanent halocline. The water body can therefore mix thermally during high wind speeds and/or cool periods during the summer. After such a situation the thermocline redevelops. This is typical for shallow coastal areas throughout the GoF.

The time evolution of temperature below the surface layer is more complicated than at the surface (Fig. 4) (Haapala and Alenius 1994). The annual date of maximum temperature varies linearly with depth only in the surface layer (0–30 m). In the mid-layer, between depths of 30 and 60 m, the date of the maximum temperature follows a logarithmic rate of change, but below a depth of 60 m the observations are very scattered, showing no systematic behaviour. The mean time lag between the date of maximum temperature at the surface and at a depth of 30 m is around 70 days, and correspondingly 110 days between the surface and a depth of 60 m.

Seasonal variations of the sea-surface temperature and the thickness of the upper mixed layer in the GoF were studied using various models. Tyrväinen (1978) used the Kraus-Turner 1D-model (Kraus and Turner 1967) in order to simulate the annual cycle of surface temperature and the thickness of the upper mixed-layer in the western GoF. Her main conclusion was that the mixed-layer depth can be forecasted with an accuracy of 5 m, and that the mixed layer temperature can be predicted with an accuracy of 2–4 °C, the predicted values being always too small. One major reason for the inaccuracies is that the role of advection is not described in the model. Omstedt (1990) and Omstedt and Nyberg (1996) simulated wintertime conditions in the Baltic Sea area using a 1.5-dimensional model, where they resolved the vertical dimension and parameterized the horizontal effects. The calculated sea-surface temperatures, e.g. in the GoF, showed good agreement with measurements.

Tamsalu and Myrberg (1995) used a two-dimensional model, in which advective processes were included. Their results reproduced quite well the seasonal changes of the surface temperature in the central GoF (Fig. 5). For June–August, when a well-developed thermocline exists, the model simulated the temperatures with errors of usually not more than 1–2 degrees. For autumn, when the mixed layer is deepening because of convection and mechanical mixing, there are inaccuracies in the model results. This is a result of many factors; inaccuracies in the atmospheric forcing (wind, temperature) and in the calculation procedure for the mixed layer thickness, as well as in the way of parameterizing the heat exchange between the upper mixed layer and the thermocline. Local upwelling phenomena also cause difficulties to prognoses because of shortcomings in the model dynamics (*see* Upwelling). So, in order to simulate accurately the sea-surface temperature, detailed parameterization of vertical processes is needed in the model; the use of 1.5 dimensional models (*see* above) or fully three-dimensional models is necessary.

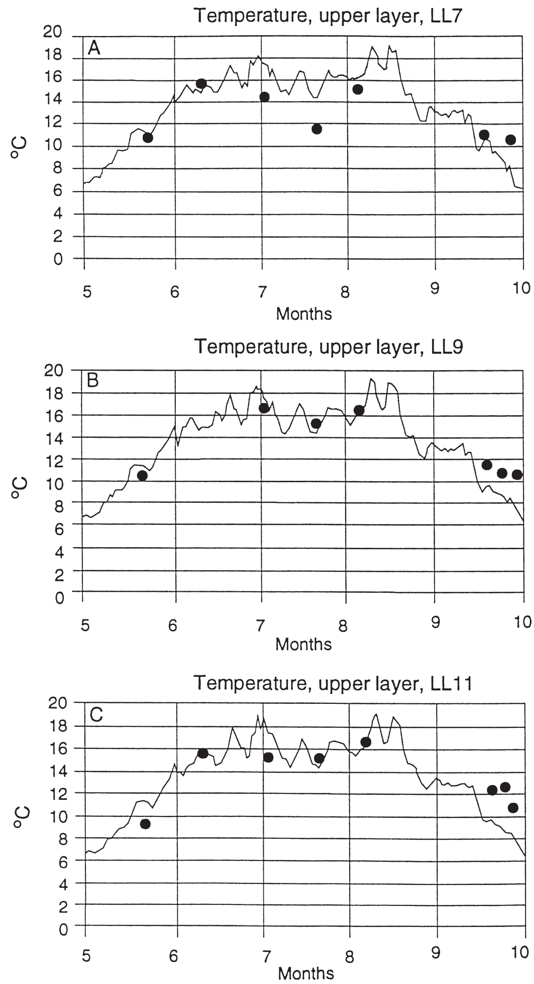

Fig. 5. Time evolution of surface temperature in the central Gulf of Finland in 1992: observed values (black points) and those modelled by a 2D-numerical model (solid line). — A: station LL7. — B: station LL9. — C: station LL11 (from Tamsalu and Myrberg 1995).

In the western GoF, a permanent halocline exists throughout the year between depths of some 60–80 m (Fig. 2). The existence of the halocline prevents vertical mixing of the water body down to the bottom. The deep waters below the halocline are thus decoupled from direct atmospheric forcing. The renewal of the deep waters in the western GoF is therefore caused by horizontal advection of the deep water from the Baltic Sea

Proper. Towards the east, the difference between surface and bottom salinities decreases. Haapala and Alenius (1994) concluded that the influence of river runoff is seen both in the long-term variations and in the seasonal variations of salinity. This feature is typical along the whole GoF where the salinity has wide seasonal variations. The surface layer salinity decreases from winter to midsummer and at the same time the salinity of the deep layers increases. In early spring the salinity isolines are evenly inclined. During the spring and summer the isohalines turn into more horizontal position in the surface layer while in the bottom layer the isohalines shift towards the inner GoF. This shows that the observed salinity variations are caused by advection and that the dynamics of the GoF play an essential role in the seasonal variations of salinity. The surface layer salinity variations are easily understood as being related to the melting of the ice cover and the increased spring-time Neva runoff. The surface layer outflow seems to generate an inflow into the GoF in the deeper layers (Haapala and Alenius 1994).

Physical processes

The geographic features and water mass properties of the GoF form the background for numerous physical processes that on the one hand depend on these properties and on the other hand try to change them. The main effect of these processes is to mix the water masses and substances on different space and time-scales. Some processes play, however, a different role; they cause convergence of the water masses and substances, forming frontal zones and reinforcing gradients. In the following sections we try to outline the nature of both types of such processes in the GoF.

Circulation

The main forcing factor for currents in the GoF is the wind stress (the main wind direction is from the south-west). The density-driven (buoyancy) currents also play an important role in the overall circulation due to the pronounced horizontal buoyancy gradients caused by variations of temperature and salinity. The sea-surface slope that results from the permanent water supply to the eastern part of the GoF also contributes appreciably to the existing circulation pattern. The GoF is large enough to see the effects of the earth's rotation on the circulation, too (Witting 1912, Palmén 1930, Hela 1946).

The classic studies of Witting (1912) and Palmén (1930) on the general circulation of the northern Baltic Sea elucidated the annual resultant surface current field of the Gulf of Finland and the Gulf of Bothnia. The residual circulation field is cyclonic (counter-clockwise in the northern hemisphere) with currents having a horizontally variable speed and stability. The stability, or persistency, of the current was defined as the ratio between the vector mean speed and the scalar mean speed. The annual mean residual vector speeds are of the order of 1–2 cm s^{-1} and the stability between 6% and 26% (Palmén 1930). However, the stability of the currents during different seasons is much larger than that for the whole year (*see* Fig. 6). The relatively poor stability of the currents on an annual time-scale indicates that the counter-clockwise mean circulation is a statistical property, not a constant phenomenon. This fact is often forgotten, with the cyclonic circulation treated as a permanent circulation type for the GoF.

The low stability of the currents shows the importance of the current variability associated mainly with the variability in the wind forcing. Even though there is a large continuous fresh water input from the rivers in the eastern GoF, the mean vector velocity to the west is almost an order of magnitude smaller than the average speed near the Finnish coast. The stability of the currents is smaller along the northern coast than along the southern coast, which means that although there may exist a long-term mean circulation, the instantaneous currents can be very variable both in speed and direction. Periodic processes like the inertial oscillations and seiches of the GoF and of the combined Gulf of Finland–Baltic Proper system make the current system even more non-homogeneous and non-stationary.

Despite the difficulties in determining what the mean circulation in the GoF is, many studies have been devoted to this problem. The question of the permanent flow in the Gulf of Finland was dealt with by Palmén (1930) and by Hela (1952) on the basis of studies of observations from the

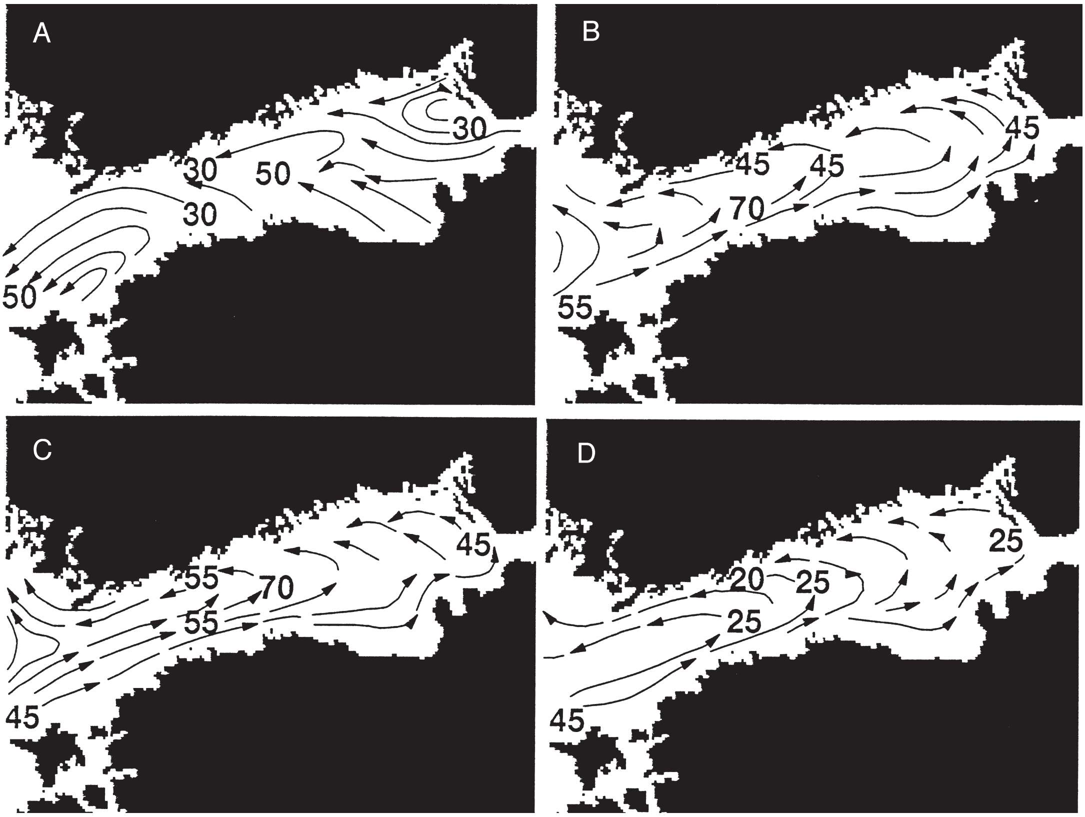

Fig. 6. Average mean surface circulation in the Gulf of Finland in (A) June, (B) August, (C) October, (D) June-October (from the Atlas of Finland 1910). The numbers in the figure show the stability of the current at different locations (*see* also the text).

lightships of Helsinki and Tallinn. These researchers tried to eliminate the drift currents from the actual current observations in order to get the permanent flow as a remainder. The drift current was defined as the flow in stationary wind and current conditions: a condition that very seldom exists. They found that the permanent flow, determined in this manner, was still dependent on the wind. Another way to roughly estimate the permanent flow is to calculate the resultant of the surface currents observed during calm weather (Hela 1952). Palmén (1930) and Hela (1952) used yet another current component definition: the characteristic current. This is the sum of the permanent flow and an additional current deflected about 20° to the right of the wind (Hela 1952). The characteristic current speed was found to be very slow, 0.6–2.0 cm s^{-1}, off Helsinki on the northern coast and somewhat more, 1.7–4.9 cm s^{-1}, off Tallinn on the southern coast. Hela (1952) also gave a result with a practical use: the ratio between the total surface current (v_C) and the wind (v_W) speed: $v_c = 0.0137 \times v_w$. This can be stated in other words as a rule of thumb: the surface current speed is 1.4% of the prevailing wind speed. The mean deflection angle of the current vector was 19° to the right of the wind direction. This deflection from the wind direction is naturally not valid near coasts.

The question of an average background flow was also discussed by Sarkkula (1989, 1991) and Pitkänen *et al.* (1993). They constructed a linear regression between the easterly and northerly wind components and the easterly component of the current at a depth of, e.g., 5 m, off Kotka in the northern Gulf of Finland based on 25 days of observations. They interpreted the constant of the regression equation, –4.8 cm s^{-1} (r^2 = 0.69), as being the speed of the background flow. They suggested the horizontal salinity gradient and bottom relief as reasons for the background flow. The value for the background flow given is somewhat larger than the mean residual flow calculated by Palmén (1930) for the ice-free period. This may follow from the fact that Palmén used much longer time series (5 years of observations) in which seasonal differences damp each other. Another possibility appears when considering the short-term large-scale dynamics of the GoF. The dominating south-westerly winds pile up the water at the eastern end of the gulf. In a situation where the wind stress is zero (the zero point of the regression), the water level returns to an equilibrium stage and thus causes water transport to the west. In this situation, the baroclinic effects and circulation caused by the river water runoff do not necessarily play a major role. The linear wind regression probably gives too high an estimate for the background flow because the wind stress is seldom zero for long enough for the equilibrium stage to be reached.

An indirect estimate of the long-term mean flow towards the west was determined from the time difference between the annual minimum of the surface layer salinity at Utö and the maximal river runoff (Launiainen 1982). The value, 2.5 months, implies a mean flow of a couple of centimetres per second (Mälkki and Tamsalu 1985).

In a real situation the wind and density-driven currents are coupled by strong non-linear interactions that can be seen for example as eddy-like formations. Thus separation of the residual circulation (or permanent flow, or background flow) from the observed current field can be regarded as somewhat artificial. The importance of the residual, or background, flow is mainly in long-term estimates of the drift of substances, but even there the interpretation is not trivial. Leppäranta and Peltola (1986) studied the applicability of a random transport model to the Baltic Sea. They compared cases where no permanent current was used and a case where a permanent flow derived from Palmén (1930) was introduced. One of the results was the probability distribution of the transport after a one-year period from a prescribed starting point (in the GoF they used Helsinki). This distribution shows that though there is a probability maximum in the west, the probability density is non-zero in the east, indicating eastward net transport in some of the cases.

The spatial and temporal variability of the currents were investigated by several authors, who based their studies on analyses of moored current measurements. Talpsepp (1986) and Talpsepp *et al*. (1994) indicated that current speeds are generally higher on the Estonian coast than on the Finnish coast. Strong currents (with a velocity of about 20–30 cm s^{-1}) on the Estonian coast can sometimes even be called coastal jets. The long-term mean (scalar) speed off Helsinki is only about 10 cm s^{-1}. The current variability was reported to be conspicuous for its 18–20 day periods (Talpsepp *et al*. 1994). Unidirectional currents prevail over several days and then change direction (Laakkonen *et al*. 1981, Talpsepp *et al*. 1994). Eddy-like short life-time structures that affect the circulation in the near bottom layer were observed by Mälkki and Elken (1988), Mälkki and Talpsepp (1988) and Talpsepp and Mälkki (1989). The studies of Laakkonen *et al*. (1981) and Haapala *et al*. (1990) off the Hanko peninsula (in the Tvärminne area) and Alenius (1986) near Kotka showed that mean surface current speeds are low in summer (3–5 cm s^{-1}) and in autumn (4–6 cm s^{-1}). In the study of Haapala *et al*. (1990), data from the years 1987–1989 were used. The mean current direction was westwards at the coast, supporting the cyclonic circulation idea. However, the current measurement showed that strong westerly winds can turn the currents eastward on the Finnish coast even for several weeks at a time. The largest instantaneous flow speeds were recorded in summer, when the wind stress only distributes into the relatively thin upper mixed layer above the thermocline.

Recent measurements in 1994–1995 (unpublished data), carried out by the Finnish Institute of Marine Research at five stations across the entrance region of the GoF show that across the GoF there exist regions of flow in opposite directions. The typical velocity in these flows is of the order of 10–20 cm s^{-1} in the surface layer. Simultaneous Acoustic Doppler Current Profiler (ADCP) observations (Stipa 1994) show still more clearly these regions of opposite flows. Because the flow conditions remained quite steady for several days, those ADCP measurements across the GoF can be considered representative of that situation in spite of their instantaneous character. They show

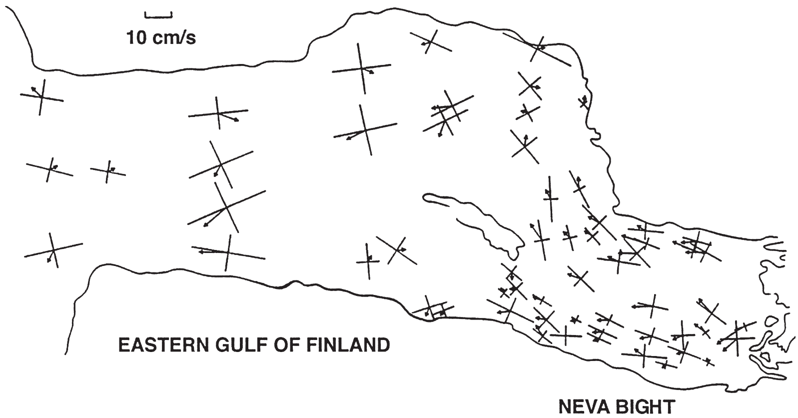

Fig. 7. Surface currents observed in Neva Bight and the eastern Gulf of Finland (east of 29°E) illustrated as main axes of dispersive ellipses and the mean vector, averaged over periods from two weeks to six months. The major and minor axes of the scattering ellipses are shown as crosses with a mean vector directed basically to the west with a speed reaching 10 cm s^{-1}. (from Belyshev and Preobrazhensky 1988).

that the oppositely-flowing currents had dimensions of about 10 km in the horizontal.

The observations at the entrance of the GoF support the idea of cyclonic circulation at least in the summer time when the stratification is well developed. On the Finnish side the mean vector velocity was 5–9 cm s^{-1} westwards during the summer and the stability of the current was high (near to or above 50%). In the middle of the GoF the current direction is highly variable. On the Estonian coast the current direction was mainly north-eastwards. In the autumn there was also a long period of inflow along the Finnish coast.

The eastern end of the GoF is, from a dynamic point of view, an estuarine of the GoF, where the current system has a specific nature. The study of local currents in the eastern GoF, though started as early as 1911–1912, was particularly stimulated by the design and construction of the flood barriers. The greatest number of regular observations was obtained during 1981–1990 for a fixed grid of stations mainly situated within the Neva Bight but also with a number of stations located to the west of the "dam". The general picture of currents in the Neva Bight and the extreme eastern GoF was discussed in numerous Russian studies (*see* e.g., Altshuler and Schumacher 1985, Azernikova and Monosov 1981, Mikhailov and Shakhverdov 1977, Shakhverdov *et al.* 1981). The most comprehensive and reliable results were obtained and described by Belyshev and Preobrazhensky (1988) on the basis of observations from the navigational seasons.

The main and general feature of the current fields in the eastern GoF area is their non-stability, with pulsating components usually exceeding the mean (residual) values. This is clearly seen in the scattering ellipses of the observed instantaneous current vectors (Belyshev and Preobrazhensky 1988; Fig. 7). Though the mean vector is directed basically to the west with a velocity value reaching 10 cm s^{-1}, it is unlikely that any persistent "streams" could actually exist even in this part of the GoF with the possible exception, perhaps, of the middle of the Neva Bight (the "transit zone") where the influence of the Neva discharge is most clearly manifested. In the central part of the Neva Bight the movements are close to "reverse" (the major axis is 3–5 times as large as the minor one) while in the coastal zones the movements in various other directions are more probable. The observed characteristics of the current fields in the Neva Bight were supplemented and more closely analysed by the application of numerical modelling (Voltsinger *et al.* 1990). This study allowed the Bight to be divided into four domains with different current regimes showing the occurrence of counter-currents and eddies in some parts of the Bight under various conditions (small level variations, considerable rise of level, etc.).

Long surge waves, responsible for the floods in St. Petersburg, cause radically special conditions in the easternmost gulf. Extraordinary strong horizontal movements related to these waves seem to determine the orientation of the major axes of the current ellipses coinciding with the longitudinal axis of the Neva Bight. The wind effects can also be considerable. Prolonged easterly winds, reinforcing the mean movement to the west, cause at the same time a compensating eastward movement of

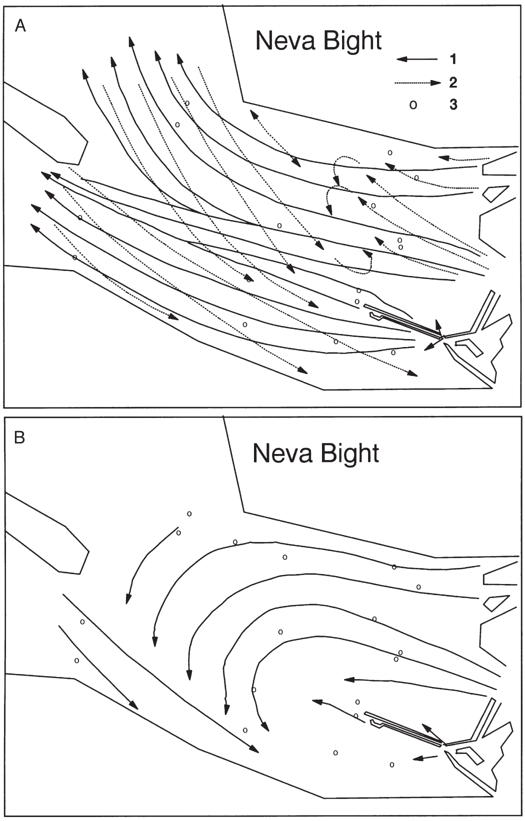

Fig. 8. Effect of water level variations on the currents in the Neva Bight. — A: Sketch of currents in the Neva Bight under the ice cover: — 1: the sea level rises or falls no faster than 1–2 cm h^{-1}; — 2: the sea level rises faster than 4 cm h^{-1}; — 3: stations with measurements. — B: Sketch of currents in the Neva Bight under the ice cover when the sea level rises with a speed of 2–4 cm h^{-1} (from Skriptunov 1974).

more saline water along the bottom of the Marine Channel (fairway; *see* Fig.1). Westerly winds may block the westward flow and reverse it to become eastward with velocities not exceeding 10 cm s^{-1} in 60%–70% of cases and reaching 25 cm s^{-1} in not more than 3%–4% of cases. Joint analysis of currents, wind and sea level oscillations allowed the division of the Neva Bight into five sub-regions having different summer current regime characteristics.

In winter, the characteristic pattern of currents in the Neva Bight is more homogeneous than in summer because the ice cover isolates the water from direct action of the wind. The mean transport velocities reduce by 20%–30 %, with pulsation decreasing by 1.5–2 times in the northern part of the Bight and by 5–6 times in its southern part. The mean transport holds in direction, but actual currents remain dependent on the surge long waves. Three types of sub-ice current regime have been established depending on the sea-level oscillations, based on observations made in 1957–1974: (1) normal (directed to the west, no significant sea-level change); (2) backward (directed to the east, rapid sea-level rise); (3) transient. Usually, the transition from normal to backward type proceeds in a clockwise manner (Skriptunov 1974, Skriptunov 1977; Fig. 8).

The regime of currents in the eastern GoF beyond the Neva Bight (to the west of the "dam") is characterised by even more space-time variability, with very slow mean transport not allowing the emergence of any ordered pattern in the mean circulation (Fig. 7) (Belyshev and Preobrazhensky 1988). Instantaneous current speeds exceed the mean transport velocity by as much as 5–10 times. The orientation of the major axes of the scattering ellipses confirms the fact that the main pulsation in velocities results from the surge long waves penetrating from the west.

From the preceding paragraphs it can be concluded that the circulation processes in the GoF fall into a large range of space-scales, from basin-wide down to small-scale vortices. The broad band of meso-scale processes encompasses the phenomena having dimensions of from tens of kilometres to kilometres. To estimate a relevant length scale, the Rossby radius of deformation R_n is commonly used. It describes the scale of geostrophic adaptation (adjustment) to either the barotropic ($n = 0$) or baroclinic ($n > 0$) dynamic response of water to an external disturbance (Konjaev and Sabinin 1992). Correspondingly, the barotropic or baroclinic (internal) long wave velocity c_n determines either the external or internal Rossby radius. The internal Rossby radius of deformation is a horizontal scale at which rotation effects become as important as buoyancy effects (Gill 1982). So, at length scales small compared with the internal Rossby radius, the adjustment is approximately the same as in a non-rotating system. The value of the barotropic Rossby radius is large compared to basin scales of the Baltic Sea (Fennel *et al.* 1991).

Fennel *et al.* (1991) calculated the internal Rossby radius of deformation for the different basins of the whole Baltic Sea using Brunt-Väisälä frequency profiles. In the GoF the internal Rossby radius in spring and summer is about 2.5 kilometres, reducing to 1.3 kilometres in autumn. The differences between seasons are connected with the changes in the stratification. Fennel *et al.* (1991) concluded that the changes in the internal Rossby radius are mainly connected with the changes in the thermocline, because the depth of the halocline remains nearly constant. In other parts of the Baltic Sea, the internal Rossby radius is usually larger than in the GoF, being between 3 and 7 kilometres. Only in the Belt Sea are radii smaller than 2 kilometres found. The scale of the internal Rossby radius of 1–3 kilometres in the GoF should be the minimum accuracy of horizontal resolution of future numerical models in order to describe the baroclinic features of the GoF properly. This space-scale should be kept in mind when measurement campaigns are planned.

Horizontal structures of salinity and temperature — fronts

The overall thermohaline structure of the GoF may be considered as principally understood. It is governed by spring-summer heating and stratification (except for the Neva Bight where the vertical homogeneity caused by mixing prevails), autumn-winter cooling and convection, voluminous river discharge and the rather irregular inflow of bottom saline water. The horizontal variations of temperature are mainly determined by the large variations in incoming solar radiation throughout the year. Local upwelling (*see* Upwelling) can cause rapid changes in sea-surface temperatures. The horizontal variation of salinity in an east-west direction is about 6‰–7‰/400 km, which gives an impression of the highly baroclinic nature of the GoF.

The large-scale horizontal temperature and salinity structures are associated with circulation processes. The most interesting and important phenomena are the boundaries between different water masses: i.e. areas of large horizontal gradients. These can be formed either between meso-scale circulation patterns or in transition zones between sea areas, e.g. in straits. These fronts or frontal zones are usually convergence zones with strong biological activity. The close coupling between biological and physical processes in such fronts has risen considerable interdisciplinary interest during the last decade in the frontal dynamics of the GoF (Talpsepp 1993, Elken 1994, Kononen *et al.* 1996, Pavelson *et al.* 1996, Laanemets *et al.* 1997). The distribution, dynamics and special physical features of salinity and temperature can nowadays be investigated by many different methods: satellite images, CTD-measurements, batfish, current meters, thermistor strings, ship-borne ADCP's, drifting buoys, etc. This has considerably intensified the study of frontal processes.

An attempt to clarify the study of Baltic Sea fronts was made by Pavelson (1988). He classified the fronts into five groups: (1) quasi-permanent salinity fronts, (2) salinity fronts, (3) temperature fronts, (4) density-compensated fronts, and (5) upwelling fronts. He also gave typical characteristics for these front classes, although he stated that this classification was only preliminary. This classification can in some sense be considered to be phenomenological.

Another possibility of classifying fronts is to stick to the definition of the front. By analogy with the definition of fronts in meteorology, we can define an oceanic front to be an area with a large horizontal buoyancy gradient. The buoyancy depends on temperature and salinity. Thus we can simply classify the fronts as (1) temperature fronts, in cases where temperature dominates the buoyancy variations, (2) salinity fronts, in cases where salinity is the dominating factor, and (3) thermohaline fronts where both temperature and salinity vary across the front. This classification is a dynamic one.

In this classification, the density-compensating situation where both temperature and salinity have large gradients but, however, compensate each other in the buoyancy should then be discussed. Such zones may behave as fronts for biology without having any large gradients in buoyancy. These cases are probably rare. Usually density-compensating "fronts" are rather weak and are formed in the upper layer of the sea by the deformation field of eddies, in an area where initially weak positively-correlated horizontal trends of temperature and salinity field existed (Pavelson 1988).

The GoF is a suitable area for the generation of fronts because of the large natural salinity variations. Mechanisms that generate and maintain the fronts in the GoF have been studied to some extent. In certain areas the bottom topography steers the deep, more saline, currents towards the north, favouring frontogenesis across the GoF (Talpsepp 1993). Upwelling at the coasts can lead to more saline coastal boundaries, generating fronts with salinity gradients of some 1.2‰ in three nautical miles (Mälkki and Talpsepp 1988). The deformation fields of meso-scale motions can enhance the TS-gradients against diffusion, and thus generate and maintain fronts (MacVean and Woods 1980, Elken 1994). Under strong or prolonged wind action convergent density-driven flows maintain and restore fronts after their decay (Elken 1994). The shape and location of the fronts are evidently determined by the wind-induced advection (Pavelson *et al.* 1996).

Salinity fronts of a quasi-permanent nature are the most intense and of the largest scale with relation to the general circulation (Pavelson 1988). The frontal area at the entrance to the GoF is of this type. It separates water of the Baltic Proper from the less saline water of the GoF (centre at 22°15´E). The salinity gradient of this front is typically 0.1‰ km^{-1} and the salinity difference across the front 0.5‰ at its maximum. Synoptic-scale salinity fronts are the most frequent (Pavelson 1988). They are, however, relatively rare in summer probably because of an insufficient energy supply for frontogenesis. The distance between the fronts is larger than the internal Rossby radius. The salinity gradients in such fronts are of the order of 0.03‰–0.05‰ km^{-1}.

The potential existence of quasi-permanent frontal areas in the GoF was pointed out in numerical model experiments, too. The verified model results of Tamsalu and Myrberg (1995) gave a clear indication of the strongly baroclinic origin of the salinity fields in the GoF (Fig. 9). The horizontal non-homogeneity of the salinity patterns becomes visible in different ways depending on the time-scale of the investigation. In the verified model, results of the monthly averages of the vertical mean salinity (July, August and October 1992), the horizontal salinity field shows a complex baroclinic structure. Three major frontal zones located in the (1) easternmost part (Seskar basin), (2) central part, and (3) the entrance of the GoF could be identified. These fronts are quasi-stationary; their location and intensity varies, but on the average maps they are clearly visible. The instantaneous pattern of salinity is determined by the local wind stress distribution, stratification and river runoff. The frontal areas for the whole Baltic Sea were simulated by Elken (1994). He used the free-surface version of the 3D Bryan-Cox model (Killworth *et al.* 1991), which was applied to the Baltic Sea by Lehmann (1995). The model could e.g., reproduce the quasi-permanent salinity front at the entrance of the GoF.

The salinity front in the south-western GoF was the subject of specially intensive investigation in Finnish-Estonian cooperation during the last years (Pavelson 1988, Mälkki and Talpsepp 1988, Talpsepp 1993, Elken 1994, Kononen *et al.* 1996, Pavelson *et al.* 1996, Laanemets *et al.* 1997). Talpsepp (1993) discussed the possible factors effecting the location and intensity of this front and concluded that topographic steering effects and coastal upwelling were the major contributors. The front is often formed between a less saline tongue of surface water in the middle of the GoF and more saline coastal water. As was indicated earlier, the reason for more saline coastal water, especially on the northern side of the gulf, should then be the coastal upwelling.

Kononen *et al.* (1996) discussed this salinity front rather thoroughly. They concluded that the frontal system is usually oriented east-to-west and forms a transition zone between the saltier waters of the open Baltic flowing in along the Estonian coast and water masses of lower salinity, originating from the inner GoF, and flowing out along the Finnish coast. The frontal area thus separates different water masses. They showed that the stratification conditions on the less saline and on the more saline side of the front clearly differ from each other, which is an important factor influencing the mixing conditions. On the weakly stratified saline side, the ratio between the kinetic energy from the wind and the potential energy of the water mass due to the stratification (Richardson number) is more favourable for vertical mixing of the water column than on the less saline side. They observed a near-slope jet current on the saline side in the intermediate layer. Increased concentrations of dissolved phosphorus were measured at a depth of 15 m. The vertical transport of nutrients could therefore take place by a two-step mechanism. First the shear-induced turbulence in the jet current can transport nutrients from the intermediate layer to a shallower

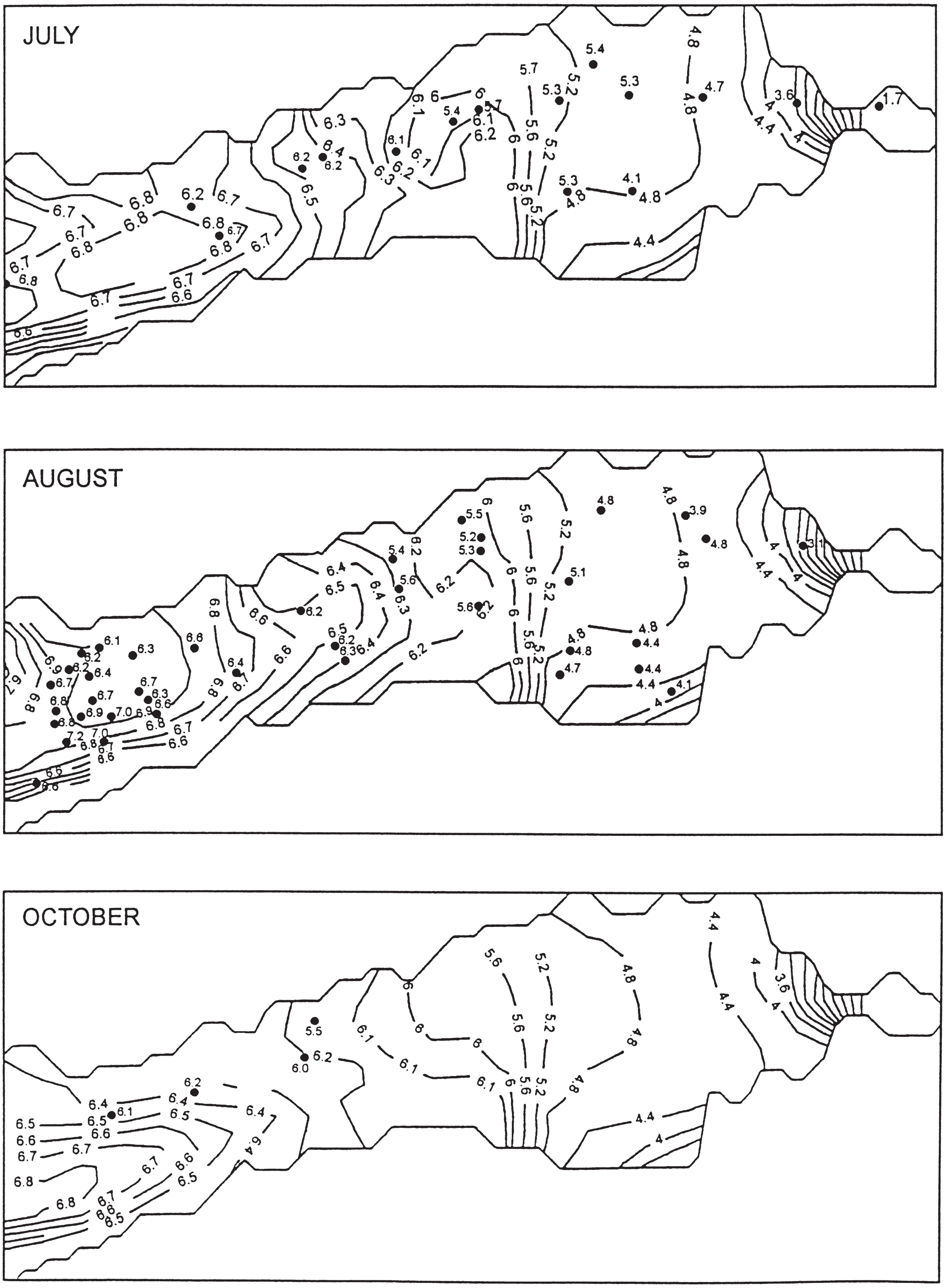

Fig. 9. Horizontal fields of the vertical mean of salinity in 1992; each panel presents monthly mean values. Observed values are marked with a black point with the corresponding value. For practical reasons isoline analyses of the model results have been made at intervals of 0.1‰ in the western part and at intervals of 0.4‰ in the eastern part. (from Tamsalu and Myrberg 1995).

depth and then episodic intensive wind mixing, capable of overcoming the diminished energy barrier, causes nutrient pulses in the upper mixed layer.

Pavelson *et al.* (1996) noticed that the offshore movement in easterly winds of the denser open Baltic water was accompanied by a sharpening and stronger inclination of the front at the entrance to the gulf. Under conditions of westerly winds the less dense GoF water moved onshore, overriding the denser water. Their calculations, using a semi-empirical wind-forcing model, showed that the changes of potential energy (the measure of stratification changes) in the upper mixed layer were related mainly to advection induced by along-front wind stress and wind-generated vertical mixing.

The horizontal temperature field in the GoF is also non-homogeneous both in space and time. Kahru *et al.* (1995) found from satellite images that temperature fronts occur predominantly in areas of straight and uniformly sloping bottom topography. Major frontal areas exist on the north-western coast of the GoF. The main effect that produces these fronts is the coastal upwelling, complemented by coastal jets, eddies, differential heating and cooling and water exchange between basins with different water characteristics. Strong winds in regions of adjacent water masses of different vertical stratification can cause the formation of frontal areas. Temperature fronts are usually accompanied by only moderate salinity changes (Pavelson 1988). The upwelling fronts in the north-western GoF are present under various weather conditions (*see* Upwelling). The upwelling filaments that emerge when the upwelling front along the coast becomes unstable are a common feature there. Kahru *et al.* (1995) concluded that it is possible that water and substance transport by these filaments plays a more important role than the mean circulation.

One of the most pronounced salinity fronts in the open GoF is at the juncture of the eastern shallow area adjacent to the Neva Bight and the deeper and wider basin lying to the west, the "Seskar basin". This area is of great interest for many reasons. The width of the sea increases considerably there, a fact which appears to be a contributory factor in the intensification of local sedimentation. It behaves as a kind of buffer for pollution transport from the Neva mouth to the west. The numerical experiments carried out by Andrejev and Sokolov (1989) and by Andrejev *et al.* (1990) with a 3D hydrodynamic model support the theory of the existence of this buffer zone. After a two months simulation, in which passive tracers were initially released from the Neva mouth area, the largest concentration of the tracers was still in the Neva Bay. The concentration rapidly decreased towards the dam area, and west of it the concentration was negligible.

Upwelling

Upwelling is an important process in bringing nutrient-rich water from deeper layers to the surface, in mixing water masses and in generating frontal areas. The most typical condition for upwelling (in the northern hemisphere) is that the wind direction is parallel to the coastline with the coast on the left hand side of the wind.

The upwelling phenomena was studied on the eastern coast of the Hanko peninsula e.g. by Alenius (1976), Hela (1976) and Haapala (1994). Alenius (1976) concluded that the upwelling is associated with along-shore winds (south-westerly winds in the study area), and the surface temperature drops about 6 °C in three days. Hela (1976) gives a first approximation for calculating the vertical velocity w in connection with upwelling by assuming that w is linearly proportional to the wind speed. He concludes that, because the Baltic Sea is relatively shallow and the isopycnal levels are in many places rather close to surface, the effects resulting from upwelling must usually be more profound than one would expect on the basis of experience from the oceans. According to Hela (1976) there was e.g. a case in the Tvärminne area in 1959 when a strong westerly wind blew for about five days. This was enough to produce a strong upwelling, in which the surface water temperature dropped from 21 degrees to 6 degrees. The values of vertical velocity in this case were between $4–10 \times 10^{-3}$ cm s^{-1}.

Haapala (1994) concluded that wind events (south-west winds) must usually prevail for at least some 60 h in order to generate upwelling in the coastal area of the Hanko peninsula. The required wind impulse seemed to depend on the stratification of the water column. During thermal stratification a 4 000–9 000 kg m^{-1} s^{-1} wind impulse per unit area (e.g. wind speed of 6 m s^{-1} with a duration of 50 h) is needed to generate upwelling. When the sea is thermally homogeneous, a 10 500–14 000 kg m^{-1} s^{-1} impulse is required (e.g. wind speed of 10 m s^{-1} with a duration of 50 h or 5 m s^{-1} for 180 h). This implies that during strongly stratified conditions the wind stress only has a direct effect on a rather short water column above the thermocline, while in thermally homogeneous conditions the influence of the wind penetrates deeper and more wind energy is needed to generate upwelling. According to Haapala (1994), surface salinity changes were negligible during upwelling, whereas below the thermocline salinity could increase by about 1‰. During summer, the surface temperature can drop several degrees. The changes in nutrients were also significant during upwelling, a fact which can have important biological consequences. The biological consequences of upwelling at Tvärminne Storfjärd were studied by Kononen and Niemi (1986). Upwelling can also take place in the open sea-area due to favourable bottom topography and baroclinic currents, which can cause upward motion too.

Bychkova and Viktorov (1987) studied upwelling areas from satellite images. They found that in the GoF the most important upwelling areas are at the entrance i.e. on the Finnish side near the Hanko peninsula and on the Estonian side in the area east of Hiiumaa. Upwelling typically takes place outside the Hanko peninsula because of the south-westerly winds with the coast on the left of the wind. The study by Kahru *et al*. (1995), in which time series of satellite images were examined and analysed, supports the view that the area off the Hanko peninsula is an important upwelling area. Upwelling-like phenomena (cold spots on the surface near the coast during and after the relevant wind conditions) have been noticed in the Luga-Koporye region (on the southern coast of the GoF) during the expeditions of the Russian State Hydrometeorological Institute (Baltic Floating University activities).

Horizontal turbulence

Horizontal turbulence in the coastal regions of the GoF plays a significant role in the transportation and spreading of substances such as the contaminants discharged from rivers and the runoff from agricultural land. The parameters of turbulence that are needed to estimate the intensity of the exchange between the coastal zone and the open sea are highly variable. However, some characteristic values of these parameters are necessary for assessment and employment in modelling, and these should be determined for the region of interest by direct field measurements. The parameter of most practical use in ecosystem modelling aiming at optimisation of the disposition and arrangement of sewage discharges is the coefficient of eddy diffusion.

Ivanov and Mikhailov (1972) and Mikhailov (1974) studied horizontal turbulence on the basis of current measurements at different distances from the coast. They estimated mean values of coefficient of eddy diffusion, the rate of turbulent kinetic energy dissipation, the scale of turbulence (the effective dimensions of eddies), the lifetime of a diffusing substance cloud and its path. They found that in the eastern part of the GoF horizontal turbulence is relatively restricted by the smaller dimensions of the basin, but the rate of energy dissipation is greater there. Rather different results were obtained by Nikolaev and Luvsk (1975) on the basis of moored current measurements.

In other cases, the parameters of horizontal turbulence in the coastal zone of the GoF were studied using the Lagrangian approach i.e. by observing floating indicators (drifters, buoys, dye spots). Mikhailov (1981) used airborne methods for keeping track of floating drifters in the eastern GoF for different "phases" (rise and fall) of sea level oscillations. Their results show that horizontal turbulence appears to be less intense in coastal regions than in more open areas. An attempt to estimate the intensity of subsurface horizontal turbulent exchange was done for the Luga Bight (V. Y. Chantsev pers. comm.) by keeping track of a non-inertial floating drifter with subsurface sail.

Experimental studies of horizontal turbulence under the ice cover were done by Smelov *et al.* (1985) using radioactive indicators in the northern part of the Neva Bight. These measurements were made within a fairly small horizontal scale range (up to 100 m), and showed the dependence of coefficient of eddy diffusion on scale of turbulence and on ice conditions. The influence of ice manifested itself on the one hand in an amplification of the hydraulic resistance reducing the turbulent diffusivity, and on the other hand in inducing additional disturbances in the water layer adjacent to the lower surface of the ice.

Sea level, seiches and tides

Variations in the sea level are forced by three main factors: changes in the wind direction and speed, fluctuations in the air pressure and changes in the density of the sea water (Lisitzin 1958, 1974, Ekman and Mäkinen 1996). The sea level has been measured systematically and scientifically along the coasts of Finland since 1887, Hanko being the first station. These measurements are meaningful, even globally, because their reference level is fixed to the bedrock. Thus, the time series can be considered homogeneous and are very little disturbed by the local effects of the measuring station. The only relevant disturbing effect is the land uplift which, according to the most recent analysis in the GoF, is about 2.7 mm y^{-1} in Hanko and 1.7 mm y^{-1} in Hamina (Vermeer *et al.* 1988; for earlier estimates *see* Hela 1953). The decreasing trend of the water level in relation to the bedrock has been quite clear during this century. Recent studies by Kahma (1995), however, indicated a possible statistically significant change in that trend. The sea level decrease seems to have stopped. The true reason for this is under investigation (K. K. Kahma, pers. comm.).

Several systematic studies on sea level oscillations and tides in the northern parts of the Baltic Sea, including the GoF, were made by Witting (1911), Hela (1944), Lisitzin (1944, 1959a, 1966) and Stenij and Hela (1947). The frequency distributions of water level at the Finnish sea level stations were given by Stenij and Hela (1947) (data from Helsinki, Hamina, Viipuri and Koivisto in the GoF) and Lisitzin (1959a) (data from Helsinki and Hamina). This work is routinely continued and these frequency distribution tables for the coasts of Finland are continuously updated by the Finnish Institute of Marine Research.

The frequency distributions show e.g., that in Helsinki the sea level variability is smallest in July (between –36 and +79 cm compared to the mean sea level), and largest in January (between –92 and +136 cm compared to the mean sea level). Thus both the extreme minimum and maximum water levels occur in winter, whereas in summer, the water level varies much less around the mean sea level. The maximum sea level values (compared to the mean sea level) along the Gulf of Finland were +123 cm in Hanko (9.I.1975), +136 cm in Helsinki (27.I.1990) and +166 cm in Hamina (7.XII.1986). The minimum values (below the mean sea level) were –78 cm in Hanko (10.III.1934), –92 cm in Helsinki (22.III.1916), and –110 cm in Hamina (20.XI.1975).

Sea level data can be used not only in sea level studies but also in investigations into differing climate conditions. It seems possible that the variance of the sea level variations has decreased during this century. This may indicate changes in the cyclonic activity in the area after the first decades of the century (Makkonen *et al.* 1981).

It can be seen that the intensity of sea level oscillations increases from west to east. In the eastern GoF high water levels have been a real trouble, in the form of floods, for the population and authorities of St. Petersburg since its foundation in 1703 because of the low-lying position of the city. A water level rise of 1.6 m above mean sea level is considered a flood. Since 1703 the city has been flooded more than 280 times, i.e. the probability of a flood is about once a year. Table 1 shows the list of the highest (catastrophic) floods.

Table 1. Extreme floods in St. Petersburg (in cm above the mean).

Year	1777	1824	1924	1955	1986
Sea level (cm)	321	421	380	286	260

According to observations and estimations, the spectral structure of sea level oscillations in the Neva Bight is characterised by energy peaks concentrated mainly in the proximity of periods of 200, 90, 60, 30–40, 24–25, 12–13, 6–8 hours. During the period of the autumn storms, oscillations with periods of 30–40 hours are especially developed. The most significant rises related with floods occur with periods from 10–12 to 20–30 hours (Anonymous 1989). In winter the damping effect of the ice reduces the oscillations by 10%–15%.

The main cause of major sea level oscillations in the GoF is a storm surge produced by a deep atmospheric depression over the Baltic. The storm surge occurs together with this depression from the west principally in the form of a forced progressive wave. On its way to the eastern GoF this wave is reinforced by a decrease in the basin cross-section as well as by local stormy winds. The sea level oscillations may grow from tens of centimetres at the entrance to the GoF to over 3–4 m at the eastern end, accompanied by strong currents of up to 2 m s^{-1} in the coastal zone and up to 1 m s^{-1} in the open sea.

The building of the dam cutting off the Neva Bight from the eastern GoF and protecting St. Petersburg from storm surges and floods started at the beginning of the 1980s (Rodenhuis 1992). Although this building work has not been completed yet, the constructions already erected have exerted a perceptible impact on sea level oscillations by filtering out and damping constituents with periods of less than about 6 hours (L. Preobrazhensky, pers. comm.). The numerical simulations by Andrejev and Sokolov (1992) are in agreement with the analysis of sea-level measurements. The model result showed that the dam filters waves with periods of less than 8.5 hours. In general the dam does not have a major influence on the longer period oscillations that are of practical interest.

Although the largest sea level rise is related to a forced progressive wave reaching the eastern end of the GoF, a significant part of sea level oscillations is caused by free-standing waves, seiches, that are manifested once the external forcing ceases. Seiches in the Baltic Sea were studied for the first time by Witting (1911), who concluded that the period of an uninodal seiche of the Baltic Proper-GoF system is between 28 and 31 h. The results of a careful harmonic analysis of the sea level observations in Hamina by Lisitzin (1959b) showed that the period of the uninodal oscillation of the system Baltic Proper-GoF system is on average 26.2 h. Practically, all the Baltic Proper seiche modes reach the GoF. Krauss and Magaard (1962), using a channel theory, revealed four main modes encompassing the GoF with periods of 27.4, 19.1, 13.0 and 9.6 hours related to 1-, 2-, 3- and 4-nodal mode patterns. Wübber and Krauss (1979) used a refined 2D-model in which the Coriolis effect was taken into account, as well as the response of the Baltic water body to an air pressure forcing function. They found that the four main two-dimensional Baltic modes encompassing the GoF have periods of 26.4, 19.8, 13.04 and 10.45 hours. These results have been supported by the studies of Klevanny (1994) and Klevanny *et al.* (1994) in which a two-dimensional numerical model was used. They obtained about 27 h for the period of the most important seiche of the GoF.

The Baltic Sea is practically a non-tidal sea. The mean amplitudes of the tidal components in sea level are only millimetres or some centimetres (Lisitzin 1944), and their effects practically disappear beneath other short-term variations of the sea level. In some exceptional cases the tidal range (difference between high and low water) can reach values of 0.1 metres in the Hanko–Helsinki area (Witting 1911). Magaard and Krauss (1966) made an extensive study of the tides in the Baltic Sea. They constructed co-tidal maps for the diurnal and semi-diurnal tidal constituents separately and calculated relevant tidal amplitudes for the whole Baltic Sea.

Tidal oscillations are more pronounced in the eastern GoF, where the maximum possible ranges can reach 20 cm in Primorsk (Koivisto) and 24 cm in the mouth of the Neva river (based on the data of Jakobi (1923), Berezkin (1936), Lisitzin (1944), Pomytkin (1977) and Altshuler (1980). The spectral structure of the tides in the GoF is characterised by the considerable predominance of diurnal (O_1 and K_1) constituents over semi-diurnal (M_2 and S_2) ones. The maximum tidal ranges therefore occur in June and December during the "tropical" intensification of diurnal tidal oscillations. The corresponding tidal currents are also predominantly diurnal and weak. According to Evdokimov *et al.* (1974), the diurnal tidal currents in the eastern GoF can reach 8 cm s^{-1} and the semi-diurnal ones 4 cm s^{-1}, giving a maximum possible value reaching 12 cm s^{-1}.

Surface waves

Both the theoretical and practical the aspects of surface waves have been investigated in the GoF. Surface waves are an important factor in the physics of the GoF as well as for seafaring.

Kahma and Pettersson (1993) gave surface wave statistics for the GoF based on wave buoy measurements in 1982–1985. The observation site was almost in the middle of the GoF at 59°58.5´N, 25°14.0´E, near Kalbådagrund, where the bottom depth is about 60 m. According to their statistics, wave conditions in the GoF can be characterised by mean significant wave heights between 0.5 m in spring and 1.3 m in winter. The corresponding wave peak periods are 3.8 s and 5.5 s. Maximum significant wave heights were 1.8 m and 3.8 m (with the maximum now in the autumn), with corresponding periods of 5.7 s and 8.5 s. Pettersson (1991) studied the possible maximum wave occurring once in 100 years in the GoF. She used the same wave data set for estimating wave distributions and probabilities. The wind statistics were based on the observations of the Finnish Meteorological Institute at the Kalbådagrund caisson lighthouse and at Katajaluoto near Helsinki. According to Pettersson (1991), the empirical distributions of the data follow a Weibull probability distribution. By using this and the wind statistics for corrections she got a relation between the cross-over level of the significant wave height and the maximum wave. The final result was 3.8 m for the cross-over level of the significant wave height and 7.1 m for the maximum wave height. Limiting factors for the growth of waves in the GoF are the narrowness of the gulf and the decrease of the effective fetch due to the refraction caused by the bottom topography (Pettersson 1991).

Lopatukhin and Shatov (1988) calculated the surface wave characteristics for the whole Baltic Sea on the basis of more than 30 000 en-route ship observations carried out in the years 1955–1975. Their results show that the strongest winds over the GoF (with a mean speed of 7–9 m s^{-1}), and accordingly the most developed surface waves (mean height 1.4–1.5 m in the western part of the GoF and 0.8 m in the eastern part) take place during autumn and winter. Summer is the calmest season with a mean wind speed of about 5–6 m s^{-1}, being slightly higher in the western part of the GoF than in its eastern part. In summer, the mean wave height does not exceed 1 m while the maximum wave height may reach 5–6 m (Lopatukhin and Shatov 1988).

These two surface wave statistics based on quite different types of observations well support each other, and we can assume that the results well describe the overall surface wave conditions in the whole GoF area. The western end of the gulf opens into the northern Baltic Sea Proper, where the surface wave conditions during autumn storms can be the most difficult for seafaring in the entire Baltic Sea area.

The GoF, as well as other parts of the Baltic Sea, has been a suitable area for theoretical studies of wave growth because sometimes the conditions are ideal for them almost resembling laboratory conditions and such as practically never occur in the oceans (Kahma 1995). The factors that limit the growth of surface waves were studied by Kahma and Pettersson (1994) using data from the GoF and from the Gulf of Bothnia. They showed that the narrow fetch geometry prevailing in the GoF has important effects on the wave growth. Taking these effects into account for practical purposes requires special knowledge. For estimating the waves from wind speeds and fetch, nomograms published by Kahma (1986) may be used also in the GoF; then, however, the inaccuracy of such generalised nomograms must also be considered.

The surface wave climate in the Baltic Sea has become an important issue for sea faring safety after the tragic accident of the passenger ferry Estonia on September 28, 1994, in which over 850 people lost their lives. The refraction of the waves due to the bottom topography can in certain rare conditions (suitable wind conditions) focus the surface waves so that in the focus area the waves grow much higher than in surrounding areas (Kahma 1995). This shows the effect of very local conditions on the physical processes of the GoF.

Designing and building the "dam" has set the task of estimating the protection against the wind waves for the Neva Bight afforded by this construction. This problem was considered by means of prognostic calculations (Nikolaev 1981). Maximum wave heights were found to be from 2.5 to 3.3 m in the Marine Channel. A moderate protection effect (up to 20%) was obtained in westerly winds only within a zone of about 800 m behind the dam. In easterly winds no effect of the dam was found inside the Neva Bight.

Ice conditions

Ice cover is a regular winter phenomenon in the GoF. It affects the physics of the gulf and its sea-

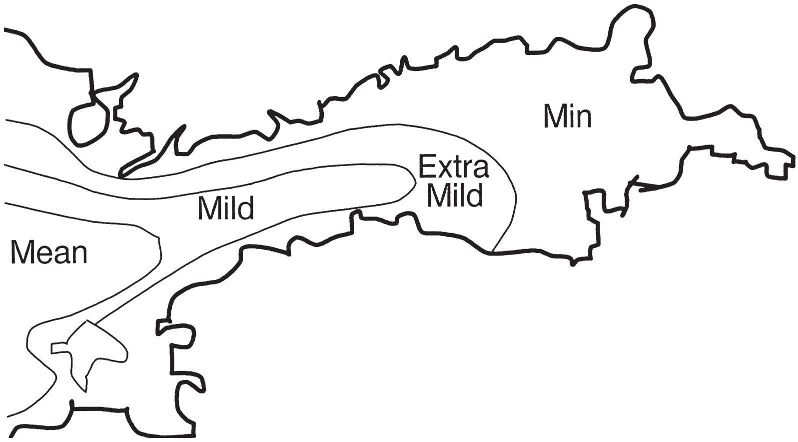

Fig. 10. Largest ice extent of the Gulf of Finland in different ice winter types (extracted from Seinä and Palosuo 1993).

faring dramatically. The piling-up of ice against the coasts is a severe problems for marine navigation. Because of its importance to human activity, ice conditions in the GoF have been very well documented for about the last 100 years. The statistical properties of ice conditions were published as statistical studies and ice atlases (Jurva 1937a, Palosuo 1957, 1965, Anonymous 1982, Leppäranta and Seinä 1982, Leppäranta *et al.* 1988, Seinä and Peltola 1991, Seinä and Palosuo 1993, Seinä 1994).

In spite of its great importance, the question of ice is considered to be slightly outside the main scope of our paper. Thus we only refer to some results of ice research in the GoF. The eastern GoF freezes every winter and the same is true for most of the rest of the GoF (Fig. 10) (*see* Jurva 1952b, Seinä and Palosuo 1993). In general, total freezing is much more probable for the GoF than for the Baltic Proper, e.g. the GoF was completely covered by ice 22 times between 1958/1959 and 1982/1983, while the Baltic Proper has been only twice covered (in 1939/1940 and 1941/1942) during the 20th century (Drabkin and Kljachkin 1990). The average date of formation of permanent ice cover at the entrance to the GoF is February 10, and the ice disappears from the GoF during mid-April (Seinä and Peltola 1991). The average number of ice days varies from 40 at the entrance to the GoF to 130 in some coastal bays in the north-eastern GoF (Seinä and Peltola 1991). The total volume of ice in the GoF varies from a few to about twenty cubic kilometres, of which about 20% are accounted for by the ice thickness increments due to hummocks resulting from ice compression of the ice floes (Drabkin and Kljachkin 1990).

Some general features of the ice cover in the Neva Bight are outlined by Nezhikhovsky (1988). The formation of ice starts there 1–2 days after the stabilised transition of the air temperature to sub-zero values. In calm and frosty weather conditions the whole Bight is covered with fast ice in 2–3 days, but with only a slight frost and a strong wind this process may take up to 2–2.5 months. The mean ice winter in Kronstadt lasts about 120 days (from the end of December to the end of April). Westerly winds and a sea level rise may lead to a breaking of the ice cover and its movement, accompanied by the creating of hummocks. The largest ice thickness (90–95 cm) was observed on the Neva Bight in the winters of 1939–40 and 1941–42. Most of the ice in the Neva Bight also melts there in the spring. The Neva Bight receives ice from the Neva river and Lake Ladoga (on average 25 million tonnes). That ice also melts in the Bight. Kljachkin (1992) forecasted the position of floating and fast ice boundaries and other ice characteristics (drift velocities, concentration, hummocking parameters) using a hydrodynamic model.

The influence of the ice cover on the underlying water masses has been studied much less than the ice conditions themselves. Only a few studies of hydrography and currents in the GoF during the ice-covered period exist. Such studies are generally very local, confined mostly to the vicinity of nuclear power plants (*see* e.g. Hari 1978, 1982 and 1984). For the Neva Bight a study of currents under the ice was made by Skriptunov (1977; Fig. 8) who investigated the relations between the currents and level oscillations produced by surge waves in winter. The damping effect of the ice cover on the sea level oscillations was considered in Kupriyanova and Freidson 1981, Kupriyanova and Tretyakova 1981 and Monosov *et al.* 1981.

The role of the ice in transporting and concentrating both airborne material and material of coastal origin is an issue that has not been studied in the GoF, although it may deserve some attention.

Marine meteorological conditions and air-sea interaction

The scale of the GoF is small compared to the typical length scale of extratropical cyclones, which is of the order of 3 000 km (*see* e.g. Holton 1979). The meteorological forcing in that sense would therefore be expected to be rather uniform over the GoF. However, the GoF is narrow and surrounded by land areas in all directions except to the west. Thus, patterns of atmospheric temperature and wind stress are very in-homogeneous because of the variable surface roughness and variable heat exchange between the sea and the atmosphere.

The wind distribution (*see* e.g. Launiainen and Laurila 1984) shows that the most common wind direction is south-west and the mean wind speed at the open-sea station of Kalbådagrund in the GoF is about 7–8 m s^{-1} (data from 1977–82). Using Hela's (1952) equation, the resulting sea-surface current speed would be about 10 cm s^{-1}. According to Launiainen and Laurila (1984), the highest wind speeds are observed during the winter, with the highest monthly mean wind speed at Kalbådagrund being 9.4 m s^{-1} in November. The highest wind speeds occur during the winter due to the strongest cyclonic activity then. The calmest period is the early summer, when typically high pressure is found over Scandinavia. The lowest monthly mean wind speed at Kalbådagrund is 5.5 m s^{-1} in July. The mean annual course of the wind speed and its variance over the sea can be expressed by a harmonic wave.

The oceanographic processes are very much wind-dependent. The dependency is not only confined to wind speed, but also to the horizontal variability of the wind field — the curl of the wind stress. The number of representative wind stations in the GoF area is very limited. The only real open-sea meteorological observation station in the GoF is at present the automatic weather station at Kalbådagrund (59°58´N, 25°37´E). The other stations are usually located near the coastline. Launiainen and Saarinen (1982) compared atmospheric parameters between Kalbådagrund and the coastal automatic sea-mast station in the Loviisa area (60°22´N, 26°22´E) of the GoF. They found that the wind speeds over the open sea are much higher than those near the coast besides which the ratio between these wind speeds is not a constant. The wind speed difference is especially dependent on wind direction, which represents the effects of variable surface roughness and orography. The atmospheric surface layer stability also plays a significant role. The wind speed ratio between the open-sea and coastal stations is smallest for winds from the open sea directions, from south-east to south-west, and correspondingly largest from land directions from west to north-east. For winds from the land, the open sea winds are 1.3–1.6 times higher than those observed in the coastal area (Loviisa). Under unstable conditions the stability effect diminishes the wind speed ratio, while under stable conditions the stability effect is the opposite.

The wind pattern over the GoF is also characterised during both spring and summer by the sea-breeze in coastal areas. The main factors contributing to the sea breeze are the temperature difference between land and sea areas and the effects of the prevailing flow. The sea-breeze is most typical during early summer, when the air-temperature is still low over the sea, but the surrounding land areas are strongly heated during the day. A pronounced horizontal temperature (density) difference develops between sea and land, which induces a circulation system, in which the wind at the surface is directed from the sea to the land, while in the upper layers (ca. 1.5 km) it blows from the land to the sea.

The results of a numerical sea model (Myrberg 1997) show that the horizontal resolution of the numerical weather prediction model used (ca. 55 × 55 km) is not accurate enough to describe the complicated non-homogenous patterns of wind speed and direction, temperature etc. over the narrow GoF. However, the results by Myrberg (1997) clearly show that even the use of the present meteorological forcing fields improves the accuracy of the sea model's forecasts compared to those simulations, in which the atmospheric input is taken only from the data of a single marine weather station.

The interaction between the sea and the air can be described in terms of the energy balance at the sea-surface. The energy balance at the sea-surface consists of the sum of incoming solar radiation, outgoing losses of energy through long-wave radiation and latent and sensible heat fluxes. The annual cycles of sea-surface temperature and the mixed-layer thickness are principally determined by the energy balance. In the GoF the energy balance of the sea-surface is positive between April and August. This is mainly because of the strong solar radiation during that part of the year. Strong variability in the fluxes of latent heat and sensible heat also takes place in the GoF area. These fluxes can reach values of $\pm$ 100–200 W m^{-2}. Detailed studies of air-sea interaction in the GoF area, especially near Loviisa, were carried out by Launiainen (1976, 1977, 1979).

The main components of the sea-surface heat balance — the radiation balance, the sensible heat exchange and the latent heat exchange — were estimated for the GoF during the "Baltica" Project (1983). The radiation balance averages between 1.4 and 1.6 GJ m^{-2} y^{-1}. The value is higher over the open part of the GoF and lower near the coast. It is positive from April to September and negative from October to March. The annual mean values of the sensible heat exchange with the atmosphere seems to increase towards the east; the reverse is true for the annual means of the evaporation heat loss (Nõmm 1977).

The heat content of the GoF is a measure of the heat balance. It is highly variable during the year in accordance with the above-mentioned variable heat exchange between the sea and the atmosphere. The heat content has an annual course similar to the net heat balance at the sea-surface; however, it has a time lag compared with the maximum surface incoming solar radiation because of the thermal inertia of the sea. The heat content of various sea areas adjacent to Finland including the GoF was already calculated earlier by Jurva (1937b).

Discussion

Having summarised the published literature on the physics of the GoF, we now give in this section some thoughts on subjects that should be studied further and some indications as to which processes are less understood than others.

The geography of the GoF is well-known, except for the detailed bottom topography. The present needs for knowledge on topography for modelling and sediment studies have proved to be greater than those for navigation in the open-sea parts of the GoF. Even surface wave conditions in this area depend so much on exact bottom topography (K. Kahma, pers. comm. related to the Estonia accident) that improved information is needed, at least for some special areas.

Discussion about the general circulation of the GoF is commonly focused on the background flow or mean cyclonic circulation. This is probably correct for long-term processes, but for estimating dispersion of substances along the coasts, shorter time-scales are relevant. Then the variability of current speed and direction and current field patterns are driven by different dynamics, in which transient features play an important role. This can lead to spreading of substances in opposite directions to that which the mean flow idea suggests. Also contemporary biological research needs more accurate knowledge on the meso-scale dynamics of the currents.

Knowledge of the spatial and temporal variability of the currents in the GoF should be improved by using modern measurement techniques e.g. ADCP, batfish and numerical modelling. As for the measured data, one should aim at getting field information sufficient for its presentation in the form of, say, scattering ellipses (*see* Figs. 7 and 8) for the GoF as a whole. This requires a great deal of field work over a long period (several years at least), which is hardly possible at present. The resolution of the models should be increased to the scale of the internal Rossby radius (about 1–3 km in the GoF). An increase in model resolution has been limited by lack of computing power and too coarse a knowledge of bottom topography. The possibility of carrying out the verification of present models has been very limited, too. When these limiting factors are overcome, the energy transfer between different scales and the coast-open sea interaction and the water exchange between the Gulf of Finland and the Baltic Sea Proper can be studied by uniting measurement campaigns and modelling. The modelling efforts suffer, however, from too coarse an atmospheric input, even from the highest resolution meteorological models presently available.

The overall temperature and salinity conditions can be considered to be more conservative than the currents. The annual cycles of temperature and salinity are well known from observations and can be modelled quite accurately. However, high-resolution horizontal temperature and salinity structures, such as frontal areas, are, by the same reasoning as for the currents, not well enough known. The frontal areas are known to be important for biological processes because of their active small-scale dynamics. Also the question of semi-permanent frontal areas in the GoF seems to still need more investigation.

Sea-level oscillations are relevant for environmental issues because the rapid sea-level changes indicate movements of large water masses and the oscillations can effectively wash the coasts. Similar reasoning holds for studies of surface waves. Their effects on resuspension in the GoF is largely unknown.

The question of horizontal turbulence is essential in estimating the interaction between the coastal and open sea waters. This is an important issue in the Baltic Sea, where contaminants originating from the coasts have been observed in the open parts of the sea as well. This is closely linked to environmental decision-making and planning for the outlets of purified waste waters, as an example. Turbulent processes that cause mixing in the water body have not been adequately investigated. This problem includes both vertical and horizontal processes. A particular problem to be investigated is that of which spatial and temporal scales are mostly responsible for really effective contaminant transports. It is felt that, in spite of the great importance of direct field measurements of the parameters of turbulence, the relevant research is not sufficiently advanced in the Gulf of Finland.

Although the ice conditions of the GoF can be considered to be well-known, the dynamics of the water masses under the ice cover is still poorly known. The "annual" residual flows, although constructed on the basis of data from several years, still mainly represent ice-free conditions. Thus winter-time investigations are needed, although they can be difficult because of the continuous movement and piling-up of the ice.

The unified field activities should be focused on regions playing a particular role in the physical and chemical processes of the GoF: the entrance to the Gulf from the Baltic Sea Proper, and the Seskar basin acting as a buffer for substance fluxes and quasi-permanent frontal zones. On the other hand, certain relatively restricted areas (coinciding favourably with the boxes of models) could be selected for especially detailed investigations serving as control sites for verification and comparison with the results of modelling.

The GoF is a challenging marine environment. It is too large to be only an estuary, but small enough to be covered reasonably well by international joint expeditions. It features a large spectrum of physical processes from small-scale turbulence to meso-scale eddies and basin-wide circulation. The main physical features of the GoF, large fresh water runoffs coupled with pronounced gradients of hydrographic variables, and the related environmental problems in a highly populated area, are common to many estuarine seas around the world. The results of this paper therefore represents to some extent a more general overview of estuarine dynamics.

The "Gulf of Finland Year 1996" project has certainly increased our knowledge of the physics of the GoF and raised more interest in the yet-unsolved problems. This knowledge of the different processes supports the development of numerical models. Thoroughly validated numerical models can then be used operationally, and may also assist in decision-making and estimating the response of the sea to different loading scenarios.

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Appendix 1. Typical parameter values for the Gulf of Finland

Length	400 km
Width	48–135 km
Mean depth	37 m
Maximum depth	123 m
Area	29 571 km^2
Volume	1103 km^3
Mean river inflow	114 km^3 y^{-1} (3 600 m^3 s^{-1}, of which the Neva accounts for 2 400 m^3 s^{-1})
Drainage area	420 990 km^2
Precipitation	593 mm y^{-1}
Evaporation	490 mm y^{-1}
Water exchange estimate at the entrance	600 km^3 y^{-1} out, 480 km^3 y^{-1} in
Mean surface salinity	0‰–7‰ (from east to west)
Mean horizontal salinity gradient along the gulf	0.016‰ km^{-1}
Freezing point temperature at the surface	–0.17 °C (east) to –0.33) °C (west)
Maximum density temperature	4 °C (fresh water) to 2.56 °C (at 7‰)
Mean maximum surface temperature	16.5 °C (can be well over 20 °C) at the very beginning of August
Baroclinic Rossby radius of deformation	1–3 km
Mean residual current speed	1–2 cm s^{-1}
Typical current speed in the upper layer	10 cm s^{-1}
Stability of the current	6%–26%
Typical vertical velocity during upwelling	(3–10) $\times$ 10^{-3} cm s^{-1}
Mean wind speed in the open sea area	7 m s^{-1}
Mean significant surface wave height	0.5 m (in spring)–1.3 m (in autumn)
Mean surface wave peak period	3.8 s–5.5 s
Average number of ice days	40 (west)–130 (east)

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