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Abstract

1. Geographical position, geological development, hydrographical features, climate and physical drivers together create the Baltic Sea environment.
2. Baltic Sea water is brackish and characterised by pronounced salinity gradients, both in horizontal and vertical directions, because of the large volume of freshwater runoff from over 100 rivers, which mixes with the saline water from the Kattegat that enters the Baltic Sea via narrow shallow straits.
3. Being a semi-enclosed continental sea with a large drainage area compared to its water volume, the Baltic Sea ecosystem is heavily impacted by the surrounding landmasses.
4. The water residence time in the Baltic Sea is long (30–40 years), and therefore discharged nutrients and toxic compounds circulate within the sea for a long time, which contributes to its vulnerability to eutrophication and chemical contamination by hazardous substances.
5. The Baltic Sea Area is geologically young and the Baltic Sea ecosystem is extremely young in an evolutionary perspective. Only few macroscopic species are fully adapted to its low-salinity environment.
6. Chief factors that affect species distributions in the Baltic Sea along local, regional and ecosystem-wide gradients are salinity, climate, ice cover, currents, permanent salinity stratification, hypoxia, and benthic substrate types (rock, sand, mud).
7. Environmental drivers vary either in time or space or both and contribute to the north-south “large-scale Baltic Sea gradient”, along which many species experience physiological stress, lose the ability to reproduce sexually and reach the ecological limit of their occurrence.
8. In an ecosystem-wide perspective, the large-scale Baltic Sea gradient is the principal ecological characteristic of the Baltic Sea.

Keywords

Ecological characteristics • Environmental gradients • Geography • Geology • Human impacts • Hydrography

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2.1 The Baltic Sea in perspective

2.1.1 Baltic Sea, East Sea and West Sea

The brackish Baltic Sea is an arm of the Atlantic Ocean, extending from northern Germany and Poland in the south almost to the Arctic Circle in the north (Fig. 2.1). It is a continental mediterranean (inland) sea, *i.e.* a semi-enclosed sea that has limited water exchange with the ocean, in which the water circulation is dominated by salinity and temperature differences rather than by winds. Although the salinity of the Baltic Sea is far below 30, and thus not within the marine range (*cf.* Fig. 1.10), it is classified as one of the world's 66 “large marine ecosystems” (<http://www.lme.noaa.gov>). The marine influence on the Baltic Sea ecosystem is large, but it is strongly influenced by freshwater as well.

The Baltic Sea separates the Scandinavian Peninsula from the rest of continental Europe and is surrounded by nine countries. The Latin name “Mare Balticum” (Baltic Sea) was already used in the 11th century and may refer to the Danish straits (known as “belts”) or because the sea stretches through the land as a belt. It is also possible that “baltas” refers to the white colour of ice and snow.

In four riparian countries of the Baltic Sea, its name has a component sounding similar to “Baltic”: in Latvia (Baltijas jūra), Lithuania (Baltijos jūra), Poland (Morze Bałtyckie) and Russia (Baltiyskoye Morye). In four other countries, the Baltic Sea is known as the “East Sea”: in Denmark (Østersøen), Germany (Ostsee), Sweden (Östersjön), and

even in Finland (Itämeri). The latter name does not reflect the sea's geographic position in relation to Finland, but is the result of the long common history of Sweden and Finland as one country (until ~200 years ago). In Estonia the name is, geographically correctly, the “West Sea” (Läänemeri).

2.1.2 The large-scale Baltic Sea gradient is unique

The Baltic Sea exhibits gradients of critical environmental drivers, created by the sea's semi-enclosed location and the strong influences of the surrounding landmasses and climate. The large-scale Baltic Sea gradient from temperate marine to subarctic limnic is unique in the world with respect to the broad ranges of the environmental drivers in combination with the large geographical size of the area.

Surface-water salinity, the strength of the halocline (a jump layer in salinity) and surface-water temperature decrease northwards in the Baltic Sea, while the influence of a winter ice cover increases. The depth of the halocline increases from the Kattegat to the Baltic Sea proper. Nutrient dynamics, superimposed by human-made nutrient emissions, generate low productivity in the north, high productivity in the south and eutrophication in the east.

The large-scale Baltic Sea gradient strongly affects the open-water and coastal systems. However, the coastal areas are also influenced by local gradients in salinity, temperature, depth of the photic zone and ice cover, as well as by

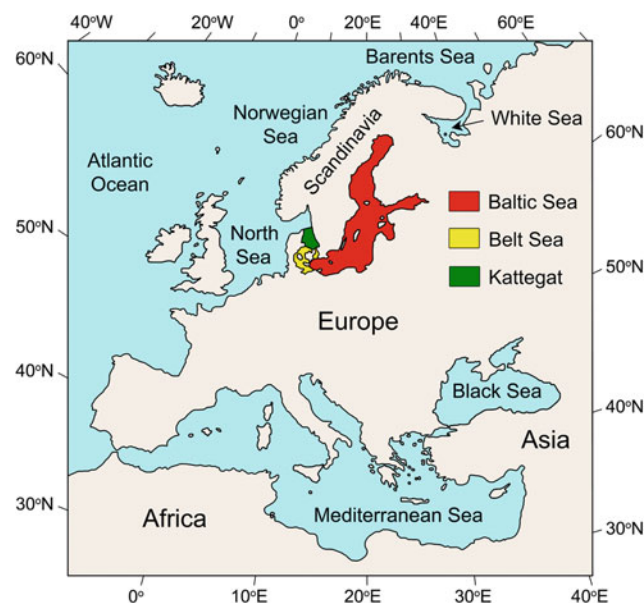


Fig. 2.1 Map of Europe, showing the semi-enclosed position of the Baltic Sea and its transition zone to the North Sea (Belt Sea and Kattegat) in northern Europe. Figure: © Pauline Snoeijs-Leijonmalm

other factors such as bottom type and water movement. Taken together, all these environmental drivers create a mosaic of habitats, supporting different types of organisms, populations and communities.

2.1.3 Not an estuary

While the Baltic Sea and estuaries on oceanic coasts share the common feature of being brackish transitional waters between freshwater and marine systems, the Baltic Sea is not an estuary (Elliott and McLusky 2002; McLusky and Elliott 2007). An estuary is a partly enclosed coastal body of water with one or a few rivers or streams flowing into it, and with a free connection to the open sea. An estuary usually involves the outflow from one river system only. The Baltic Sea receives outflows from over 200 rivers, is much larger and deeper than an average estuary and has narrow and shallow connections with the ocean. In the latter respect, the Baltic Sea would be more similar to a gigantic threshold fjord with a series of subbasins.

An estuary is subject to large daily and seasonal fluctuations in salinity due to factors such as freshwater runoff, tides and winds (*cf.* Sect. 13.1.2). The Baltic Sea, on the contrary, has a long, stable salinity gradient. Many organisms living in the Baltic Sea belong to species of estuarine origin. However, the Baltic representatives differ from their estuarine conspecifics by adaptations to Baltic Sea conditions, *e.g.* to a permanently submerged life in water with constant low salinity.

2.1.4 Comparisons with other water bodies

The Baltic Sea is one of the world's largest brackish water bodies, together with the Black Sea and the Caspian Sea. The Black Sea has a 25 % larger surface area than the Baltic Sea, while that of the Caspian Sea is about the same (Table 2.1). As they are deeper, the respective water volumes of the Black Sea and the Caspian Sea are 26 and 4 times larger than that of the Baltic Sea. However, since only ~30 % of the Black Sea is located on the continental shelf and the Caspian Sea is in fact a lake (being fully enclosed by land), it is safe to say that the Baltic Sea is the world's largest continental brackish-water sea.

The coastlines of the continents feature numerous semi-enclosed inland waters and some of them show certain similarities with the Baltic Sea (Table 2.1). Although located in a totally different climate zone and having a salinity higher than that of the ocean due to evaporation, the Red Sea and the Persian Gulf suffer, like the Baltic Sea, from large-scale hypoxia ($>0 \text{ mL O}_2 \text{ L}^{-1}$ and $<2 \text{ mL O}_2 \text{ L}^{-1}$)

and anoxic conditions ($\leq 0 \text{ mL O}_2 \text{ L}^{-1}$) because of limited water exchange with the ocean. However, the most hypoxic water body is the strongly stratified Black Sea; ~90 % of its water and ~75 % of its bottoms are anoxic because of its limited and shallow connection with the intercontinental Mediterranean Sea (Murray et al. 1989). In some areas of the Black Sea, hypoxia is also closely related to anthropogenic nutrient inputs. For example, between 1990 and 2000 the oxygen conditions on the northwestern shelf part of the Black Sea improved substantially when nutrient loads from the Danube river decreased because of the end of intensive farming through economic decline as a result of the dissolution of the Soviet Union (Rabalais et al. 2010). The tidal North Sea, the White Sea and the Hudson Bay have open connections with the ocean and in these water bodies hypoxia is local and mainly of anthropogenic origin.

The Chesapeake Bay, the Gulf of Ob and the Caspian Sea are examples of water bodies with salinity levels close to that of the Baltic Sea (Table 2.1). When organisms from such areas are introduced to the Baltic Sea, they may establish persistent populations and impact the ecosystem. As a brackish-water lake with a long evolutionary heritage, the Caspian Sea is an important source area for non-indigenous species introductions to the Baltic Sea (*cf.* Sect. 5.1).

2.1.5 Humans and the Baltic Sea

Human colonisation of the Baltic Sea Area started a few centuries after the end of the last glaciation (~15,000 years ago) and has continued without any notable interruption until today (Jöns 2011). The different developmental stages of the Baltic Sea since the ice age have provided food and means of transportation to the people living around them: palaeolithic reindeer hunters during the Baltic Ice Lake and Yoldia Sea stages, mesolithic and early neolithic hunters-gatherers and fishermen during the Ancylus Lake and Littorina Sea stages, and seamen and traders during the post-Littorina stage (Jöns 2011).

In the course of the 20th century, the Baltic Sea ecosystem degraded as a result of imprudent anthropogenic activities such as eutrophication, chemical pollution and overfishing. Some environmental conditions are improving (*e.g.* eutrophication, some forms of contamination), while new threats emerge (*e.g.* climate change, new forms of contamination). Today, the people and governments of the Baltic Sea countries are well aware of the ecosystem services provided by the Baltic Sea and the importance of wise ecosystem management for sustainable use of our common resource (*cf.* Sect. 18.5).

Table 2.1 Comparison between the physical properties of the Baltic Sea and those of some other semi-enclosed seas, large estuaries and lakes

Water body	Location centre	Climate type	Average winter ice cover (% of surface area)	Surface area (10^3 km^2)	Water volume (10^3 km^3)	Average depth (m)	Freshwater budget	Average surface-water salinity	Classification	Connection with the ocean	Lunar tides	Hypoxia
Chesapeake Bay	38°N, 76°E	Temperate	<5 %	12	0.17	6.5	+	15	Estuary	Open	Strong	In summer
Gulf of Ob	73°N, 74°E	Arctic	100 %	41	0.49	12	+	5	Estuary	Open	Strong	Local
North Sea	56°N, 03°E	Temperate	0 %	700	67	95	+	35	Semi-enclosed sea	Open	Strong	Local
Hudson Bay	60°N, 85°E	Subarctic-Arctic	100 %	1,233	123	100	+	30	Semi-enclosed sea	Limited	Strong	Local
White Sea	66°N, 37°E	Subarctic-Arctic	100 %	90	5	60	+	30	Semi-enclosed sea	Limited	Strong	Local
Mediterranean Sea	37°N, 17°E	Mediterranean	0 %	2,500	3,750	1,500	–	38	Semi-enclosed sea	Very limited	Weak	Local
Persian Gulf	27°N, 52°E	Desert	0 %	239	9	36	–	40	Semi-enclosed sea	Very Limited	Weak	Large-scale
Red Sea	22°N, 38°E	Desert	0 %	438	215	490	–	40	Semi-enclosed sea	Very limited	Nearly absent	Large-scale
Baltic Sea	60°N, 20°E	Subarctic-temperate	~50 %	369	21	57	+	5–8	Semi-enclosed sea	Very limited	Nearly absent	Large-scale
Black Sea	43°N, 35°E	Mediterranean-continental	~20 %	461	552	1,197	+	20	Semi-enclosed sea	Very limited	Absent	Extreme
Caspian Sea	43°N, 50°E	Continental	~20 %	371	78	211	0	12	Brackish lake	None	Absent	Local
Lake Superior	48°N, 88°E	Continental	~60 %	83	12	149	0	<0.1	Freshwater lake	None	Absent	Local

2.1.6 A vulnerable ecosystem

Owing to its natural geographical, hydrographical, geological and climatic features, the Baltic Sea may be considered an ecosystem with low ecological resilience, *i.e.* a system susceptible to change (*cf.* Sect. 4.8 and 17.2). For example, many organisms in the Baltic Sea live at the limit of their salinity distributions in species-poor communities with low functional diversity (*cf.* Sect. 4.5). Key species that dominate their functional group, and thus have crucial roles in the Baltic Sea ecosystem, are, in the pelagic zone, the piscivorous fish the Atlantic cod *Gadus morhua* (*cf.* Fig. 4.12) and the two planktivorous clupeids: the Atlantic herring *Clupea harengus* (*cf.* Fig. 4.12) and the European sprat *Sprattus sprattus* (*cf.* Fig. 4.12). In the benthic zone, key species are the habitat-forming macrophytes: the bladderwrack *Fucus vesiculosus* (*cf.* Fig. 4.27b) on rocky coasts and common eelgrass *Zostera marina* (*cf.* Fig. 12.13) on sandy coasts, in addition to the filter-feeding animals: the blue mussel *Mytilus trossulus* (*cf.* Fig. 4.29) on rocky coasts and the sand gaper *Mya arenaria* (*cf.* Box Fig. 5.1) on sandy coasts. Even if these species run a low risk of being lost from the entire Baltic Sea, there may be no other species capable of fulfilling their role in areas where they would become extinct. Threats to these species include *e.g.* overexploitation of the fish stocks by humans, habitat destruction in coastal areas and chemical pollution.

2.2 Geography

2.2.1 Geography and the distribution of organisms

The geographical position of the Baltic Sea creates environmental gradients that affect the distribution of species and impinge on evolutionary processes. The shallow sills at the entrance of the Baltic Sea, and between its subbasins (Figs. 2.2 and 2.3), influence species distributions by modifying environmental drivers such as salinity and oxygen supply. Insolation, sediment composition and activities on the land within the drainage area vary along the ~2,000 km long large-scale Baltic Sea gradient from the Skagerrak to the northern Bothnian Bay and eastern Gulf of Finland.

The sandy coasts that dominate in the south of the Baltic Sea host different biological communities than those found on the rocky coasts that dominate in the north. The densely populated southern drainage area, which is dominated by agricultural land-use, influences the sea biota in a different way compared to the sparsely populated northern part, which is dominated by boreal forests.

Typical of the ~8,000 km long Baltic Sea coastline are archipelagos with numerous skerries stretching out from the coastline into the open sea, as well as estuaries and lagoons which extend landward from the coastline. Estuaries, and often also lagoons and archipelagos, receive freshwater runoff from land, which creates local, and often temporally variable, gradients in *e.g.* salinity, nutrient availability, water movement and sedimentation all around the Baltic Sea coasts. These local environmental gradients also strongly influence species distributions.

2.2.2 The boundaries of the Baltic Sea

The “Baltic Sea Area” was defined by HELCOM (1993) as “the Baltic Sea proper with the Gulf of Bothnia, the Gulf of Finland and the entrance of the Baltic Sea bounded by the parallel of the Skaw in the Skagerrak at 57°44.8' N”. This is the boundary between the Kattegat and the Skagerrak, also known as the “Skagerrak-Kattegat front” (Fig. 2.2). Thus, the Baltic Sea Area includes the whole sea area that is significantly influenced by the brackish-water outflow of the Baltic Sea. The relatively deep Skagerrak (average depth 210 m) north of the boundary is part of the North Sea.

The shallow Kattegat and Belt Sea (average depths 23 m and 14 m, respectively) comprise the transition zone between the North Sea and the Baltic Sea (HELCOM 1996). The Kattegat and Belt Sea are part of the Baltic Sea Area (as defined by HELCOM), but not of the Baltic Sea *sensu stricto* (Fig. 2.1, Table 2.2). The transition zone is heavily influenced by marine water inflow from the Skagerrak as well as by brackish-water outflow from the Baltic Sea. Geographically, the whole Belt Sea, or the Belt Sea south of the Lillebælt sill and the Storebælt sill, are often considered part of the Baltic Sea. In the latter case, the southern part of the Belt Sea is referred to as the “Western Baltic Sea”. In a biological sense, as already recognised by *e.g.* Remane (1934), the southern Belt Sea typically belongs to the transition zone to the North Sea because many marine organisms can still live there.

The Baltic Sea (as used in this book) is the area east of the Belt Sea, being located between the outflow of the Szczecin Lagoon at the Polish-German border (latitude 53°55' N) and the estuary of the Torne älv at the Finnish-Swedish border (65° 48' N) (Fig. 2.1). From west to east, it stretches from the lower tip of the Danish island of Falster (longitude 11°59' E) to the inner Neva Bay in Russia (30°59' E). The western boundaries of the Baltic Sea consist of the shallowest sill between Denmark and Sweden in the Öresund (the Drogden sill, 8 m water depth) and the shallowest sill between Denmark and Germany (the Darß sill, 18 m, Lemke et al. 1994) (Figs. 2.2 and 2.3). The Drogden and Darß sills are also the natural biological

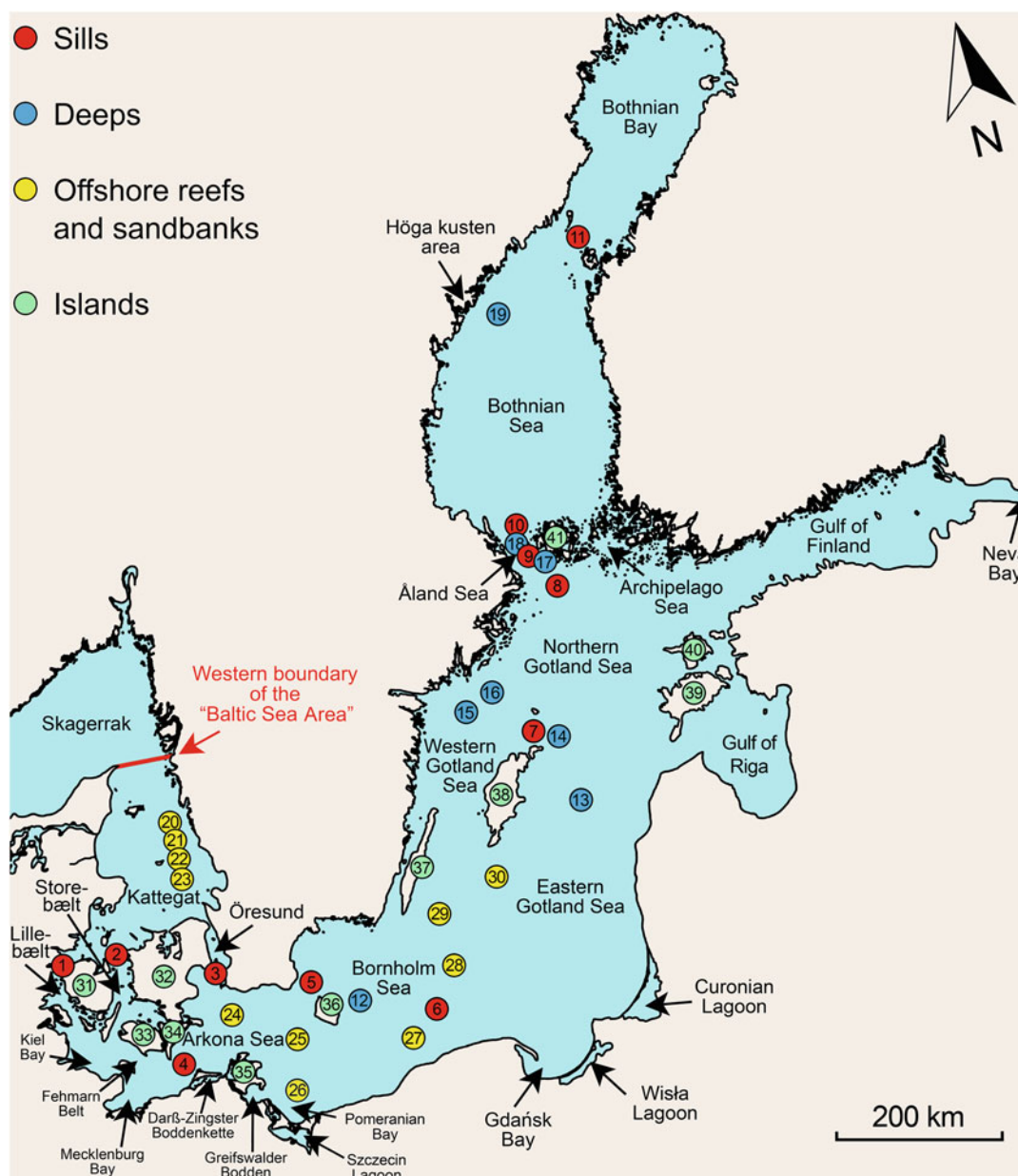


Fig. 2.2 Map of the Baltic Sea Area, showing its subregions. The Arkona Sea, Bornholm Sea, Gdańsk Bay and Gotland Sea are together called “the Baltic Sea proper”. The Arkona Sea, Bornholm Sea and the southern part of the Eastern Gotland Sea are together called “the southern Baltic Sea proper”. The Western Gotland Sea and the northern part of the Eastern Gotland Sea are together called “the central Baltic Sea proper”. The Northern Gotland Sea is called “the northern Baltic Sea proper”. The Bothnian Sea and the Bothnian Bay are together called “the Gulf of Bothnia”. Straits, channels and sills are indicated in red: 1 = Lillebælt (maximum depth 81 m), 2 = Storebælt (maximum depth 60 m), 3 = Drogden sill (8 m), 4 = Darß sill (18 m), 5 = Bornholmsgattet (maximum depth 45 m), 6 = Słupsk channel (maximum depth 56 m), 7 = Fårö sill (115 m), 8 = Southern Åland sill (70 m), 9 = Middle Åland sill (70 m), 10 = Södra Kvarken sill (100 m), 11 = Norra Kvarken sill (25 m). The major deeps are indicated in blue: 12 = Bornholm deep (105 m), 13 = Gotland deep (249 m), 14 = Fårö deep (205 m), 15 = Norrköping deep (205 m), 16 = Landsort deep (459 m), 17 = Lågskär deep (220 m), 18 = Åland deep (301 m), 19 = Ulvö deep (293 m). The major offshore stone reefs and sand banks are indicated in yellow: 20 = Fladen, 21 = Lilla Middelgrund, 22 = Morups bank, 23 = Stora Middelgrund, 24 = Kriegers flak, 25 = Adlergrund, 26 = Odra bank, 27 = Słupsk bank, 28 = Södra Midsjö bank, 29 = Norra Midsjö bank, 30 = Hoburgs bank. The major islands are indicated in green: 31 = Fyn (2984 km²), 32 = Sjælland (7031 km²), 33 = Lolland (1243 km²), 34 = Falster (514 km²), 35 = Rügen (926 km²), 36 = Bornholm (588 km²), 37 = Öland (1347 km²), 38 = Gotland (3184 km²), 39 = Saaremaa (2922 km²), 40 = Hiiumaa (1023 km²), 41 = Åland (1552 km²). Figure: © Pauline Snoeijs-Leijonmalm

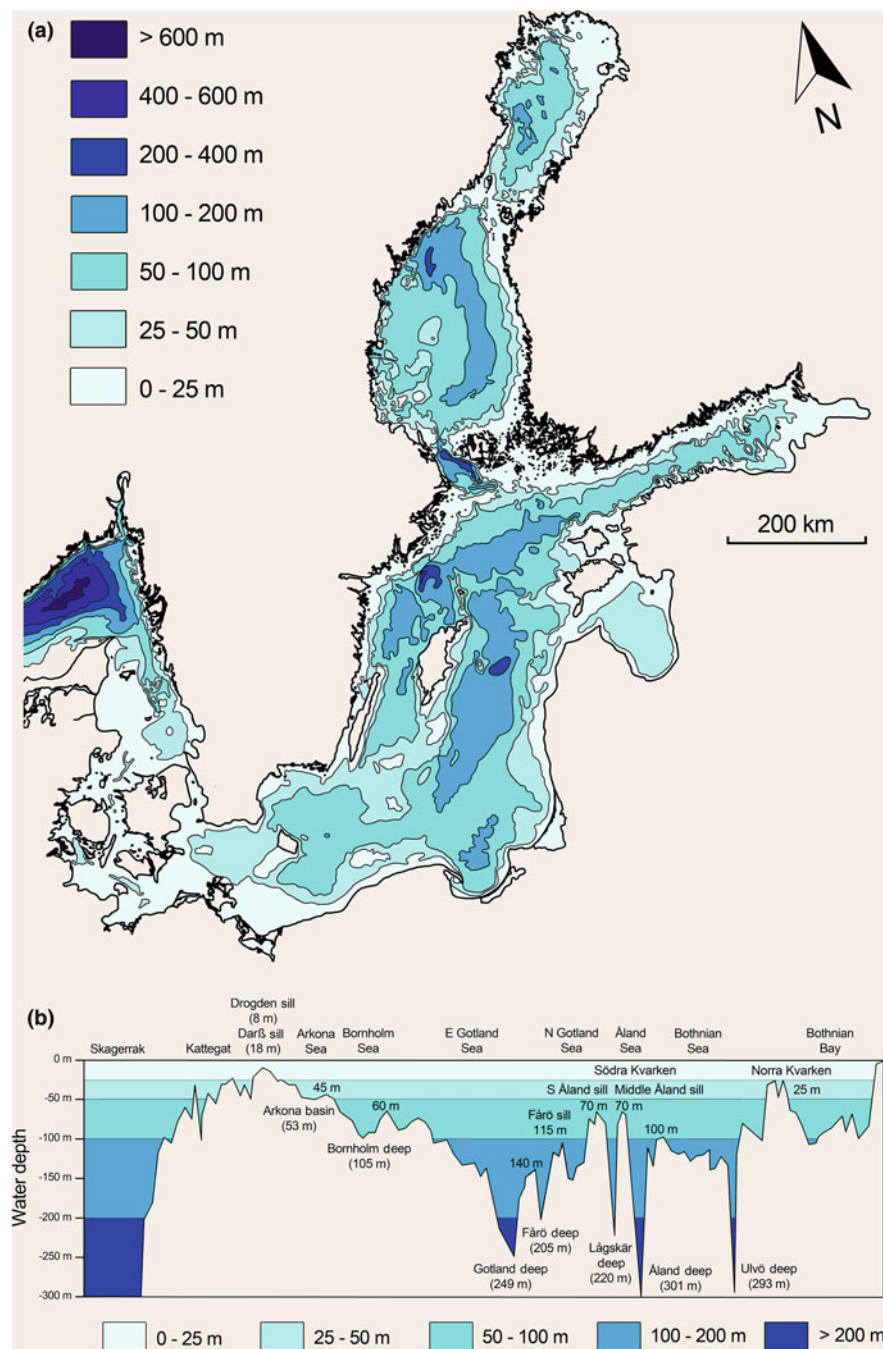


Fig. 2.3 Water depth in the Baltic Sea Area. (a) Geographical map showing water depth. (b) Schematic cross-section from the Skagerrak to the Bothnian Bay via the Eastern Gotland Sea, showing the maximum depths of the major deeps and the minimum depths of the major sills (cf. Fig. 2.2). Figure (a) modified from Bernes (2005), (b) modified from Sjöberg (1992)

boundaries, *e.g.* for species distributions, between the the Belt Sea (with strong marine influence) and the low-salinity brackish water of the Baltic Sea (cf. Sect. 4.2.2).

Altogether, the Baltic Sea covers a water area of $\sim 369,000 \text{ km}^2$ with an average volume of $\sim 21,000 \text{ km}^3$

(Table 2.2). The largest part of the Baltic Sea, containing 64 % of its total water volume, is called the “Baltic Sea proper” (Fig. 2.2), which is sometimes also referred to as the “Baltic Proper”. The Gulfs of Riga, Finland and Bothnia are connected to the northern and northeastern Baltic Sea proper.

Table 2.2 Dimensions of the Baltic Sea Area (Baltic Sea and the transition zone) and the Skagerrak (North Sea). *The Belt Sea includes the Danish straits (Storebælt 7,765 km², Lillebælt 2,942 km², Öresund 2,278 km²) and the southern Belt Sea, also known as the “Western Baltic Sea” (Kiel Bay 3,382 km² and Mecklenburg Bay 4,636 km²). **The Baltic Sea *sensu stricto* does not include the transition zone. Surface area data from HELCOM (2012), water depth data from Leppäranta and Myrberg (2009), and surface-water salinity data (shown as long-term averages 1970–2008) from Andersen et al. (2015)

Subregion	Surface area (km ²)	Water volume (km ³)	Average depth (m)	Maximum depth (m)	Surface-water salinity
Skagerrak (North Sea)	33,400	7,281	218	700	34–35
Transition zone	43,105	801	19	109	9.6–30.2
Kattegat	22,102	508	23	130	12.2–30.2
Belt Sea*	21,003	293	14	81	9.6–22.9
Baltic Sea proper	208,253	13,313	64	459	5.0–11.3
Arkona Sea	16,502	380	23	53	7.6–11.3
Bornholm Sea	41,970	1,931	46	105	4.3–8.1
Gdańsk Bay	5,806	331	57	114	5.0–7.5
Eastern Gotland Sea	74,985	5,774	77	249	
Western Gotland Sea	27,876	1,979	71	459	
Northern Gotland Sea	41,114	2,919	71	150	
Gulf of Riga	18,796	432	23	51	4.1–6.2
Gulf of Finland	29,901	1,106	37	123	1.2–5.6
Gulf of Bothnia	112,384	6,106	54	301	1.8–6.6
Åland Sea	4,433	332	75	301	3.8–6.6
Archipelago Sea	11,077	210	19	104	
Bothnian Sea	63,650	4,201	66	293	
Bothnian Bay	33,224	1,362	41	146	1.8–3.9
Baltic Sea**	369,334	20,958	57	459	1.8–11.3

2.2.3 A semi-enclosed sea

The Baltic Sea is located on the Eurasian continental shelf and is almost completely enclosed by landmasses made up by the European mainland and the Scandinavian peninsula (Fig. 2.1). It is artificially linked to the North Sea by the Kiel Canal (via the Belt Sea), to the northern Kattegat by the Göta Canal through Sweden, to the White Sea (Beloye Morye) by the White Sea Canal via Lake Onega and the Volga-Baltic Waterway, and even to the Ponto-Caspian region (the Black, Azov and Caspian Seas) via a ramified network of inland waterways including human-made canals. Water exchange through these waterways is negligible, but the building of the canals has provided possibilities for non-indigenous species to reach the Baltic Sea (*cf.* Sect. 5.3.2).

2.2.4 A shallow sea

The average depth of the Baltic Sea is only ~57 m (Fig. 2.3, Table 2.2). The deepest place is the Landsort deep with a recorded depth of 459 m (Figs. 2.2 and 2.3). Other deep sites

are the Gotland deep in the Eastern Gotland Sea (249 m), the Åland deep in the Åland Sea (301 m) and the Ulvö deep in the Bothnian Sea (293 m).

The major sills within the Baltic Sea are the Słupsk sill (56 m) in the southern Baltic Sea proper and those associated with Södra Kvarken between the Åland Sea and the Bothnian Sea (70 m) and Norra Kvarken between the Bothnian Sea and the Bothnian Bay (25 m). There is no threshold between the Baltic Sea proper and the Gulf of Finland, while the Gulf of Riga is enclosed by large islands and shallow waters. The Gulf of Riga (average depth 23 m) is shallower than the Gulf of Finland (37 m), which is shallower than the Bothnian Sea (66 m) and the Bothnian Bay (41 m).

The largest islands in the Baltic Sea are, from north to south: Åland (Finland), Hiiumaa and Saaremaa (Estonia), Gotland and Öland (Sweden), Bornholm, Lolland, Falster, Fyn and Sjælland (Denmark) and Rügen (Germany).

In the Baltic Sea and the shallow Kattegat, there are also a number of offshore stone reefs and sand banks. These shallow areas, with water depths of 5–20 m and surrounded by deeper waters, are not directly affected by terrestrial

runoff, and if undisturbed by trawling they host a pristine macroalgal vegetation with associated fauna. The offshore stone reefs and sand banks are important foraging areas for waterbirds and act as refuges for endangered coastal species. The largest banks inside the Baltic Sea are Kriegers flak, Adlergrund, Odra bank, Slupsk bank, Södra Midsjö bank, Norra Midsjö bank and Hoburgs bank (Fig. 2.2). In the Kattegat, the largest banks are Fladen, Lilla Middelgrund and Stora Middelgrund.

2.2.5 A large drainage area

The drainage area (syn. drainage basin or catchment area) of the Baltic Sea is located between latitudes 49 °N and 69 °N and between longitudes 8 °E and 37 °E (Fig. 2.4). Its climate varies from mild in the south to subarctic in the north and from oceanic in the west to continental in the east. The size of the Baltic Sea drainage area is 1.7 million km², which is more than four times larger than the water surface area. Compared with most other seas, the Baltic Sea water surface area is large in relation to its volume. Water renewal time (turnover time) in the Baltic Sea is slow, around 30 years in the southern part and 40 years in the northern part, because of the shallow sills in the southwest. Altogether, this means that the Baltic Sea is heavily influenced by runoff from the surrounding land-masses and by anthropogenic activities on land.

The Baltic Sea is strongly affected by activities in the nine countries that are directly bordering it: Sweden, Finland, Russia, Estonia, Latvia, Lithuania, Poland, Germany and Denmark (Fig. 2.5a). In six of these countries, most of the land surfaces belong to the drainage area of the Baltic Sea, but those of Denmark, Germany and the Russia only partially. The nine Baltic Sea countries, together with the European Union, co-operate in monitoring the environmental status of their common sea and jointly devise measures for its adequate management. The international governmental body coordinating these activities is the Baltic Marine Environment Protection Commission (Helsinki Commission, HELCOM, cf. Sect. 17.8.4). Five more countries, Norway, Belarus, Ukraine, Slovakia and the Czech Republic, are partially located within the drainage area, and runoff from these countries affects the Baltic Sea after passing through one of the riparian countries.

2.2.6 A large human population in the drainage area

Altogether, ~85 million people live in the Baltic drainage area, almost 18 % of them within 10 km of the coast (HELCOM 2010a). The population is unevenly distributed, with roughly 10 % in the northern part, 15 % in the eastern

part and 75 % in the southern part (Fig. 2.5b). The large human population around the Baltic Sea is served by some of the busiest shipping routes in the world. Around 2,000 large vessels are normally at sea at any one time, including cargo ships (>50 % of the large vessels in 2009), oil tankers (~18 %) and passenger ferries (~11 %) (HELCOM 2010b).

According to the Oslo/Paris Convention for the Protection of the Marine Environment of the North-East Atlantic (OSPAR 1998, cf. Box 14.1), the eutrophication of marine waters refers to the enrichment of water by nutrients, which causes an accelerated growth of algae and plants that produce an undesirable disturbance in the balance of organisms present in the water and to the quality of the water concerned. Hence, eutrophication includes the undesirable effects resulting from anthropogenic nutrient inputs. In the southern part of the drainage area, ~65 % of the land area is used for agriculture (Fig. 2.5c), which is the major cause of eutrophication of the Baltic Sea. In the northern part, agriculture uses less than 5 % of the land area and over 80 % is covered by boreal forest.

The southern part of the drainage area of the Baltic Sea also contains the largest number of “hotspots” (HELCOM 2010a), *i.e.* sites known for large discharges of hazardous substances (toxic, persistent and bioaccumulating compounds), which cause chemical pollution of the Baltic Sea (Fig. 2.5d). Even though a large number of these hotspots have disappeared during the last decades, the Baltic Sea water still contains high amounts of contaminants, one of the main reasons being the long water renewal time.

2.2.7 Patterns in sedimentation

The bedrock in the Baltic basin is superimposed by gradually younger sediments (Fig. 2.6, Table 2.3). On top of the bedrock, tills (unsorted sediments deposited during the ice ages) fill up its troughs and other depressions. Glaciofluvial deposits of sand, gravel and stones (eskers and river deltas) are occasionally found as remnants of glaciofluvial systems. On top of the tills and coarser glaciofluvial deposits, glacial clays drape the bottom topography, hence glacial and post-glacial deposits can be well over 100 m thick. The youngest sediment consists of post-glacial mud, a fine-grained, organic-rich material with a high capacity to bind various types of chemical pollutants. The mean annual sediment accumulation rate in the open Baltic Sea is ~0.5–2 mm, but there is considerable local variation.

The seabed can be subdivided into two sedimentation types: non-depositional, where erosion or transportation of sediment occurs, and depositional, where sediments accumulate. In-between the depositional and non-depositional bottoms there are areas where sediments are

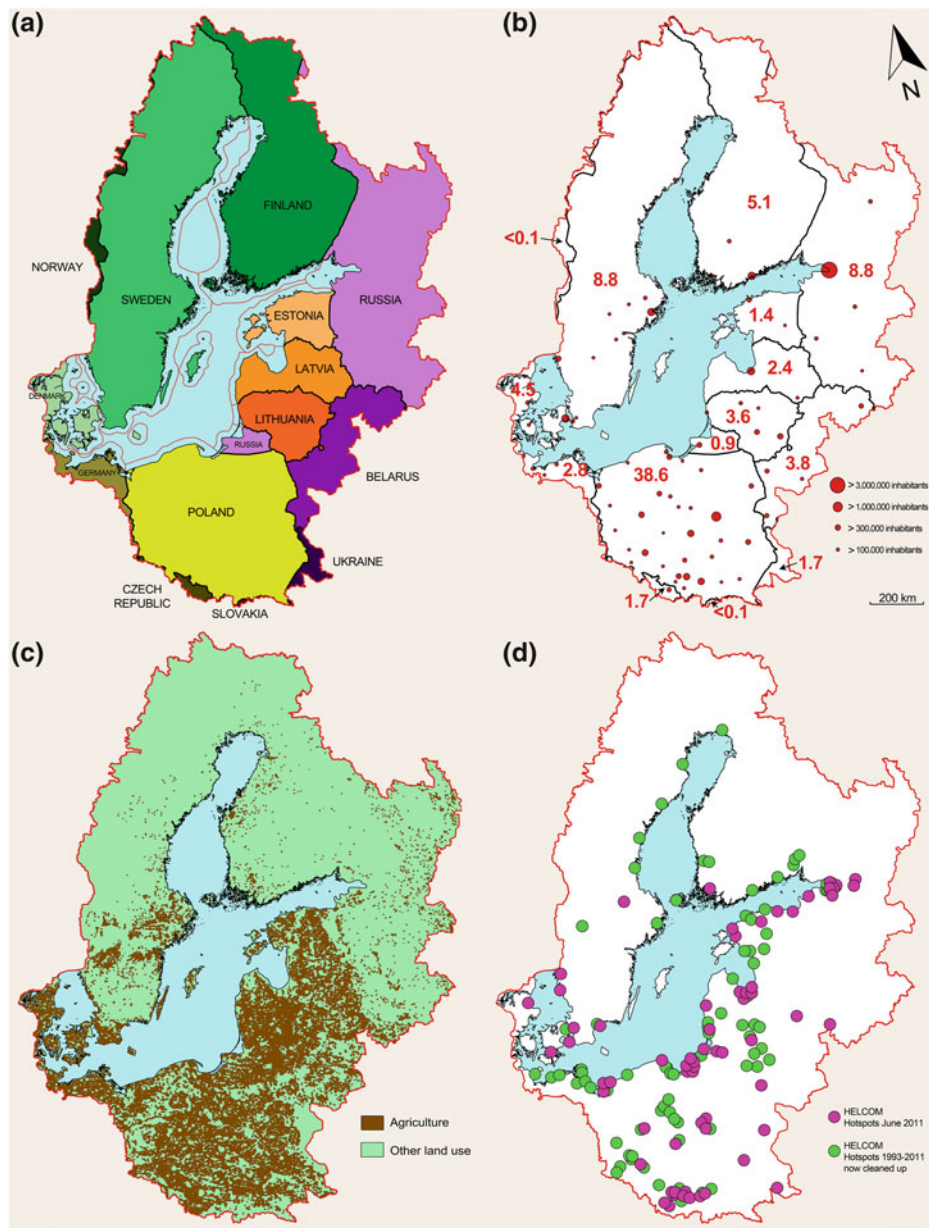


Fig. 2.5 Maps of the drainage area of the Baltic Sea Area, showing (a) The drainage areas of the nine countries with coastlines on the Baltic Sea with their territorial waters (thin red lines) and the five additional countries with runoff to the Baltic Sea. (b) The size of the population living in each of the 14 countries within the drainage area in millions of people, and the locations of the cities with more than 100,000 inhabitants (for city names, see Fig. 2.4). (c) Agricultural land use. (d) HELCOM hotspots in June 2011 and previous HELCOM hotspots that were cleaned up by June 2011. Figure (a) © Pauline Snoeijs-Leijonmalm, (b) based on population data in Hannerz and Destouni (2006), (c) modified from Bernes (2005), (d) modified from HELCOM (2010a), updated with data from the HELCOM web site (<http://www.helcom.fi>) until June 2011

more dynamic, occasionally accumulating, but they can be resuspended and transported easily. Resuspension of sediments occurs naturally as a result of wave action, currents and mass-movements affecting steep slopes, and as a result of land uplift or land subduction. However, it is also induced by anthropogenic activities such as trawling and dredging.

The boundaries between deposition and non-deposition on the seabed are not constant, especially in the northern part of the Baltic Sea where on-going land uplift shifts the baseline of erosion and new bottom areas emerge above the wave base. About half the Baltic Sea bottoms are of the non-depositional type. Erosion results in bottoms consisting of exposed bedrock, hard coarse deposits such as gravel and

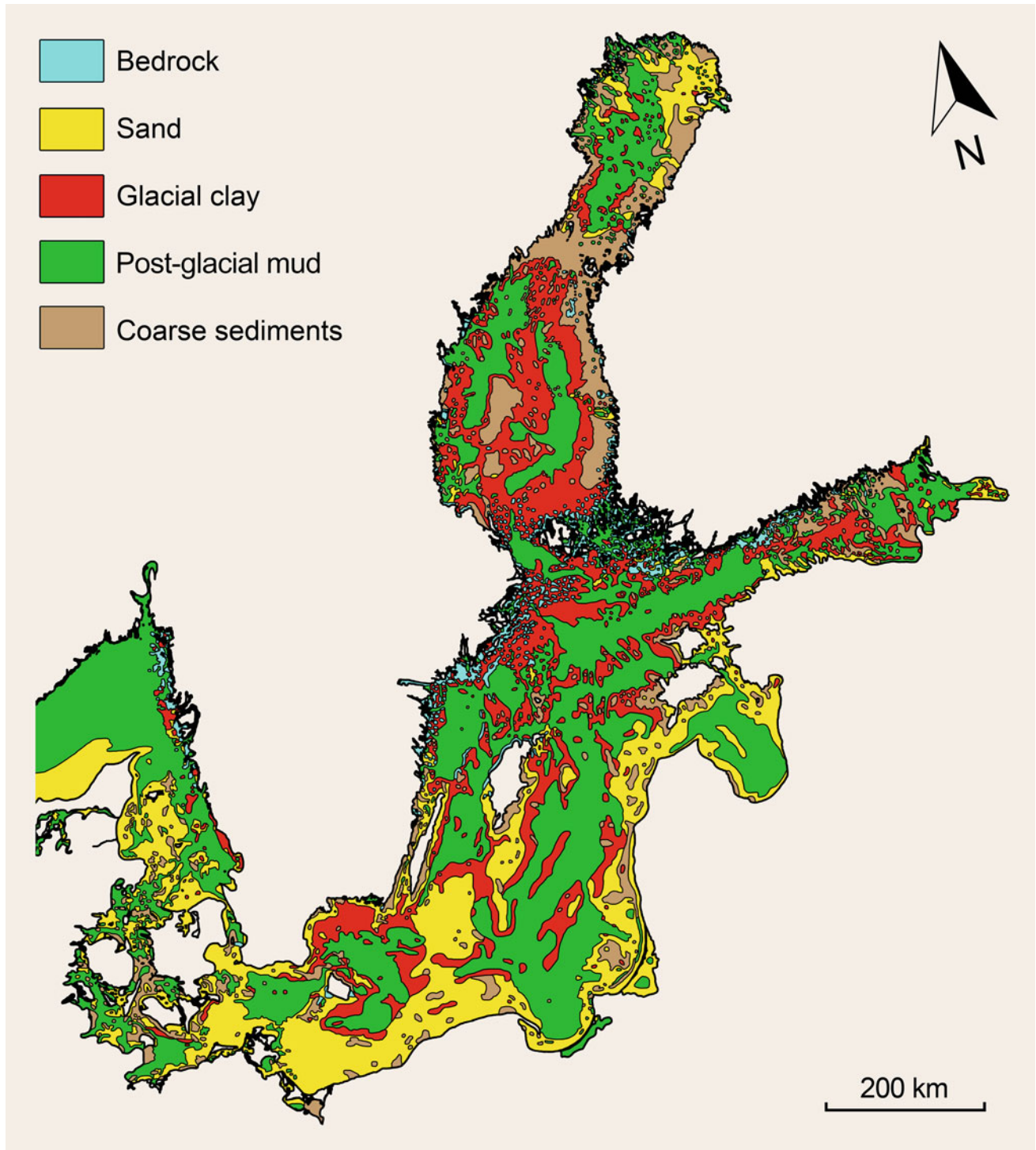


Fig. 2.6 Map of the Baltic Sea Area, showing its bottom sediment composition. Figure modified from Al-Hamdani and Reker (2007)

stones or resistant glacial clays. This implies that sea bottom does not necessarily consist of recent sediment but can have deposits with an origin of thousands of years.

Sediment transport in the Baltic Sea takes place in both vertical and lateral directions and roughly follows the water

depth contour, resulting in an accumulation of sediment beneath the halocline (compare Figs. 2.3a and 2.6). Post-glacial mud accumulates in the deeper areas, which function as the so-called “depocentres”, but can also accumulate in sheltered coastal embayments. Glacial clay is

Table 2.3 Simplified grain size table after Wentworth (1922). For sand and silt each scale value is half the size of the scale value above. The logarithmic phi scale (Krumbein 1934) is a modification that allows grain size data to be expressed in units of equal value for the purpose of statistical analyses and graphical plotting.

Substrate type	Wentworth size class	Lower grain size limit (mm)	Phi (Φ)
Gravel	Boulder	256	−8.0
	Cobble	64	−6.0
	Pebble	4	−2.0
	Granule	2.00	−1.0
Sand	Very coarse sand	1.00	0.0
	Coarse sand	0.50 (1/2)	1.0
	Medium sand	0.25 (1/4)	2.0
	Fine sand	0.125 (1/8)	3.0
	Very fine sand	0.0625 (1/16)	4.0
Silt	Coarse silt	0.031 (1/32)	5.0
	Medium silt	0.0156 (1/64)	6.0
	Fine silt	0.0078 (1/128)	7.0
	Very fine silt	0.0039 (1/256)	8.0
Mud	Clay	0.00006	14.0

found at the fringes of these deeper areas where erosion or transportation of sediments prevail. Sand and coarse sediments occur in shallow areas, mainly along the coasts. Sand dominates in the Kattegat, the southern and eastern Baltic Sea proper and the northeastern corner of the Bothnian Bay (Fig. 2.7). Coarse sediment is a generic name for gravel of different size (Table 2.3), which was deposited by ice sheets as till or is of glaciofluvial origin. These coarse deposits dominate along the rocky coasts of Sweden and Finland.

2.2.8 Rocky shores and sandy beaches

The Baltic Sea coasts consist of many different coast types formed along varying time scales and by various factors, such as tectonic movements, glacial erosion, glacial deposition and on-going wind- and coastal processes. According to a simplified system, the entire coast of the Baltic Sea Area can be divided into five coast types (Fig. 2.8). While the southern and southeastern Baltic Sea coast is dominated by open low sandy beaches or lagoon and bodden coasts (Fig. 2.7), the northwestern part is dominated by rocky archipelago coasts (Fig. 2.9). Fjord and klint coasts are rare.

The grain size of rocks and sediments is an important environmental driver, *e.g.* it determines the type of macrophyte vegetation and its accompanying fauna (*cf.* Sect. 11.1). The following grain size limits are commonly used: clay/mud has a grain size <3.9 μm , silt 3.9–63 μm , sand 63–2,000 μm , gravel >2 mm. Gravel is subdivided into different sizes of stones: granules 2–4 mm, pebbles 0.4–6.4 cm, cobbles 6.4–25.6 cm and boulders >25.6 cm (Table 2.3).

2.2.9 Fjords and fjärds

High fjord-like coasts are in the Baltic Sea Area found only in one small area in the western Bothnian Sea (the Höga Kusten area). The fjord-like inlets here mainly consist of water-filled valleys in exposed bedrock with no or minor coverage of glacial deposits. This is different from the “true” fjords of the type found in Norway, Greenland, Alaska and Chile, which are long, narrow inlets with steep walls formed by glacial erosion. Fjords typically have deep basins in the inner part due to glacial erosion, and since the glacier deposited gravel and sand where it met the sea, a threshold partly separates the fjord from the sea. There are no fjords of this type in the Baltic Sea Area; the closest one is the ~25 km long and 1–3 km wide Gullmarsfjorden in Bohuslän on the Swedish Skagerrak coast, with a threshold of ~40 m and a maximum depth of ~120 m in the inner part. The estuary of the river Ångermanälven in the Höga Kusten area is a fjord-like estuary. With its length of ~35 km, width of 1–3 km, threshold of ~20 m and maximum depth of ~100 m, it has a structure similar to a true fjord, but with less steep walls.

However, along the rocky archipelago coasts of Sweden and Finland there are many “fjärds”, which are drowned shallow glacial valleys that form large inlets of the Baltic Sea. For example, Bråviken on the Swedish coast of the Baltic Sea proper near Norrköping is over 50 km long. Fjärds are broader and shallower than fjords and lack steep walls. Eroded local materials were deposited into the fjärds, and after the last glaciation they were filled with seawater during the eustatic sea level rise. Contrary to fjords, fjärds may contain mud flats,



Fig. 2.7 Beaches of the southern and eastern Baltic Sea. **(a, b)** Views from the air. **(c)** Sandy beach with dispersed boulders. **(d)** Stony beach below a chalk cliff. **(e)** Sandy beach with empty shells and collected driftwood. **(f)** Construction for protection against beach erosion. Photo: **(a, b, d–f)** © Hendrik Schubert, **(c)** © Hans Kautsky

salt marshes and flood plains. The Swedish word “fjärd” is also used for *e.g.* broad open water areas between archipelago islands away from the mainland because the word simply means “fare way” (like the Norwegian word “fjord”).

A third type of inlets of glacial origin is found in the Belt Sea area, and to complicate the matter, they are also called “fjords” in Danish and “Förden” in German. However, the glacial mechanics were different from those

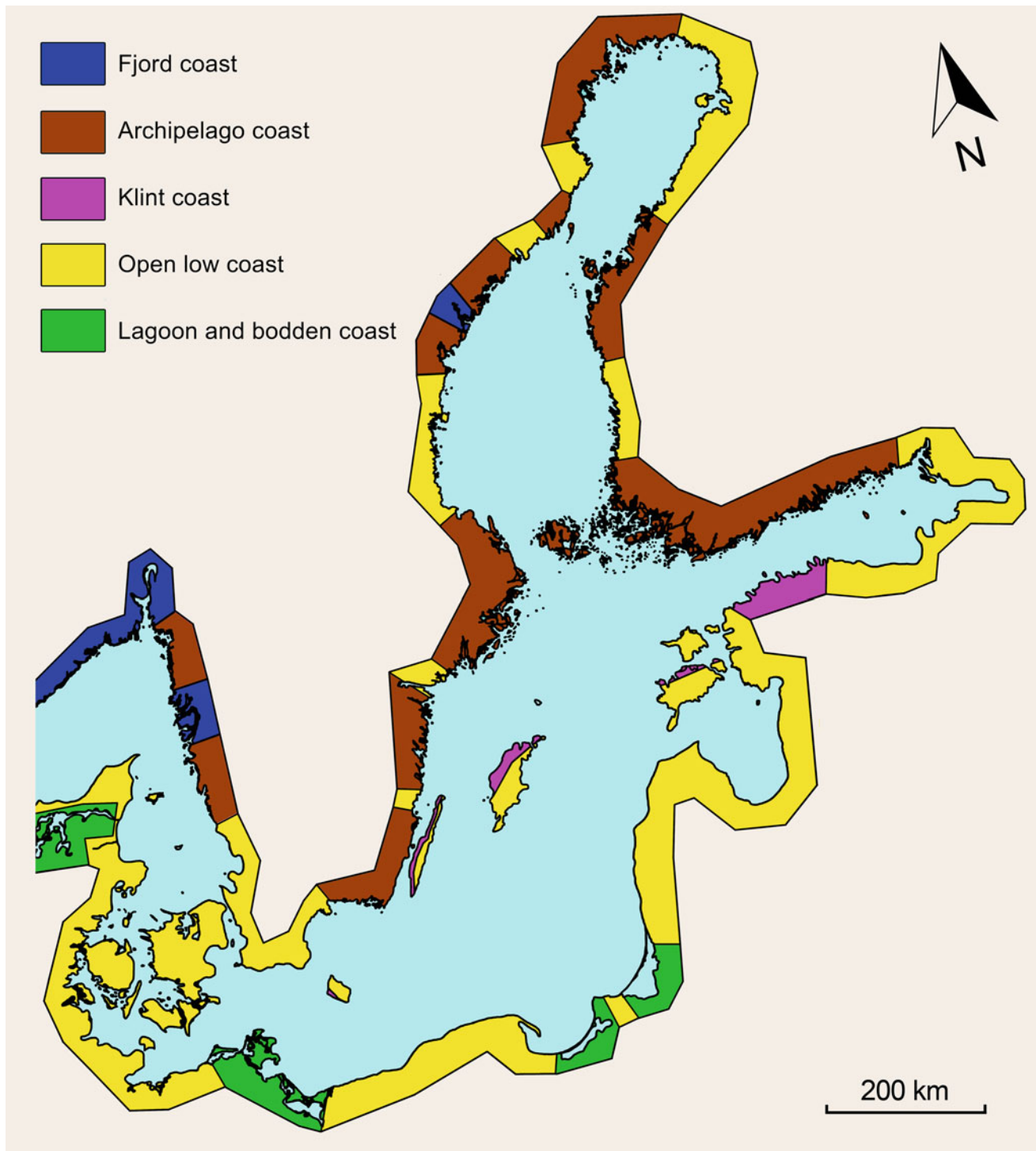


Fig. 2.8 Map of the Baltic Sea Area, showing the large-scale pattern of where different coast types dominate. Note that this only refers to the coastline (not inland geology) and that when zooming in to certain areas there is often a mosaic of different coast types. Figure based on coastal data in Boedeker and Knapp (1995), Lampe (1995) and Soesoo and Middel (2007). Figure: © Pauline Snoeijs-Leijonmalm

of Norwegian fjords and Swedish and Finnish fjärds. In the Belt Sea area, the movement of the Weichselian ice sheet's edge carved out depressions in the land, which were later filled with seawater during the eustatic sea level rise.

2.2.10 Archipelago coasts and klint coasts

Archipelago coasts, rocky coasts with numerous islands, skerries (small islands) and fjärds prevail in the northwestern part of the Baltic Sea Area and the eastern Kattegat



Fig. 2.9 Rocky coasts and archipelagos of the Baltic Sea. (a) Bornholm, southern Baltic Sea proper. (b) Ekenäs archipelago, Gulf of Finland. (c, d) Kråkelund, Baltic Sea proper. (e) Inner Stockholm archipelago, northern Baltic Sea proper. (f) Sheltered bay at Oskarshamn, Baltic Sea proper. Photo: © Pauline Snoeijs-Leijonmalm

(Figs. 2.8 and 2.9). The largest archipelago areas are the Stockholm archipelago (>30,000 islands and skerries) and the Archipelago Sea (>50,000 islands and skerries) between Åland and the Finnish mainland. Archipelago coasts can be

formed by different processes, *e.g.* large-scale tectonic movements in the bedrock and glacial erosion creating fissure valley landscapes where the fissures are water-filled, creating abundant islands.

Larger areas with klint coasts, with up to 50 m high cliffs of stratified sedimentary limestone bedrock, occur in the southern Gulf of Finland, on the Estonian coast and on the islands of Gotland and Bornholm (Fig. 2.8). Klint coasts are created when waves undercut a cliff, which results in rock-fall, and the talus masses (accumulations of broken rock debris) are subsequently washed away by waves. The klint on the coasts of Öland and northern Estonia belongs to the “Baltic klint ridge”, while the klint on the coasts of Gotland and Saaremaa belongs to the “Silurian Klint” ridge. These ridges continue on the seafloor across the Baltic Sea proper.

2.2.11 Low open coasts and boddens coasts

Low and open coasts, consisting mainly of sand, prevail in the southeastern Baltic Sea (Figs. 2.7 and 2.8). The shore profile is usually low because the continent is old and mature compared to the Scandinavian mountain range, but chalk cliffs occur at some places (Fig. 2.7d). The sand, derived from the weathering of bedrock, was carried to the Baltic basin by fluvial transport from the continent for several hundred thousand years, with only relatively short breaks of glacial coverage. Sandy beaches are easily eroded, by both waves and winds, which creates coasts featuring sand dunes and spits. However, in the southern Baltic Sea there is also a considerable redeposition of sediments as a result of a transgressive water level.

Areas with low coasts interrupted by shallow coastal lagoons (boddens) and with bottoms covered by muddy sediments, often forming chains, occur in the southern Baltic Sea, *e.g.* the Darß-Zingster Boddenkette in northern Germany. Boddens have only narrow connections with the sea as they are semi-enclosed by peninsulas and are strongly affected by freshwater runoff. Most of the peninsulas were originally formed as till ridges at the ice sheet margins, and have later been modified by wind erosion and coastal processes in areas subjected to land subsidence. Some of the peninsulas are pure sand deposits.

2.3 Hydrography

2.3.1 Hydrography and the distribution of organisms

Species distributions in the Baltic Sea Area strongly depend on water circulation. The Baltic Sea basically lacks an intertidal zone because there are no regular daily water level fluctuations between high and low water level. However, irregular water level changes make the upper littoral of the Baltic Sea a highly dynamic zone with periods of total desiccation. This has forced the perennial Atlantic intertidal species that can

penetrate into the brackish Baltic Sea to adapt to a permanently submerged life. Freshwater transports eroded materials and nutrients from the drainage area to the sea. This fertilises the sea and enhances the growth of primary producers.

The weak water exchange with the ocean promotes oxygen deficiency in the deeper areas of the Baltic Sea. This generates oxygen stress for organisms and affects animal life as well as the functional diversity of bacteria. The net water budget and long water residence time in the Baltic Sea create a nearly stable salinity gradient, along which species are distributed according to their salinity optima and tolerances.

For many species in the Baltic Sea, a major prerequisite for life involves large inflows of saline water from the Kattegat (Box 2.1). When an inflow occurs, salinity stress for marine species living in the Baltic Sea decreases and oxygen stress for animals living in the deeper areas is alleviated, at least temporarily. The surface-water currents in the Baltic Sea affect the transport-related processes, such as the dispersal of organisms.

2.3.2 Currents

The average net surface-water circulation in the main sub-regions of the Baltic Sea is counter-clockwise (Fig. 2.10). These currents are induced by complex interactions between the Coriolis force, wind stress at the sea surface, sea level tilt, thermohaline gradient of water density, the (minimal) tidal forces, bottom topography and friction. The Baltic Sea surface-water currents are, on average, rather weak ($\sim 5 \text{ cm s}^{-1}$), but during storms, wind-driven currents can reach 50 cm s^{-1} and up to 100 cm s^{-1} in straits (Leppäranta and Myrberg 2009). The counter-clockwise surface-water circulation transports salt, heat, nutrients, contaminants, sediments, plankton and propagules of species (including non-native ones) and thus affects many aspects of the Baltic Sea ecosystem. For example, the circulation continuously shapes the large sandy beaches in the southeastern Baltic Sea.

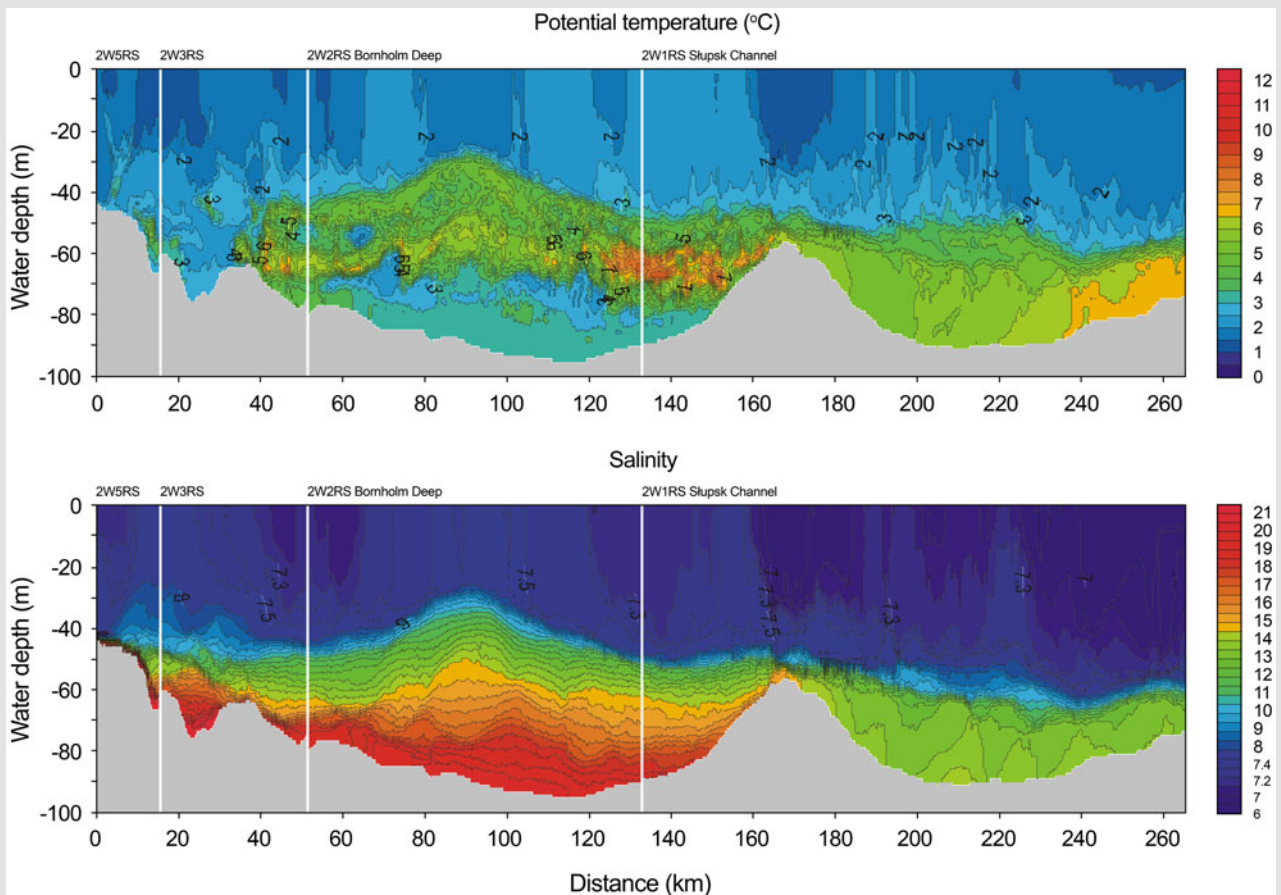
Zooplankton and fish species that live in deeper water can be transported from the Kattegat into the Baltic Sea with saltwater inflows. Their dispersal is thus mainly controlled by the baroclinic flow field and bottom topography (Box 2.1). Hydrodynamic drift modelling has shown that the potential dispersion of *e.g.* comb jellies follows the deep-water currents from the Bornholm Sea towards the north and the east of the Baltic Sea and is limited by topographic features and low advection velocities (Lehtiniemi et al. 2012). However, if such species are new invaders in the area (*cf.* Sect. 5.1), and the conditions for growth and reproduction are favourable in the Baltic Sea, they will be able to form stable populations despite the fact that most individuals are hampered by the hydrodynamics of the deep water.

Box 2.1: Major Baltic inflows (MBIs)

Kai Myrberg

Barotropic inflows

Since the mid-1970s, the frequency and intensity of barotropic MBIs (based on sea level differences between the Baltic Sea and the North Sea) have decreased (*cf.* Fig. 2.13a). They were completely missing between 1983 and 1993 (Lass and Matthäus 1996; Matthäus et al. 2008), which was the longest break of MBIs ever since the measurements started in 1890s. After another 10 years without a large MBI between 1993 and 2014, a strong event occurred in December 2014 (Mohrholz et al. 2015; Gräwe et al. 2015). Several theories have been proposed to explain the decrease of the frequency of MBIs, which are *e.g.* coupled to changing riverine runoff or meteorological patterns (*e.g.* Leppäranta and Myrberg 2009).



Box Fig. 2.1 Potential temperature and salinity on 16–18 February 2003, along the axis Arkona basin—Bornholmsgattet—Bornholm deep—Slupsk channel—Gdansk deep. Figure reprinted from Piechura and Beszczyńska-Möller (2004) with permission from Oceanologia (Institute of Oceanology PAN, Sopot, Poland)

Theories explaining the frequency of barotropic inflows: riverine runoff

From long-term observations it becomes evident that there is a good fit between the minimum deep-water salinity and the maximal riverine runoff if there is a time interval of six years between salinity and runoff measurements (Lau-niainen and Vihma 1990). From the dynamic viewpoint, there are two main mechanisms driven by the riverine runoff variability and counteracting the inflows (Matthäus and Schinke 1999). Firstly, the low-salinity outflowing water mixes in the surface layer with the more saline lower layer which penetrates into the Baltic Sea in the near-bottom layer of the sill areas. Increased water supply from rivers reduces the salinity of the outflowing water and strongly dilutes the inflowing waters. Secondly, an increase in the freshwater supply to the Baltic Sea produces larger outflow, reduces or impedes the inflow of saline water and causes unfavourable conditions for MBIs.

Theories explaining the frequency of barotropic inflows: meteorology

On a larger scale, strong wintertime westerly winds (associated with a high NAO-index) transport intensively moist air masses from the North Atlantic to Europe, resulting in increased precipitation in the Baltic Sea region, with lower evaporation and increased riverine runoff. Above-normal Baltic Sea levels occur frequently for long periods, which hampers saltwater inflows (Zorita and Laine 2000). Increases of stagnation periods can be due to the high salinity in the bottom waters. High bottom salinities in the 1950s and 1960s may have been caused by the MBI in 1951 (Meier et al. 2006), filling the Baltic deeps with highly saline water and making their replacement by later inflows difficult. There are indications of changes in the Baltic Sea local wind climatology. An anomalous west wind component at the Kap Arkona station was found between August and October during 1951–1990 for seasons without a MBI as compared with the corresponding years with MBIs. In the years without MBIs, the period with easterly winds is shortened. Such changes of local wind patterns may cause variations in long-term salinity patterns, which cannot be explained by accumulated freshwater inflow or by low-frequency variability of the zonal wind (Lass and Matthäus 1996). Lehmann and Post (2015) examined atmospheric circulation conditions necessary to force large volume changes (LVCs, total volume changes of the Baltic Sea of at least 100 km³). MBIs can be considered as subset of LVCs transporting additionally a large amount of salt into the Baltic Sea. An LVC is a necessary condition for a MBI, but an LVC alone is not sufficient. Lehmann and Post (2015) confirmed earlier conclusions about the importance of the pre-inflow period when prevailing easterly winds increase the Baltic Sea brackish water outflow, lower the mean sea level and hinder the inflow of Kattegat water through the Danish straits (Storebælt, Lillebælt and Öresund).

Baroclinic inflows

Warm baroclinic inflows into the Baltic Sea (based on water-density differences between the Baltic Sea and the North Sea) also occur. Such inflows regularly take place in late summer and autumn in the southern Baltic Sea Area, as shown by the mean long-term annual temperature cycles in the deep water of the Arkona, Bornholm and Gdańsk basins (Matthäus 2006). Inflows with exceptionally warm waters recently occurred, in 2002 and 2003 (Box Fig. 2.1). Two types of such inflows have been observed with specific dynamic mechanisms (Matthäus 2006). The first type is caused by heavy westerly gales which pass over the Darß and the Drogden sills. The second type is a long-lasting baroclinic inflow, which only passes over the Darß sill, caused by calm weather conditions over central Europe. In such a situation the inflows are driven by baroclinic pressure gradients, especially caused by horizontal salinity differences (Feistel et al. 2006). Warm baroclinic inflows can transport large volumes of exceptionally warm water into the deeper layers of the Gotland Sea. However, these inflows in fact import oxygen-deficient waters, although they seem to be important for ventilation of intermediate layers in the Eastern Gotland basin deep water through entrainment (Feistel et al. 2006). On the other hand, warm water inflows (baroclinic or barotropic) do transport less oxygen to the Baltic Sea than cold-water inflows, and higher temperatures increase the rate of oxygen consumption in the deep water and facilitate formation of hydrogen sulphide (Matthäus 2006).

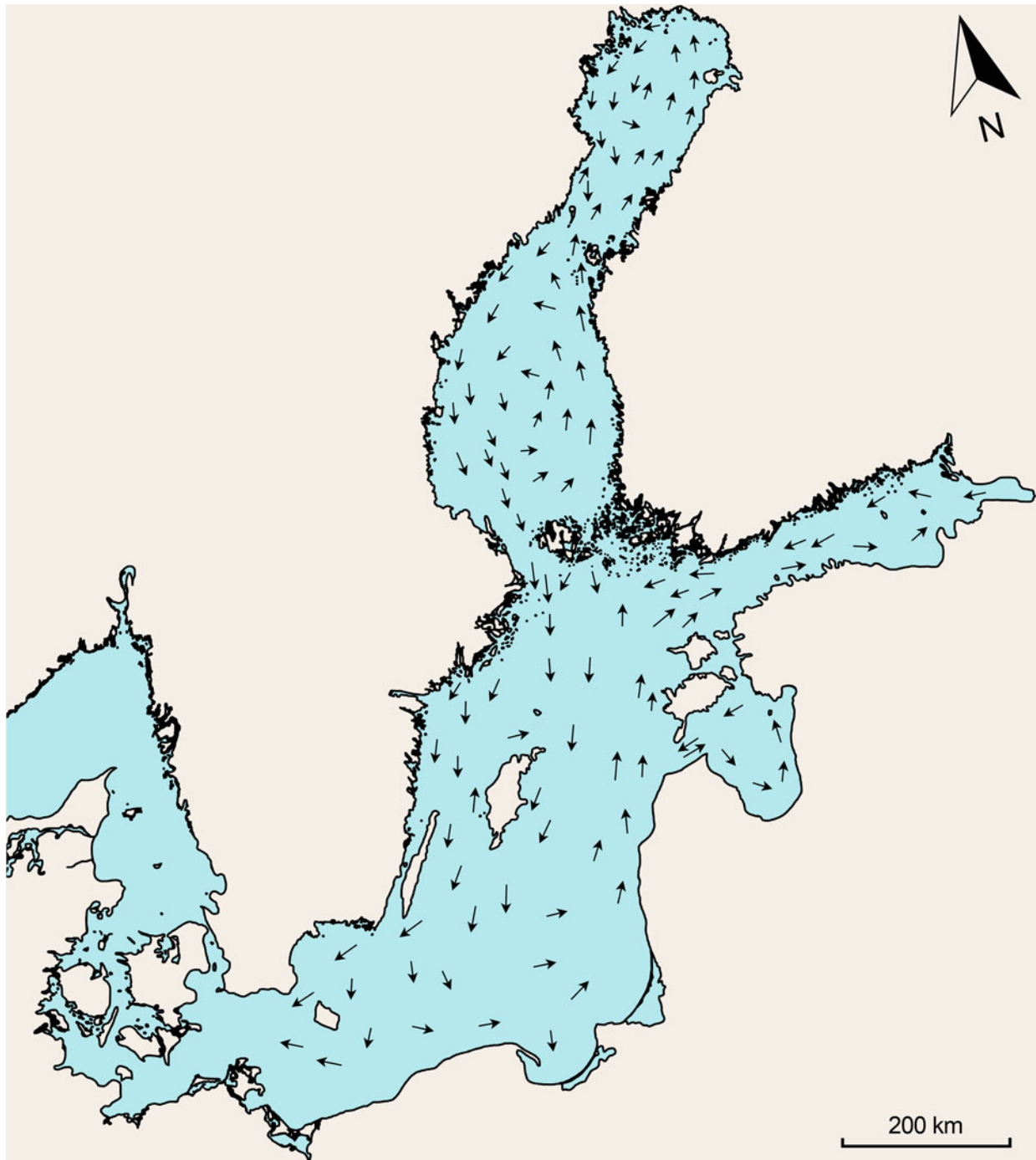


Fig. 2.10 Map of the Baltic Sea Area, showing the average (net) directions of the surface-water circulation based on measurements obtained from drifters deployed and followed by lightships. Figure modified from Leppäranta and Myrberg (2009)

2.3.3 A microtidal sea

The Baltic Sea is “microtidal”, which is defined as a tidal amplitude <2 m (Hayes 1979). In fact, the tidal amplitude in the Baltic Sea is much smaller than that. In the Belt Sea, the tidal amplitude in sea level is ~ 10 cm, over most of the Baltic Sea it is 2–5 cm and only from the eastern Gulf of

Finland amplitudes of >10 cm have been reported (Leppäranta and Myrberg 2009). These tidal changes are so small that the difference between high and low tide is basically undetectable anywhere in the Baltic Sea because they are masked by the much larger water level fluctuations caused by air-pressure changes and winds (Novotny et al. 2006).

Tides involve the rise and fall of the water level caused by the combined effects of the gravitational forces exerted by the moon and the sun and the rotation of the Earth. The subbasins of the Baltic Sea themselves are too small and shallow to have their own significant tides. Instead, the few cm of tidal amplitude in the southern Baltic Sea originate from the tidal waves coming in from the North Sea via the Skagerrak and the transition zone.

The North Sea tidal amplitude depends on counter-clockwise surface-water circulation (due to the Coriolis force) and distance from central amphidromic points (sites with zero amplitude in the open sea). Since the North Sea gets narrower southwards to the Strait of Dover, the tidal amplitude increases southward along the British coast to macrotidal (>4 m of tidal amplitude). When the water masses move northward again along the shallow sandy Belgian, Dutch and Danish coasts, the tidal waves are reduced by friction. Finally, when the circulating water reaches the Skagerrak, the tidal amplitude is only a few dm, and has decreased to <10 cm when it enters the Baltic Sea over the the Darß and Drogden sills.

2.3.4 Sea level changes

Sea level changes in the Baltic Sea are predominantly controlled by meteorological forcing. In any location, the water level is directly influenced by the local air pressure. High air pressure produces low water levels, and *vice versa*. Strong winds also affect the water level by pushing the water up against the coast or by pressing the water away from the coast depending on the wind direction. These weather-dependent sea level changes are largest when the combination of wind and air pressure results in storm surges, especially in shallow sea areas and near the coast, and can be up to 1–2 m (*cf.* Fig. 11.22b). Before the construction of the Neva Bay dam outside Sankt-Petersburg, extremes of up to 4 m in the easternmost Gulf of Finland were experienced.

As a consequence of the weather dependence, the irregular water level changes in the Baltic Sea can be fast and drastic, *e.g.* in some areas it can increase or decrease by more than 1 m over one day. For example, in winter there can suddenly be a 1-m thick air layer between the sea ice and the water in the Bothnian Bay (*cf.* Fig. 11.22c).

A low water level in one part of the Baltic Sea raises the level in another part, a phenomenon known as “seiche”. For example, in the northern Baltic Sea, autumn and winter water levels tend to be higher because of an increased frequency of conditions with low air pressure and strong westerly winds. In spring and summer, high air pressure and gentle winds dominate and water levels tend to be lower. However, the pattern is different in other parts of the Baltic Sea and seasonal patterns may be obscured by *e.g.* stochastic extreme weather conditions or large freshwater inflows.

The water level of the Baltic Sea is also subject to two slow long-term trends: (a) the land uplift in the north of the area (*cf.* Fig. 2.26b) which decreases the water level by up to ~1 cm per year and (b) the global sea level increase due to the melting of the Earth’s glaciers and expansion of the seawater volume by global warming. As a result of the latter trend, the average water level of the Baltic Sea has increased by 11 cm in the 80 years between 1890 and 1970 and by another 11 cm in the 45 years between 1970 and 2015 as measured by the Swedish Meteorological and Hydrological Institute (SMHI, <http://www.smhi.se>).

2.3.5 Coastal zonation

Oceanic coasts are typically subdivided into the supralittoral zone (splash zone), the eulittoral (intertidal) zone and the sublittoral (subtidal) zone that extends from below low tide to the edge of the continental shelf (Levinton 2010). Since tides are negligible in the Baltic Sea, and the whole sea is located on the continental shelf, its shores have only an epilittoral zone and a sublittoral zone. The epilittoral zone is not covered by water but receives saline sea spray, while the sublittoral zone is the submerged part of the shore.

Subdivisions of the sublittoral zone have been made according to regional conditions in different parts of the Baltic Sea Area, but none of these are applicable everywhere. The most widely used subdivision is that of a ~0.5–1 m wide “hydrolittoral” zone in-between the epilittoral zone and the permanently submerged sublittoral zone. The hydrolittoral zone is the part of the littoral that is most affected by the irregular sea level changes of the Baltic Sea and is defined as the zone that extends from the annual minimum water level up to the mean summertime water level (Du Rietz 1930). The hydrolittoral zone is subject to longer periods of desiccation (*cf.* Fig. 11.22a) and is inhabited by ephemeral algae and their accompanying fauna (Wærn 1952).

2.3.6 More than 200 rivers discharge into the Baltic Sea

Freshwater enters the Baltic Sea from over 200 rivers discharging along its coastline (Fig. 2.11). In the north, the runoff is usually largest in May from snow melt, and smallest in January and February when the air temperature is below 0 °C.

Twenty-eight major rivers together cover 80 % of the drainage area (Table 2.4). Twelve of these rivers are classified as “eutrophic”. In addition to nutrients from natural sources on land, they transport excess nitrogen and phosphorus from agricultural land to the sea. These 12 rivers mainly discharge into the Baltic Sea proper and the Gulf of

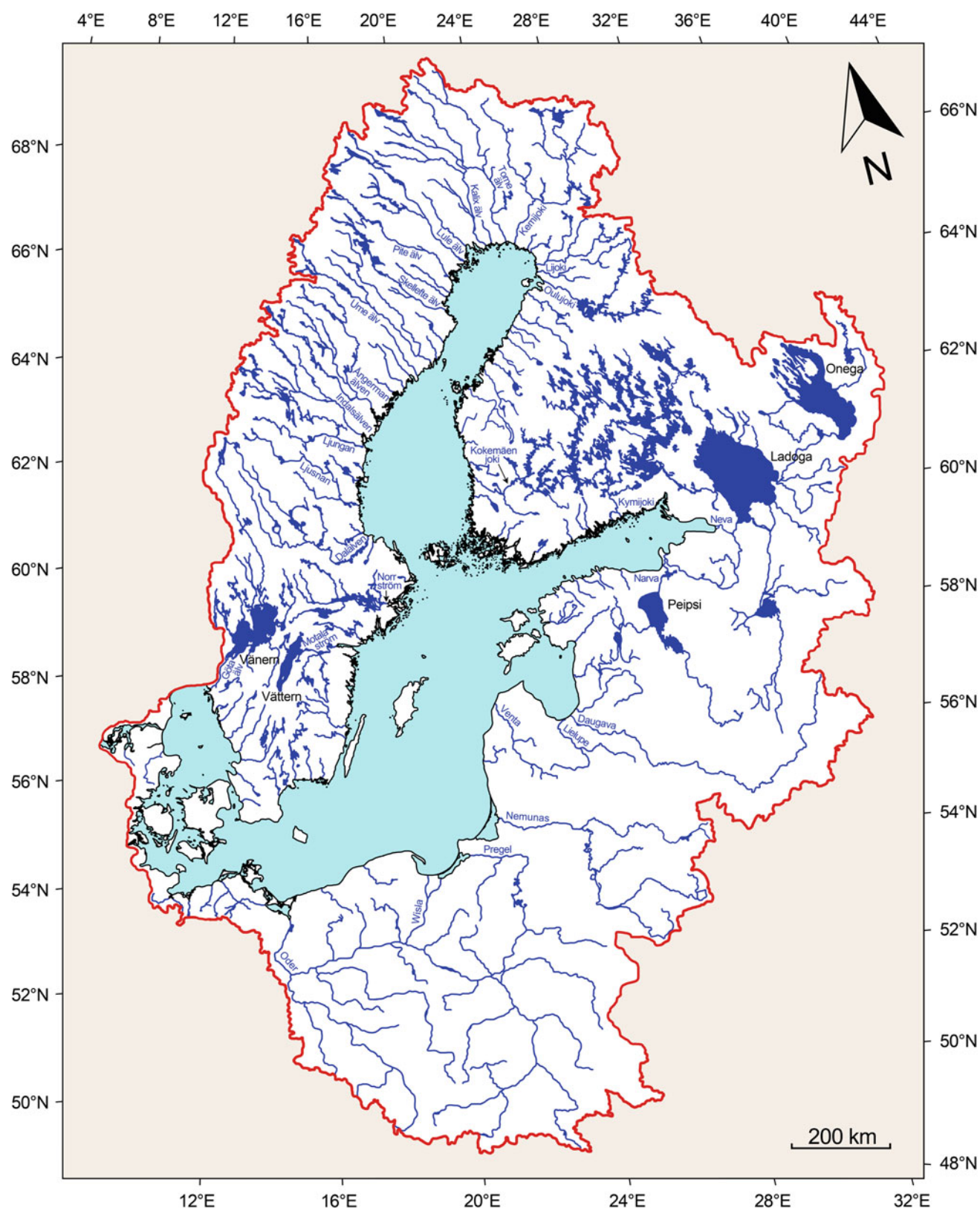


Fig. 2.11 Map of the Baltic Sea drainage area, showing the major lakes and rivers discharging into the Baltic Sea. Drainage area outline according to HELCOM (2010a). Figure: © Pauline Snoeijs-Leijonmalm

Table 2.4 List of the 28 major rivers discharging into the Baltic Sea, together covering 80 % of the drainage area, the Baltic Sea subregion to which they discharge, their drainage areas, land use in the drainage area and nutrients discharged. For the locations of these rivers, see Fig. 2.11. Monthly observations of river nutrient data (Si, N and P) and hydrological data based on measurements accessed from the major databases around the Baltic Sea with the decision support system Baltic Nest (<http://nest.su.se>, Wulff et al. 2013). These measurements represent river mouth data. The nutrient data were discharge-weighted, and averaged using monthly data from 1980 to 2000. Data from Humborg et al. (2007)

River	Subregion	Drainage area (km ²)	Forest (%)	Agriculture (%)	Bare, water, wetlands (%)	Discharge (km ³ year ⁻¹)	Nitrogen (tonnes year ⁻¹)	Phosphorus (tonnes year ⁻¹)	Silicate (tonnes year ⁻¹)
Eutrophic agricultural									
Neva	Gulf of Finland	285,835	68	15	17	80	56,996	3,707	17,927
Wisła	Baltic Sea proper	192,899	33	64	3	34	123,021	6,126	111,851
Odra	Baltic Sea proper	117,589	33	64	3	17	73,754	6,305	58,257
Nemunas	Baltic Sea proper	92,104	37	61	2	16	56,261	1,360	31,423
Daugava	Gulf of Riga	85,853	46	51	3	21	39,112	1,301	35,321
Narva	Gulf of Finland	56,665	40	52	8	14	9,367	507	19,549
Göta älv	Kattegat	48,214	67	13	20	18	15,098	328	8,586
Norrström	Baltic Sea proper	22,534	64	22	14	5	3,761	183	3,363
Lielupe	Gulf of Riga	17,835	28	70	2	3	18,436	327	6,696
Motala ström	Baltic Sea proper	15,544	60	20	20	3	2,535	105	2,414
Pregel	Baltic Sea proper	14,749	22	75	3	3	5,301	140	6,648
Venta	Baltic Sea proper	11,597	37	62	1	3	7,816	122	5,617
Regulated boreal									
Kemijoki	Bothnian Bay	50,918	84	3	13	19	6,989	401	55,486
Kymijoki	Gulf of Finland	36,522	78	3	19	10	5,683	215	10,255
Ångermanälven	Bothnian Sea	31,421	85	3	12	17	4,727	209	23,316
Dalälven	Bothnian Sea	28,873	86	5	9	12	5,319	221	23,432
Ume älv	Bothnian Sea	26,737	83	1	16	15	3,518	183	20,663
Kokenmäenjoki	Bothnian Sea	26,667	79	10	11	8	9,695	474	15,167
Indalsälven	Bothnian Sea	25,458	79	4	17	15	4,581	138	14,614
Lule älv	Bothnian Bay	24,934	62	1	37	17	3,506	155	19,780
Oulujoki	Bothnian Bay	22,825	86	2	12	9	3,205	184	10,902
Ljusnan	Bothnian Sea	19,751	89	3	8	8	3,155	116	19,149
Iijoki	Bothnian Bay	14,264	92	2	6	6	2,240	137	15,673
Ljungan	Bothnian Sea	13,042	89	2	9	4	1,627	80	8,605
Skellefte älv	Bothnian Bay	11,577	76	1	23	5	1,418	50	6,467
Pite älv	Bothnian Bay	11,209	74	0	26	6	1,528	89	10,640
Unperturbed boreal									
Torne älv	Bothnian Bay	39,613	73	1	26	14	5,085	375	32,630
Kalix älv	Bothnian Bay	17,674	72	1	27	10	3,612	246	22,286

Finland. The mouths of the five largest rivers, the Neva, Wisła (Vistula), Odra (Oder), Nemunas and Daugava, are major point sources of nutrient emissions to the Baltic Sea (*cf.* Sect. 18.8). The other 16 major rivers are boreal rivers transporting less nitrogen and phosphorus, but relatively more humic substances, from forested areas. Most of these northern rivers discharge into the Gulf of Bothnia.

Recent human interferences in the drainage area resulted in, *inter alia*, reduced inputs of dissolved silica (DSi) to the northern Baltic Sea. Only two of the large boreal rivers, the

Torne älv and the Kalix älv, are unperturbed, while the other 14 have been used for hydroelectric power generation since 1920–1970 (Table 2.4). Damming reduces the DSi input to the sea due to longer residence times for water in the river systems. DSi inputs to the sea are also reduced by eutrophication of rivers through biogenic silica production (by diatoms) and subsequent sedimentation along the river system. Overall, the river-borne DSi loads entered into the Baltic Sea were estimated to have dropped by 30–40 % during the last century (Humborg et al. 2007). This may ultimately decrease

the biogenic silica production, notably that of pelagic diatoms, in offshore parts of the Baltic Sea proper.

2.3.7 Limited water exchange with the ocean

The narrow and shallow thresholds at the entrance of the Baltic Sea hamper water exchange with the North Sea and this, together with the large freshwater discharge into the Baltic Sea, is the main reason why the Baltic Sea is brackish. Water exchange between the Baltic Sea and the North Sea is governed by sea level differences, wind-driven currents and water density, the latter depending on salinity and temperature (Lass and Matthäus 2008).

The large riverine runoff to the Baltic Sea causes an increase in the water level from the entrance of the Baltic Sea to the Gulf of Bothnia by ~ 25 cm. This permanent sea level tilt is the result of the water density decrease with lower salinity towards the north. The less dense water in the northern Baltic Sea occupies a larger water volume than the denser water in the south. The outflow from the Baltic Sea increases with easterly winds and decreases with westerly winds and is also dependent on air pressure. In years with

heavy precipitation relative to evaporation in the Baltic Sea drainage area, the water outflow increases as well.

Like the Baltic Sea, the North Sea is a continental sea, but unlike the Baltic Sea it is not a semi-enclosed sea since it has a wide and deep connection to the Atlantic Ocean. Therefore, the salinity of the North Sea is close to that of the ocean, although its surface-water salinity (especially in the north-eastern North Sea) is influenced by the brackish-water outflow from the Baltic Sea, which represents ~ 60 % of all freshwater entering the North Sea (Leppäranta and Myrberg 2009).

If not interrupted by wind-driven currents, a continuous inflow of saline water from the Skagerrak forms the deep water of the Baltic Sea, while the less saline Baltic Sea water outflow occurs at the surface. This is because the more saline water has a higher density and is thus heavier (*cf.* Sect. 1.3.1). The inflow of the saline deep water is mainly governed by the bottom topography with sills and deeper channels (Fig. 2.3). Approximately 70–80 % of the saline inflow enters the Arkona Sea via the Storebælt and Lillebælt and the southern Belt Sea over the 18 m deep Darß sill, and 20–30 % via the Öresund over the 8 m deep Drogden sill (Mattsson 1996; Jakobsen and Trébuchet 2000). From the Arkona Sea, the saline water moves via Bornholmshavet to the Bornholm deep and via the Słupsk channel to the Gotland deep in the Eastern Gotland Sea. A shallow connection with the Baltic Sea proper prevents the deep water from entering the Gulf of Riga. From the Gotland deep, the water flow continues to the Fårö deep and further on to the northern Baltic Sea proper. The Åland Sea sills (Södra Kvarken) and the shallowness of the Archipelago Sea prevent the deep water from entering the Gulf of Bothnia from the Baltic Sea proper. Thus, it flows southward into the Western Gotland Sea with the Landsort deep and the Norrköping deep and eastward into the Gulf of Finland.

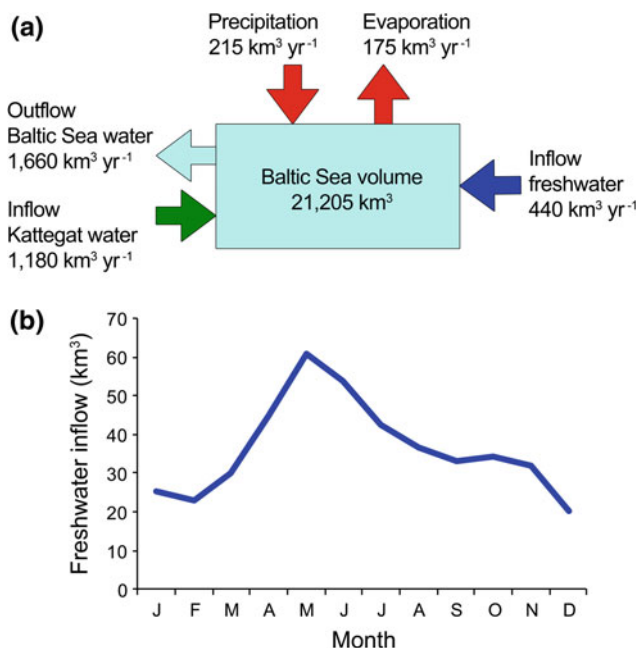


Fig. 2.12 Water budget of the Baltic Sea. (a) The average annual Baltic Sea water budget, showing freshwater runoff from land to the right, water exchange at the entrance to the Baltic Sea to the left and precipitation and evaporation at the top. According to this budget, the sum of water input and output is zero, leaving a water volume in the Baltic Sea (including the southern Belt Sea) of 21,205 km³. The net outflow to the Kattegat is 480 km³ year⁻¹. (b) The average monthly riverine runoff (inflow of freshwater) to the Baltic Sea for the time period 1901–1990. Figure (a) based on water budget data in Leppäranta and Myrberg (2009), (b) modified from Cyberski and Wróblewski (2000)

2.3.8 The Baltic Sea water budget

The annual average outflow from the Baltic Sea is estimated at 1,660 km³ ($\sim 52,600$ m³ s⁻¹). The estimated annual inflow is very similar, 1,620 km³, which is the sum of 1,180 km³ inflow from the Kattegat and 440 km³ freshwater inflow in the form of riverine runoff from the drainage area (Fig. 2.12a). This results in an annual net outflow from the Baltic Sea of 480 km³ ($\sim 15,000$ m³ s⁻¹) and constitutes ~ 60 % of the total freshwater supply to the North Sea. The riverine runoff from the drainage area to the Baltic Sea is largest in May–June and lowest in December–February (Fig. 2.12b). In long-term reconstructions for the time period 1500–1995 no significant long-term changes in the total riverine runoff to the Baltic Sea were detected, although decadal and regional variability was large and the runoff is sensitive to temperature decreasing by 3 % (450 m³ s⁻¹) per degree Celsius increase (Hansson et al. 2011a).

The outflow from the Baltic Sea equals the sum of the freshwater discharges into the Baltic Sea and the difference between precipitation and evaporation over the Baltic Sea. Precipitation is slightly higher than evaporation (Omstedt et al. 1997). Precipitation is 500–600 mm per year, with the lowest monthly values of 25–50 mm in December–May and the highest monthly values in July–September (50–75 mm). Evaporation is 450–500 mm per year with minimum values in the spring (10–20 mm during May–June) when the surface-water temperature is low, and maximum values in late autumn (70–80 mm during October–December) when the turbulent air-sea exchange is extensive (Leppäranta and Myrberg 2009).

A simple calculation of the water renewal time in the Baltic Sea, based on the total volume of 20,958 km³ and 480 km³ of freshwater runoff, yields 43.7 years. However, this is not a full estimate because water entering the Baltic Sea from the Kattegat may flow out again within a short time. Thus, water

masses closer to the entrance of the Baltic Sea tend to stay for a relatively shorter time in the sea compared to water masses farther away from the entrance. More accurate renewal times for the surface water in different parts of the Baltic Sea were estimated at 26–30 years for the Bornholm Sea, 28–34 years for the Gotland Sea, 34–38 years for the Bothnian Sea and 38–42 years for the Bothnian Bay (Meier 2007).

2.3.9 Major inflows from the Kattegat are rare

The normal water exchange with the Kattegat as described above is not strong enough to renew hypoxic and anoxic water masses in the deeper parts of the Baltic Sea, which is a prerequisite for animal life in and on deep soft seabeds (*cf.* Sect. 10.11). Renewal of deep waters only occurs during very strong and intensive inflow events of Kattegat water, which are, however, rare (Fig. 2.13a).

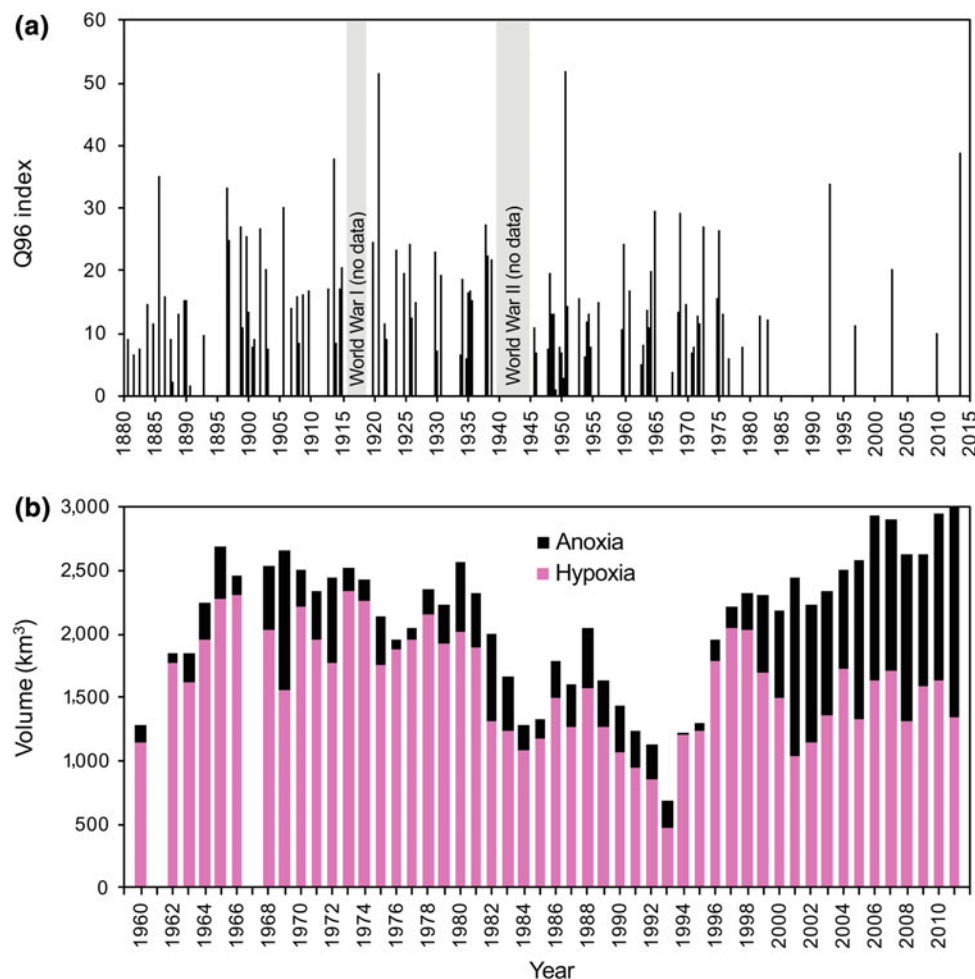


Fig. 2.13 Major inflows of saline water from the North Sea into the Baltic Sea and oxygen conditions in the Baltic Sea. (a) Inflow intensity from 1880 to 2014, expressed as “Q96”, which is an intensity index based on the total amount of salt transported through the Belt Sea and the Öresund into the Baltic Sea during an inflow event (Fisher and Matthäus 1996). (b) Oxygen conditions, shown as the total volumes of hypoxic water (>0 mL O₂ L⁻¹ and <2 mL O₂ L⁻¹) and anoxic water (≤ 0 mL O₂ L⁻¹) in the Baltic Sea 1960–2011. Figure (a) modified from Matthäus (2006) and Mohrholz et al. (2015), (b) modified from Hansson et al. (2011b)

Such major Baltic inflows (MBIs, Box 2.1) are infrequent because the required weather conditions are rare. They occur only when, at first, strong easterly winds dominate over the entrance to the Baltic Sea for several weeks, which then are followed by persistent, strong westerly winds (Lass and Matthäus 2008; Leppäranta and Myrberg 2009). The easterly winds increase the outflow of surface water and push the water level in the Baltic Sea down to a minimum level. The subsequent westerly winds force the Kattegat water to pile up in the Belt Sea and the Öresund, push the Baltic Sea water eastwards and press the Kattegat water over the sills into the Baltic Sea. During an MBI, the water level of the Baltic Sea can rise by one metre. Finally, the water masses with high salinity (and density) sink into the deeper areas of the Baltic Sea proper.

A strong MBI took place in December 2014. Together with the 1,913 MBI, this was the third largest one recorded since 1880. It was estimated that the total inflow of highly saline oxygen-rich water during the 2014 MBI was $\sim 198 \text{ km}^3$ ($\sim 4 \text{ Gt}$ salt), of which 138 km^3 (2.60 Gt salt) entered through the Storebælt and Lillebælt and 60 km^3 (1.38 Gt salt) through the Öresund (Mohrholz et al. 2015). While the MBI events in 1993 and 2003 interrupted the anoxic bottom conditions in the Baltic Sea only temporarily, the large 2014 MBI may have induced a longer-lasting improvement of oxygen levels. During 2015, this large new water inflow was slowly spreading northward and this has the potential to turn (most of) the hypoxic and anoxic deep water of the Baltic Sea proper into oxic conditions, with substantial consequences for marine life and biogeochemical cycles. However, it is not certain that this will actually happen.

Table 2.5 The solubility of oxygen in water (in mL L^{-1}) at different temperatures for salinities relevant for the Baltic Sea Area at a pressure of 1 atm. The data were calculated with the Oceanographic Calculator of the International Council for the Exploration of the Sea (<http://ocean.ices.dk/Tools/Calculator.aspx>)

Salinity	0 °C	5 °C	15 °C	25 °C
0	10.22	8.93	7.05	5.77
3	10.01	8.75	6.92	5.67
5	9.88	8.64	6.84	5.61
7	9.74	8.53	6.75	5.54
10	9.54	8.36	6.63	5.45
15	9.22	8.09	6.43	5.30
25	8.61	7.57	6.05	5.00
35	8.05	7.09	5.69	4.73

2.3.10 Hypoxia

Surface waters are always rich in O_2 because they are in contact with the atmosphere and mixed by winds. In addition, oxygen is produced from water during photosynthesis by primary producers in the upper part of the water column. In deep water, aerobic bacteria consume oxygen when they respire during the decomposition of organic material that sinks to the bottom from the upper water column. If there is a surplus of organic material in the system, the oxygen in the deep water is depleted and anaerobic bacteria take over (*cf.* Sect. 3.6).

Hypoxia (oxygen stress) occurs when there is a mismatch between oxygen supply and the demand for it. Hypoxia thresholds vary widely across marine benthic organisms (Vaquer-Sunyer and Duarte 2008). Hypoxia is usually defined as an oxygen concentration $< 2 \text{ mL O}_2 \text{ L}^{-1}$, which seems most relevant as a threshold for organisms that have evolved in normoxic waters without severe hypoxia as an evolutionary stress (Diaz and Rosenberg 1995). This threshold of $2 \text{ mL O}_2 \text{ L}^{-1}$ equals $2.9 \text{ mg O}_2 \text{ L}^{-1}$ or $91 \mu\text{M O}_2$. Other commonly used thresholds are $2 \text{ mg O}_2 \text{ L}^{-1}$ ($1.4 \text{ mL O}_2 \text{ L}^{-1}$ or $63 \mu\text{M O}_2$) or 30 % O_2 saturation (Rabalais et al. 2010).

Since the solubility of oxygen in water decreases with increasing salinity and temperature (Benson and Krause 1984, Table 2.5), the % oxygen saturation increases with increasing salinity and temperature. For example, a concentration of $1.4 \text{ mL O}_2 \text{ L}^{-1}$ generates 14 % O_2 saturation at salinity 5 and 0 °C, but 30 % O_2 saturation at salinity 35 and 25 °C. At a concentration of $2 \text{ mL O}_2 \text{ L}^{-1}$ the same salinity-temperature combinations generate 20 % and 42 % oxygen saturation, respectively. Because of the different thresholds applied and the different ways to express oxygen concentrations, it may be difficult to compare different studies on hypoxia.

2.3.11 Hypoxia and anoxia in the Baltic Sea

Hypoxia in the deeper basins of the Baltic Sea has been occurring since the Littorina Sea stage some 8,000 years ago (Zillén et al. 2008). In the geological development of the Baltic Sea, hypoxia has increased during warmer periods and decreased during colder periods. The average size of the seafloor in the Baltic Sea proper that during the last 40 years was affected by hypoxia and anoxia is $\sim 49,000 \text{ km}^2$ (23 %), including practically all deep bottoms (Conley et al. 2009; Hansson et al. 2011b). While anoxia is typical of stratified semi-enclosed seas (*e.g.* ~ 90 % of the Black Sea bottoms are anoxic), hypoxia and anoxia in the Baltic Sea

have - during the last century - increased drastically as a result of anthropogenic activities in the drainage area.

The major cause of the increased oxygen deficiency in the deep basins of the Baltic Sea proper is excess nutrient loading, which induces algal and cyanobacterial blooms and subsequently increases the sedimentation of organic material (eutrophication). An additional cause is the reduced incidence of MBIs of saline and oxygen-rich Kattegat water (Fig. 2.13a), so that bottom waters are not renewed. After large inflows of water from the Kattegat, the extent of hypoxic and anoxic bottoms decrease (Figs. 2.13b and 2.14). The relative importance of physical forcing (MBIs) *versus* eutrophication for hypoxia is still debated. A recent estimate is that hypoxia has increased 10-fold during the last 115 years, which is primarily linked to nutrient inputs from land (Box 2.2), although increased respiration due to higher temperatures during the last two decades has probably also contributed to worsening oxygen conditions (Carstensen et al. 2014).

In the Gulf of Bothnia, bottom oxygen conditions are much better than in the Baltic Sea proper because the saline below-halocline water of the Baltic Sea proper cannot enter

the Bothnian Sea over the Södra Kvarken sill. Thus, the Gulf of Bothnia has a much weaker halocline (*cf.* Sect. 2.4.3) and lower primary production in the photic zone as a result of lower nutrient concentrations, compared with the Baltic Sea proper. In the more eutrophic Gulf of Finland, hypoxia occurs because there is no sill between the Baltic Sea proper and the Gulf of Finland and the bottom oxygen conditions vary (Hansson et al. 2011b).

Anoxic bottoms are also widespread in the shallow coastal areas of the Baltic Sea (Conley et al. 2011). The lack of oxygen leads to the death of organisms that live in and on the bottom and weakens the function of the coastal zone as a nursery habitat for fish. Increases in the number and size of both small- and large-scale hypoxic areas in the Baltic Sea may be attributed to elevated nutrient levels resulting from activities in the drainage area: the excessive use of fertilisers, the presence of large animal farms with intensive livestock production, the burning of fossil fuels, discharges of effluents from *e.g.* municipal wastewater treatment plants, industrial point sources, fish farms and the disappearance of wetlands that act as nutrient traps.

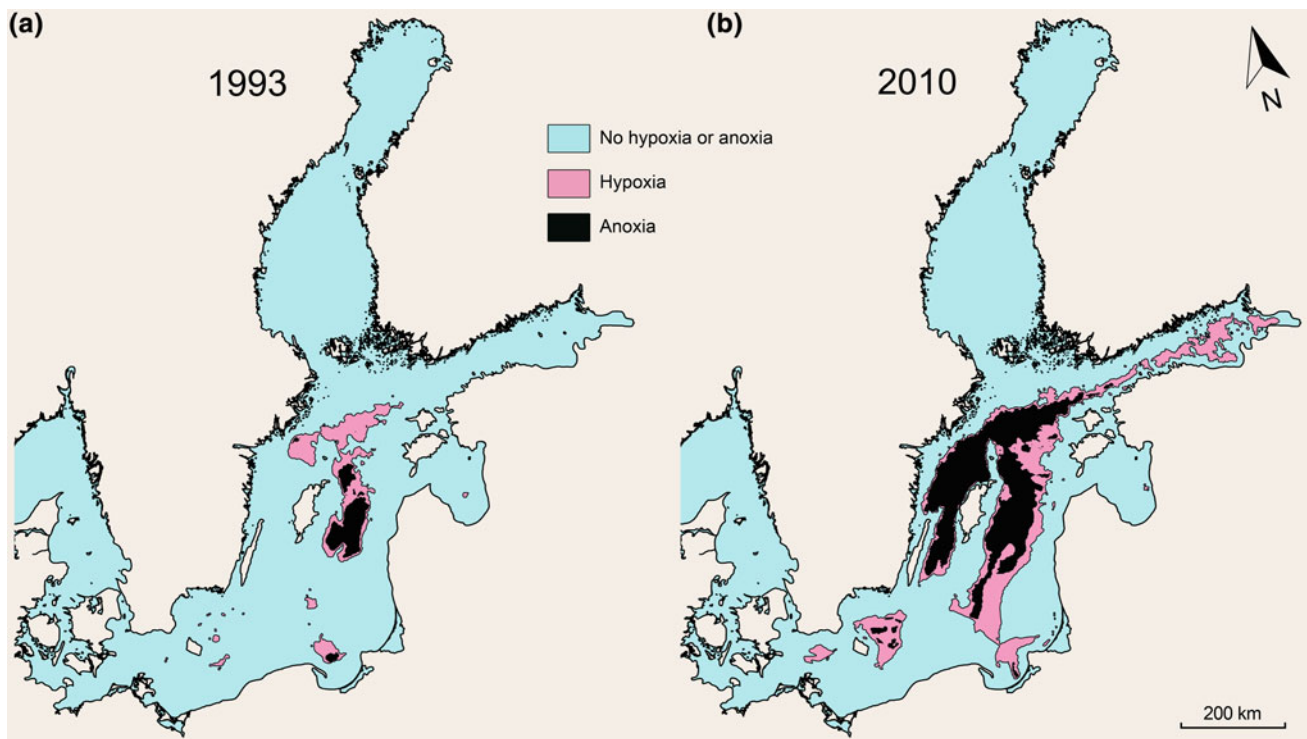


Fig. 2.14 Comparison of the extent of hypoxic conditions ($>0 \text{ mL O}_2 \text{ L}^{-1}$ and $<2 \text{ mL O}_2 \text{ L}^{-1}$) and anoxic conditions ($\leq 0 \text{ mL O}_2 \text{ L}^{-1}$) in the deep waters of the Baltic Sea with different inflows of saline water from the North Sea into the Baltic Sea (*cf.* Fig. 2.13a). (a) Oxygen conditions in the deep waters of the Baltic Sea in autumn 1993 after the large inflow in January 1993. (b) Oxygen conditions in the deep waters of the Baltic Sea in autumn 2010 after a long period of stagnation. Figure modified from Hansson et al. (2011b)

Box 2.2: External nutrient inputs to the Baltic Sea

Oleg Savchuk

Misbalance in nutrient cycles

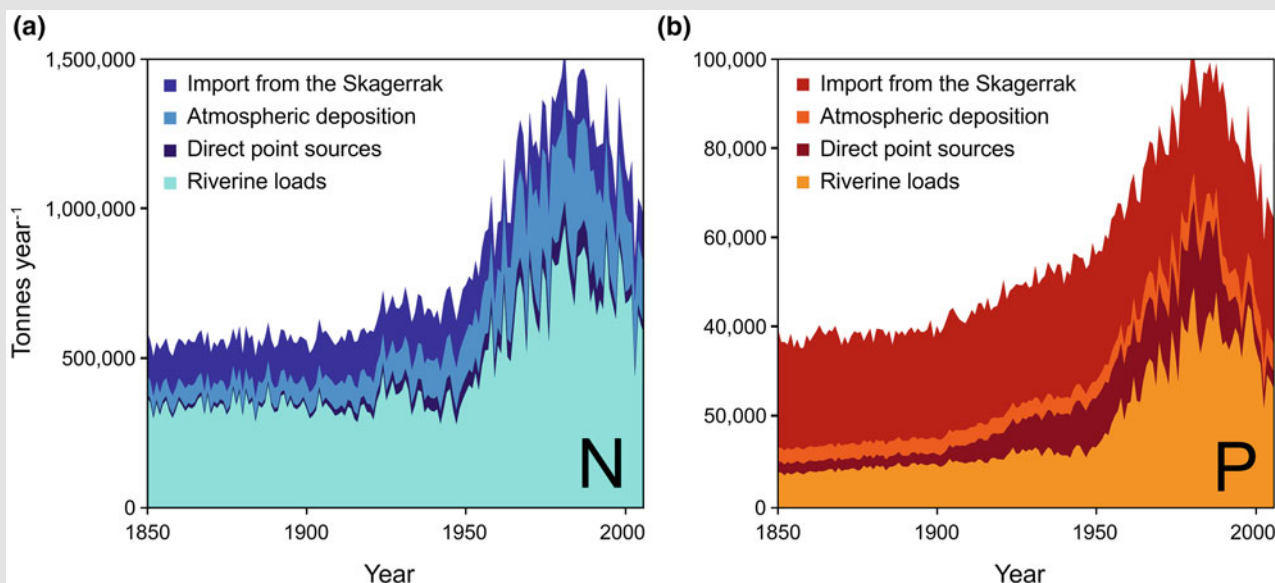
Eutrophication can be considered as the result of a misbalance in biogeochemical cycling in which more nutrients enter the system than leave it. Knowledge of the past and present nutrient inputs to the Baltic Sea is important, both for understanding the development of eutrophication (*cf.* Sect. 17.4), and for designing remedial measures in ecosystem management (*cf.* Sect. 18.5).

History of eutrophication in the Baltic Sea

The human eutrophication impact on the Baltic Sea became explicit after the 1950s (Box Fig. 2.2, Zillén et al. 2008). Current estimates of the loads of terrestrial origin and atmospheric deposition in 1970–2006 are based on good data with sufficiently high coverage and resolution (Savchuk et al. 2012a). However, because of the lack of reliable information on how historical nutrient inputs were generated by natural and human-made mechanisms within the entire drainage area and atmospheric deposition area, the temporal dynamics for 1850–1970 were reconstructed by a linear interpolation between loads calculated for as few as four points in time: 1850, 1900, 1950 and 1970 (Savchuk et al. 2012b). The water exchange and nutrient imports from the North Sea (Skagerrak) were estimated from reconstructed sea level variations and measurements of nutrient concentrations at the entrance of the Baltic Sea (Gustafsson et al. 2012), assuming a 15 % linear decrease of the concentrations from modern time back to 1900 (Savchuk et al. 2008 and references therein).

Four major sources of nutrient inputs

The relative contributions of the different nutrient sources follow well-known patterns. During 1977–2006, the most important sources of both N and P were the loads of terrestrial origin by rivers discharging into the Baltic Sea, which supplied about half of the total inputs (Box Fig. 2.2). The significance of other sources differs between the two nutrients: point sources, atmosphere, and the Skagerrak supply ~5 %, ~23 % and ~13 % of the total nitrogen input, respectively, while the contributions of these sources to the total phosphorus input are ~14 %, ~7 % and ~34 %, respectively. The pronounced decreases of the nitrogen and phosphorus inputs from the 1980s to the 2000s is caused not only by a naturally driven decline in freshwater discharges, but also by the reduction of atmospheric nitrogen emissions and by phosphorus removal in wastewater treatment plants.



Box Fig. 2.2 External inputs of nutrients to the entire Baltic Sea. (a) Total nitrogen. (b) Total phosphorus. Figure based on a reconstruction in Savchuk et al. (2012b). Figure: © Oleg Savchuk

2.4 Environmental gradients

2.4.1 Environmental gradients and the distribution of organisms

The geographical position and hydrographical features of the Baltic Sea together generate strong environmental gradients, which affect the distributions of organisms. Physical factors such as salinity, temperature, light and pH directly affect the performance and survival of organisms.

Differences in salinity produce the strongest environmental gradients in the Baltic Sea. Marine species from the North Sea and Atlantic Ocean enter the Baltic Sea via the transition zone and meet their lower salinity limit somewhere along the large-scale Baltic Sea gradient. Freshwater species enter the Baltic Sea all along its coasts and meet their higher salinity limit at some point along the inshore-offshore salinity gradient. Most of the Baltic Sea water column has a permanent halocline, which prevents water mixing and creates oxygen stress for the organisms living in the deeper areas, especially in the Baltic Sea proper. Areas with weak or no haloclines are the Gulf of Riga, the eastern Gulf of Finland and the Gulf of Bothnia.

The species distributions in the Baltic Sea also strongly depend on the climatic gradient, which stretches over 12 degrees of latitude from a temperate climate in the southern Arkona Sea to a subarctic (boreal) climate in the northern Bothnian Bay. Cold-adapted stenotherm species are abundant in the north and species requiring higher temperature populate the south. The long periods with ice cover in the north promote the occurrence of sympagic (ice-dependent) organisms, while the shading and scouring by ice hampers the growth of perennial macrophytes in the littoral zone.

Nutrient dynamics, which determine productivity, differ between the subbasins of the Baltic Sea. The Bothnian Bay is phosphorus-limited, similar to most temperate limnic environments, while the Baltic Sea proper is nitrogen-limited as are most seas. When considering both nitrogen and phosphorus, the Gulf of Riga is the most nutrient-rich area while the Gulf of Bothnia the most nutrient-poor area. Many environmental gradients (*e.g.* climate and nutrients) co-vary in some way with salinity and together they form the “large-scale Baltic Sea gradient”.

2.4.2 Large-scale and local Baltic Sea salinity gradients

The brackish water of the Baltic Sea is a mixture of marine North Sea water and freshwater from rivers and precipitation (Fig. 2.12). This creates the ~2,000 km large-scale salinity gradient between the Bothnian Bay and the Skagerrak, as

well as similar gradients from the inner Gulfs of Finland and Riga to the Skagerrak. The Baltic Sea surface-water salinity is 0 in the Neva Bay (easternmost Gulf of Finland), ~2 in the northern Bothnian Bay and the eastern Gulfs of Finland and Riga and ~10 in the western Arkona Sea (Fig. 2.15a, c). In the major part of the Baltic Sea, the year-round stable surface-water salinity varies between 5 and 8 along the gradient (Fig. 2.15a). Coastal sites in the Baltic Sea may experience considerable local salinity fluctuations which are governed by the intensity of land-runoff, and in such sites species are more adapted to salinity fluctuation than to a specific stable salinity.

Normally, the estimated daily and decadal variations of the surface-water salinity in the open Baltic Sea are below 0.1 and 0.5, respectively (Omstedt and Axell 2003; Meier and Kauker 2003). In the transition zone, the gradient is less stable and the surface-water salinity varies between 10 and 23 in the Belt Sea and between 12 and 30 in the Kattegat (Table 2.2). The transport of water masses in and out of the Baltic Sea is to a large extent controlled by winds and the surface-water salinity of both the Belt Sea and the Kattegat can vary strongly from day to day.

The Baltic Sea bottom waters display a horizontal salinity gradient as well, but with higher salinity than that of the surface waters (Fig. 2.15b, c). The high-salinity bottom water originates from the water flowing in from the Kattegat. Vertical salinity increases from the water surface to the seafloor are found in all subbasins of the Baltic Sea, but the salinity difference between surface and bottom waters decreases from south to north, depending on distance from the transition zone and bottom topography. For example, the saline deep water in the Baltic Sea proper is blocked by the Åland sills and the shallow Archipelago Sea (Figs. 2.2 and 2.3), and cannot enter the Bothnian Sea (Fig. 2.15c). The bottom-water salinity is ~15 in the southern Baltic Sea proper, ~12 in the Gotland deep (but higher after major inflow events) and only ~6.5 in the Bothnian Sea.

2.4.3 Variation of pycnoclines in space and time

The density of water is a function of salinity and temperature (*cf.* Sect. 1.2). High-salinity water and colder water are heavier than low-salinity and warmer water. In the Baltic Sea, the surface-water salinity is lower than that of the bottom water (Fig. 2.15), and in summer the surface water is warmer than the bottom water (Fig. 2.16). When water masses with different densities are not fully mixed, the denser water moves down and the less dense water moves up, and stratification ensues.

Between the water masses with different densities there is a pycnocline, a relatively thin layer in which the water

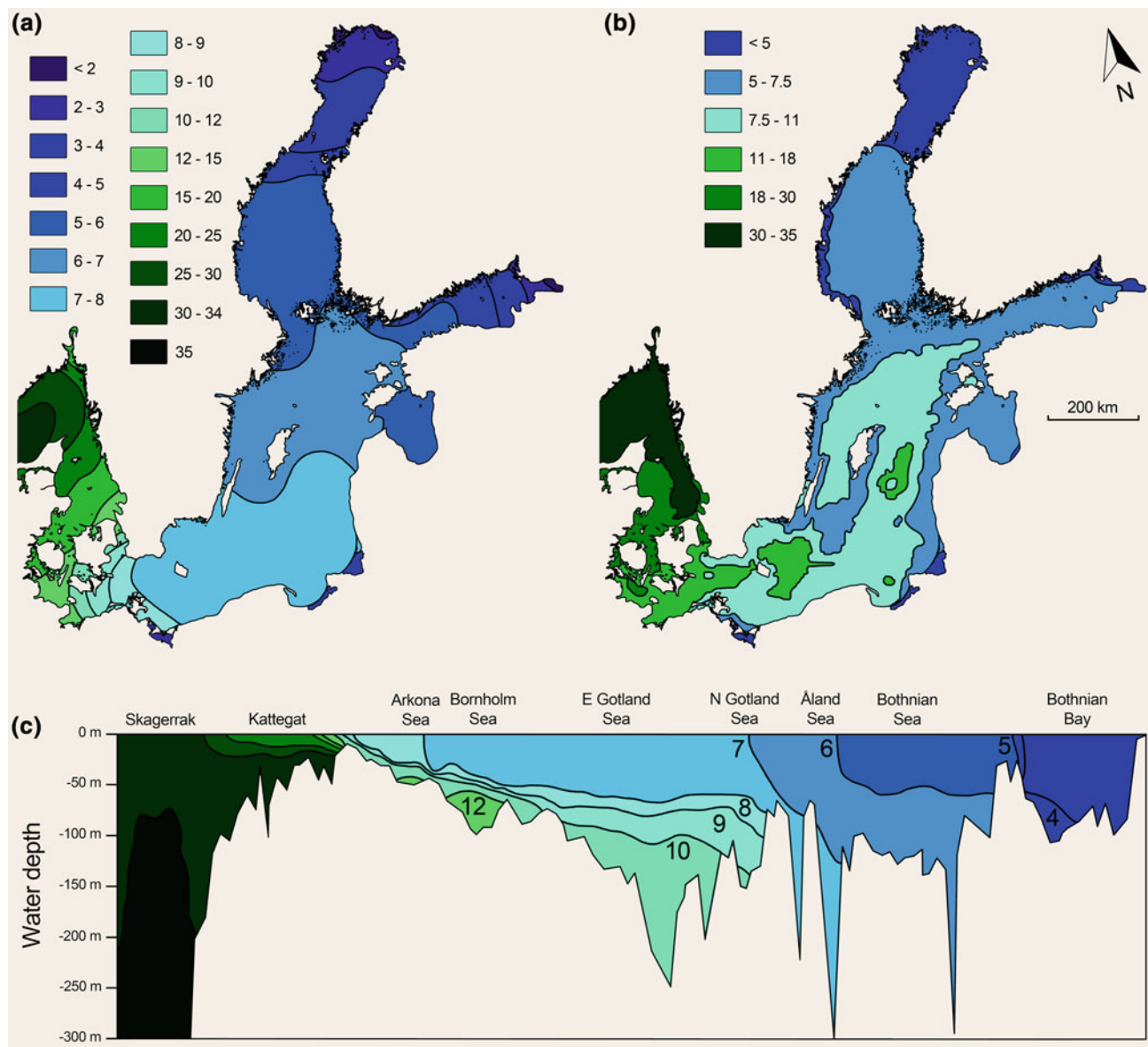


Fig. 2.15 The salinity gradient of the Baltic Sea Area. (a) Average surface-water salinity. (b) Average bottom-water salinity. (c) Schematic cross-section from the Skagerrak to the Bothnian Bay via the Eastern Gotland Sea, showing the vertical distribution of the salinity gradient. The legend for (c) is given in (a). Figure (a) modified from Bernes (2005), (b) modified from Al-Hamdani and Reker (2007), (c) modified from Sjöberg (1992)

density increases rapidly with depth. Pycnoclines constitute a barrier preventing homogeneous distribution of salt and temperature, but also of other compounds, *e.g.* nutrients and oxygen. Such a strong separation between different water masses can thus have a large impact on the distribution of organisms in the open Baltic Sea. In the coastal waters the water mass is generally well-mixed.

All basins of the Baltic Sea have a thermocline in the warm season (Fig. 2.16c). It is coincident with the pycnocline separating the warmer surface layer (upper 15–30 m) from the colder water below. In the southern Baltic Sea, the

seasonal thermocline starts building up in the beginning of May and in the Bothnian Bay this starts in June. Along the thermocline, the temperature drops by as much as 10 °C over a distance of a few metres. The surface layer is well-mixed and the vertical temperature change does not exceed 0.1 °C m⁻¹. In autumn, the surface water cools down, starts sinking and finally the thermocline disappears. During summer, wind-induced upwelling can locally produce a drastic drop in coastal surface-water temperatures (by up to 10 °C). Cooler and saltier deep water then flows up to replace wind-displaced surface water, bringing nutrient-rich

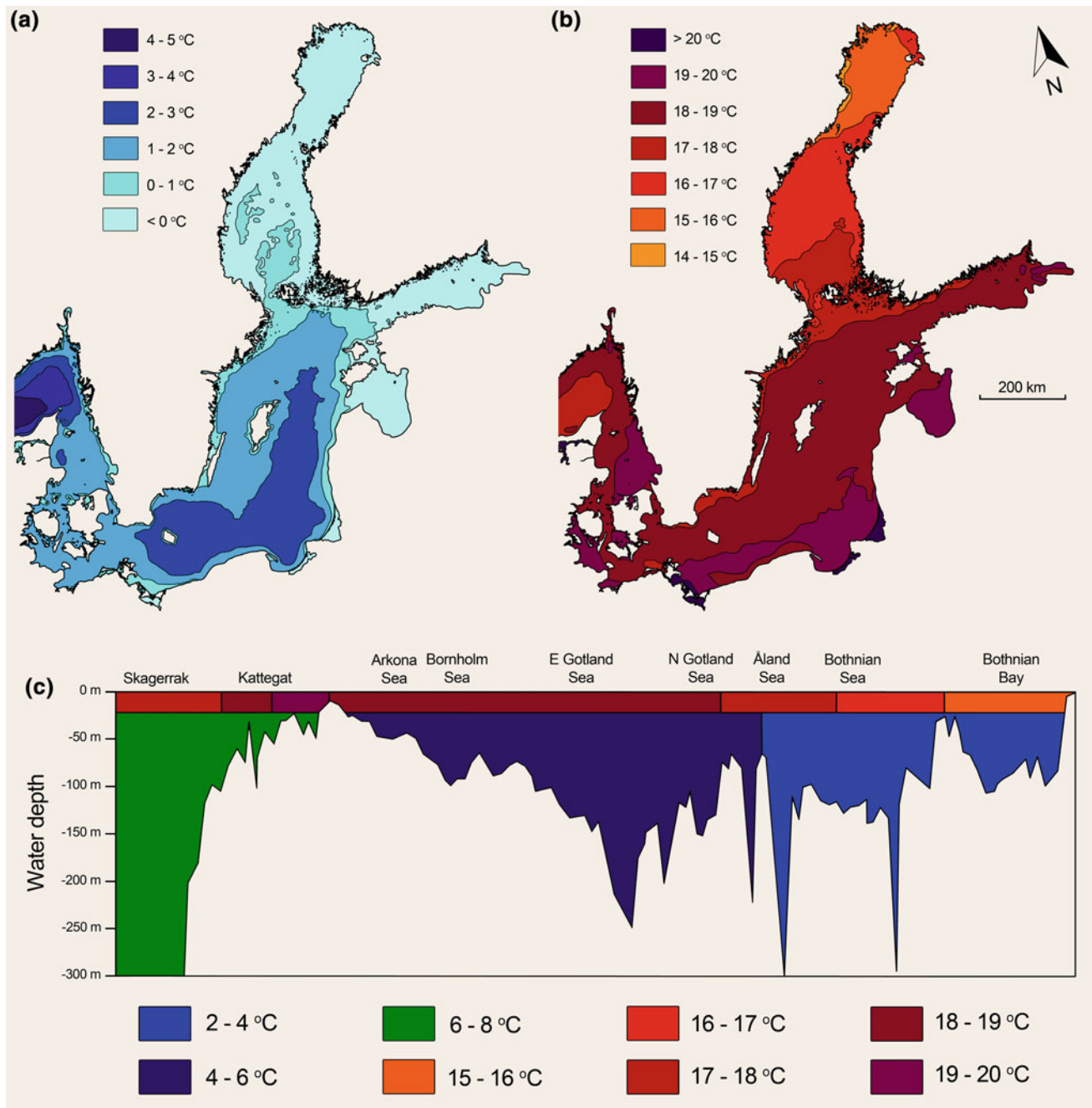


Fig. 2.16 The temperature gradient of the Baltic Sea Area. (a) Average surface-water temperature in February 1990–2005. (b) Average surface-water temperature in August 1990–2005. (c) Schematic cross-section from the Skagerrak to the Bothnian Bay via the Eastern Gotland Sea, showing the approximate vertical distribution of the temperature gradient in summer of the mixed surface-water layer with the thermocline above a 15–30 m water depth and the much lower water temperatures below 15–30 m. Figure (a, b) modified from Siegel et al. (2008), (c) © Figure: Pauline Snoeijjs-Leijonmalm

water to the photic zone (Myrberg and Andrejev 2003; Lehmann and Myrberg 2008), and this affects the growth and species composition of the biota in the sea.

The large volume of freshwater entering the Baltic Sea forms a low-salinity layer on top of the saltier bottom water throughout the sea, and not only in estuarine areas as is usually the case in fully marine seas. A permanent halocline,

which is a pycnocline between water masses of different salinities, separates the upper and lower layers in the Baltic Sea. This halocline is found at different depths in different parts of the Baltic Sea (Fig. 2.15c). A halocline occurs all the way to the Skagerrak, as long as the Baltic outflow water forms a layer with lower salinity on top of more saline deeper water. The halocline increases in depth and decreases

Table 2.6 The depth of the halocline and salinity ranges in the different subregions of the Baltic Sea Area. Data from Leppäranta and Myrberg (2009) and Andersen et al. (2015)

Subregion	Depth of the halocline (m)	Surface-water salinity	Deep-water salinity
Transition zone	15–20	10–30	30–34
Arkona Sea	20–30	8–11	10–15
Bornholm Sea	50–60	4–8	13–17
Gotland Sea	60–80	5–8	9–13
Gulf of Riga	20–30	4–6	6–7
Gulf of Finland	60–80	1–6	3–9
Bothnian Sea	60–80	4–7	6–7
Bothnian Bay	50–60	2–4	4–5

in strength from the Kattegat to the Gotland Sea (Table 2.6). In the Gulf of Bothnia, the halocline is extremely weak or absent. As the halocline is rather shallow in the Kattegat (15–20 m), it coincides with the summer thermocline, but in most of the Baltic Sea the halocline is located deeper than 50 m, *i.e.* far below the thermocline. Like a lid, the halocline limits the vertical mixing of water in the Baltic Sea. This implies that the oxygen content of the deep basins of the Baltic Sea proper can only be replenished by oxygen-rich saltwater flowing in from the Kattegat along the seafloor, and not by diffusion of atmospheric oxygen.

In the deep basins of the Baltic Sea, at water depths >100 m, a second pycnocline, a redoxcline, may occur (*cf.* Sect. 3.6.4). Below the redoxcline the water is anoxic and organic matter is oxidised by sulphate, which is reduced to toxic hydrogen sulphide.

2.4.4 The Baltic Sea climatic gradient: temperature and ice cover

The Baltic Sea stretches over more than 1,400 km from south to north. While the southern part of the Baltic Sea is located in the temperate zone, the northern part (just below the Arctic Circle) has a subarctic climate. In the south, the monthly average air temperature varies between 0 °C in winter and 20 °C in summer, while in the north it varies between –9 °C in winter and 18 °C in summer (Fig. 2.17a). This yields an average temperature gradient in the Baltic Sea surface waters of 4–5 °C between south and north, both in summer and in winter (Fig. 2.16a, b).

However, the largest differences in surface-water temperature are seasonal. February is, on average, the coldest month of the year with surface-water temperatures in the

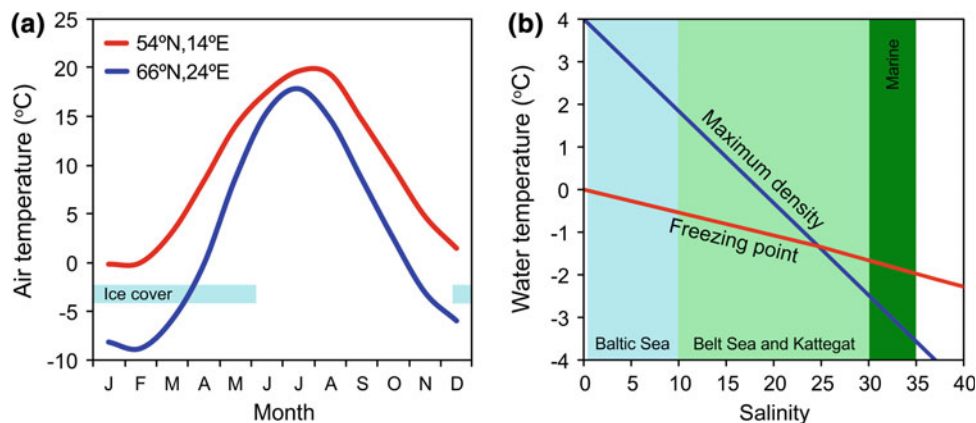


Fig. 2.17 Relationships between temperature and ice with respect to the Baltic Sea Area. (a) Annual variation of the monthly air temperature in the northernmost (66°N, 24°E) and southernmost (54°N, 14°E) Baltic Sea (averages for 1979–2011, $n = 33$ years). The horizontal light-blue line indicates the approximate time of ice coverage of the northernmost Baltic Sea. (b) The relationship between the temperature of maximum density and the freezing point for water of different salinity. At the intercept of the maximum density temperature and the freezing point (at salinity 24.7 and temperature –1.3 °C), the freezing properties of water change fundamentally. Water of salinity <24.7 is defined as brackish water from a purely physical point of view while in ecology water of salinity <30 is considered brackish (*cf.* Fig. 1.10). Figure (a) based on data extracted from the National Oceanic and Atmospheric Administration database (<http://www.noaa.gov>; 20th Century Reanalysis V2c dataset), accessed on 27 November 2015. Figure: © Pauline Snoeijs-Leijonmalm

range of <0 to $3\text{ }^{\circ}\text{C}$. August is, on average, the warmest month of the year with surface-water temperatures in the range of 15 to $20\text{ }^{\circ}\text{C}$. Bottom-water temperature is rather stable throughout the year, ranging from 4 – $6\text{ }^{\circ}\text{C}$ in the Baltic Sea proper to 2 – $4\text{ }^{\circ}\text{C}$ in the Gulf of Bothnia (Fig. 2.16c).

Because of the density anomaly of its water (*cf.* Sect. 1.2.3), the Baltic Sea behaves more like a freshwater lake than an ocean with respect to the formation of a winter ice cover (Fig. 2.18). Salt decreases the freezing point, and marine water of salinity 35 freezes around $-1.9\text{ }^{\circ}\text{C}$, but its maximum density is reached below that, at $-3.25\text{ }^{\circ}\text{C}$ (Fig. 2.17b). This means that vertical convection does not cease in an ocean like it does in a freshwater lake, and seawater does not freeze so easily, even if the air is cold. When a freshwater lake cools down below $+4\text{ }^{\circ}\text{C}$ (the maximum density of pure water), its vertical convection ceases and a thin upper water layer with lower density is formed, which freezes when the temperature decreases to $0\text{ }^{\circ}\text{C}$. Most of the Baltic Sea has the surface-water salinity between 5 and 8 (Fig. 2.15), and this water freezes around $-0.4\text{ }^{\circ}\text{C}$, while its maximum density is reached around $+2.5\text{ }^{\circ}\text{C}$ (Fig. 2.17b). Therefore, the water remains at the surface when its temperature is between -0.4 and $+2.5\text{ }^{\circ}\text{C}$, which speeds up further cooling and facilitates the formation of ice.

The surface water of the Bothnian Bay and the easternmost Gulf of Finland freezes every year (Fig. 2.19) and in the northern Bothnian Bay the ice cover persists for about half a year (Fig. 2.17a). Roads for car- and snow-scooter traffic are built on the ice in coastal areas, and previously this even made it possible to cross the Bothnian Bay between Sweden and Finland. The maximum extent of the ice cover in the Baltic Sea is normally reached in February–March.

More to the south, ice conditions vary extensively from one year to another. About once every decade in the period 1956–2005, the ice cover was so large that only a small area in the southern Baltic Sea remained ice-free. During this time period, the maximum extent of the ice cover varied between $\sim 50,000$ and $293,000\text{ km}^2$, *i.e.* 13–100 % of the whole Baltic Sea, while the yearly average was $\sim 218,000\text{ km}^2$. During World War II several winters in a row were severe and the Baltic Sea froze over completely.

The melting season starts in March–April in the south, but in the northern Bothnian Bay the last sea ice usually does not disappear before June (Fig. 2.17a).

2.4.5 The Baltic Sea insolation gradient

For primary production, the number of daylight hours is important because this determines for how many hours per day photosynthesis is possible. The light-harvesting step of photosynthesis depends on light energy (until the saturation

level, *cf.* Fig. 11.9), while carbon fixation is a temperature-dependent enzymatic process. However, primary production rates of phytoplankton are the outcome of the balance between light availability (including its penetration depth in the water column), temperature and the amount of available nutrients. Thus, longer days alone do not warrant higher primary production.

Because of the Baltic Sea's long latitudinal gradient, between $53^{\circ}55'\text{ N}$ and $65^{\circ}48'\text{ N}$, day length varies significantly between the south and the north throughout the year. In the south, mid-winter day length is ~ 7.5 hours while mid-summer day length is ~ 17 hours (Fig. 2.20a). In the northernmost Baltic Sea, the seasonal difference is much more extreme, with ~ 3 hours of daylight around mid-winter and 24 hours of daylight around mid-summer. Between the equinoxes on 21 March and 23 September (with 12 hours of light and 12 hours of darkness everywhere), the north of the Baltic Sea receives more daylight hours than the south, while the south receives more daylight hours during the cold half of the year. Despite the longer days in the north, the total daily amount of light energy in summer does not vary much between the north and the south of the Baltic Sea, but in mid-winter it is $\sim 90\%$ less in the north than in the south (Fig. 2.20b, c).

2.4.6 The photic zone

The photic zone is the upper part of the water column where the solar radiation is sufficient for photosynthesis to take place. The lower limit of the photic zone is generally defined as the depth to which 1 % of the sunlight penetrates (Kirk 2011). In clear ocean waters, the photic zone can be up to 200 m deep, but in the Baltic Sea it is only about one-tenth of that (Fig. 2.21).

Light penetration in natural waters is attenuated by a combination of coloured dissolved organic matter (CDOM), phytoplankton pigments and scattering by particles (*cf.* Sect. 15.2.3). The Baltic Sea is comparatively rich in CDOM, which mainly consists of humic substances, such as tannins, released from decaying plant detritus. The supply of CDOM from land is high, especially in the northern part of the Baltic Sea where boreal forests and bogs cover most of the land. Along the salinity gradient of the Baltic Sea, CDOM is inversely related to salinity (Fig. 2.22). However, the surface waters of the Kattegat and the Skagerrak are still significantly affected by CDOM in the outflow from the Baltic Sea (Stedmon et al. 2010).

In the Baltic Sea, the photic zone is around 20–25 m deep in the central parts of the subbasins and thins down to <5 m in coastal areas (Fig. 2.21). The photic zone is usually shallower near the coasts, partly due to the CDOM and sediment particles in the land runoff, but mainly because



Fig. 2.18 Winter conditions in the Baltic Sea. (a) The beginning of ice cover formation. (b) Snow-covered ice with open water in-between the ice sheets. (c) Snow-covered solid ice sheet; in winter, roads for car and snow-scooter traffic are made on the ice, especially in the Bothnian Bay. (d) A snow-covered beach in the southern Baltic Sea; the sea is covered with ice near the shore. (e) On board R/V Argos for seawater sampling in early March. (f) Diving to study perennial macroalgae in winter. Photo: (a–c, e) © Pauline Snoeij-Leijonmalm, (d) © Hendrik Schubert, (f) © Lies Van Nieuwerburgh

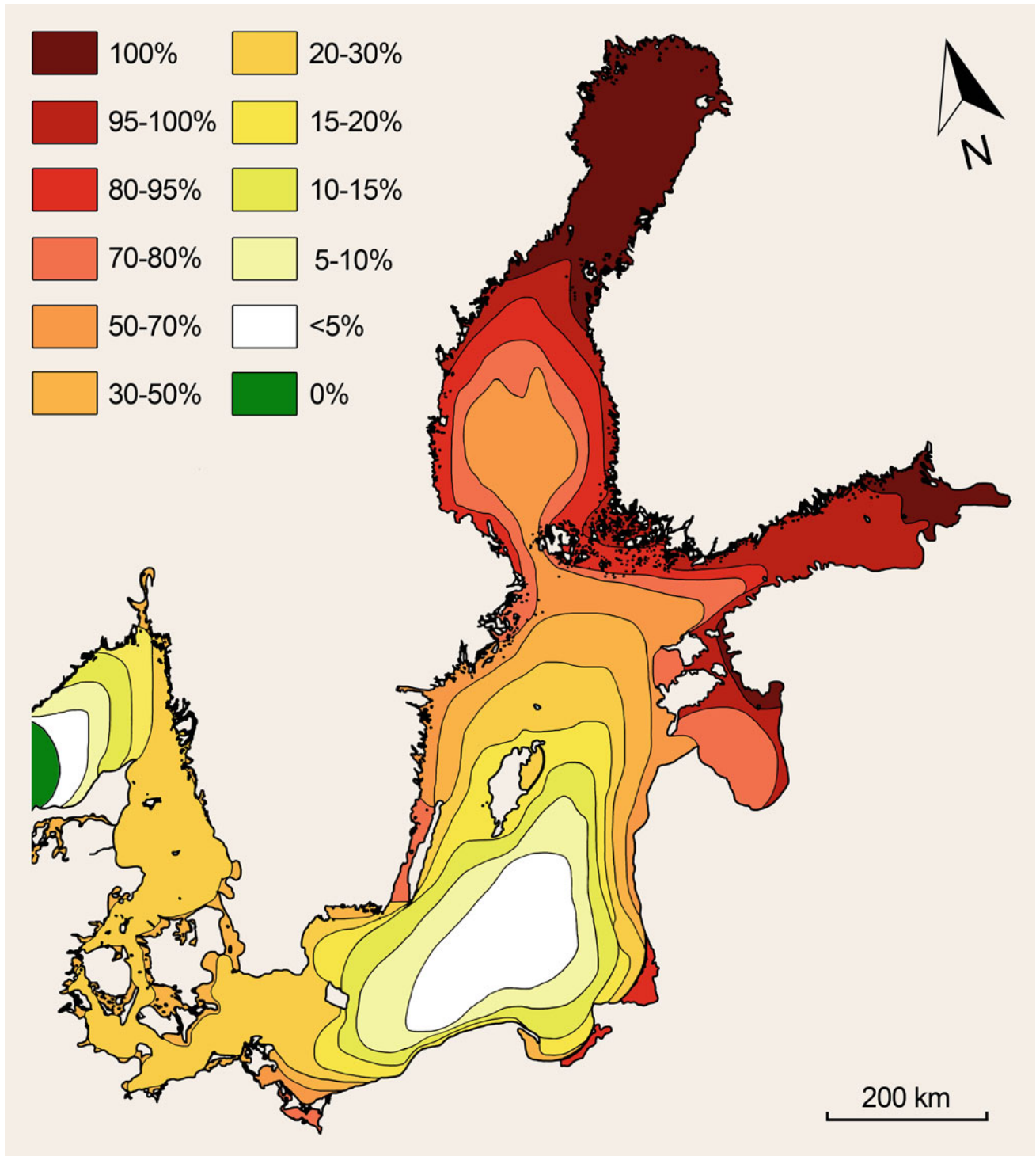


Fig. 2.19 The ice cover gradient of the Baltic Sea Area, shown as the average maximum winter ice coverage (in % of years) during 50 years (1956–2005). Figure modified from Schmelzer et al. (2008)

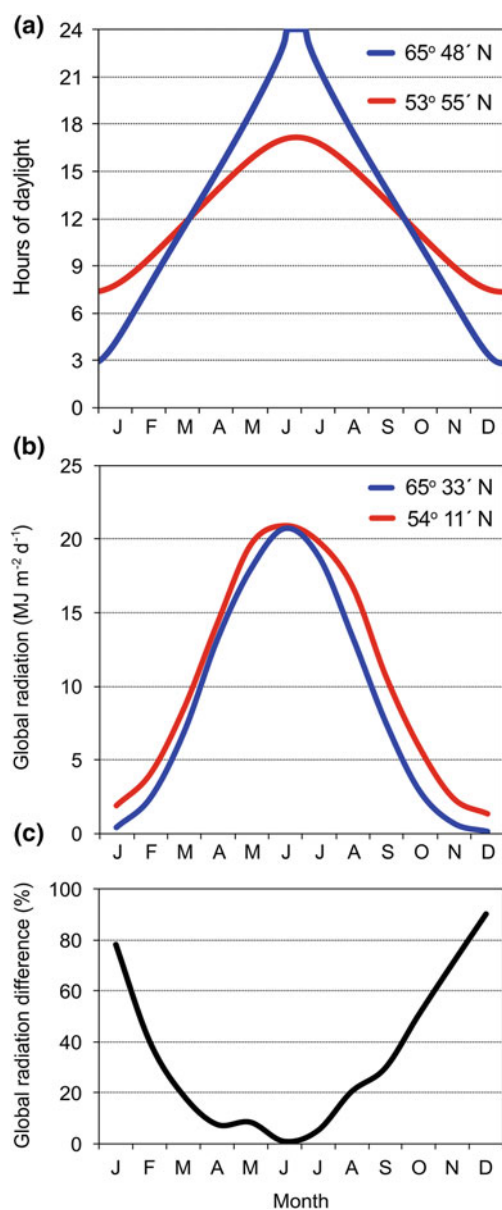


Fig. 2.20 Insolation in the northernmost and southernmost Baltic Sea. (a) Annual variation in day length during one year at 65°48' N near the outflow from the Torne älv estuary and at 53°55' N near the outflow of the Szczecin Lagoon. (b) Annual variation in measured global radiation (averages for 1965–2010, $n = 46$ years) at 65°33' N near Luleå (Sweden) and at 54°11' N near Kołobrzeg (Poland). (c) Difference in global radiation between Luleå and Kołobrzeg based on (b). Figure (a) based on data calculated with the Solar Calculator of the National Oceanic and Atmospheric Administration (<http://www.noaa.gov>), (b) based on data extracted from the Global Energy Balance Archive database (<http://www.geba.ethz.ch>), accessed on 25 November 2015. Figure: © Pauline Snoeijs-Leijonmalm

phytoplankton biomass is usually higher near the coasts than in the open waters as a result of nutrient emissions from land. The photic zone in shallow coastal areas also decreases when winds and waves whirl up particles from the sea bottom.

2.4.7 Basin-specific patterns of nutrient concentrations

The macronutrients nitrogen and phosphorus, and for some microalgae (*e.g.* diatoms and chrysophytes) also silica, are of major importance for biomass production since these elements often have low concentrations in natural waters compared to the other major constituents of organisms (C, O, H). For diazotrophic (nitrogen-fixing) cyanobacteria that are able to use elemental nitrogen (N_2) as their nitrogen source, phosphorus is the only growth-limiting macronutrient.

Thus, productivity can be limited by one or several of the nutrients nitrogen, phosphorus and silica. Primary producers most easily take up nitrogen and phosphorus as small ionic compounds, *i.e.* as dissolved inorganic nitrogen ($DIN = NO_3^- + NO_2^- + NH_4^+$) and dissolved inorganic phosphorus ($DIP = PO_4^{3-}$). Dissolved silica (DSi) occurs in natural waters as $SiO(OH)^{3-}$, $Si(OH)_2^{2-}$ and $Si(OH)_4$, of which diatoms utilise mainly $Si(OH)_4$ (Del Amo and Brzezinski 2000; Thamtrakoln and Hildebrand 2008).

Each subbasin of the Baltic Sea has its own typical nutrient concentrations, which are a combination of natural background concentrations and the eutrophication of the last 60 years (Fig. 2.23, Table 2.7, Box 2.2). The easiest way to compare the subbasins is to evaluate their winter nutrient concentrations as, during the rest of the year, nutrient concentrations are highly dynamic due to biological activity, which also differs among the subbasins (Granéli et al. 1990). The shallow Gulf of Riga has the highest winter DIN, DIP, total nitrogen (TN) and total phosphorus (TP) surface-water concentrations in the Baltic Sea, followed by the Gulf of Finland. The lowest DIN concentrations are found in the Baltic Sea proper and the Bothnian Sea, being only about half of those in the Bothnian Bay. The lowest DIP and TP concentrations occur in the Bothnian Bay, followed by the Bothnian Sea.

The DSi concentrations of the different subbasins are roughly related to the freshwater discharge they receive in relation to their water volume (compare Tables 2.2 and 2.4), and are highest in the three gulfs and lowest in the Baltic Sea proper and the Kattegat. Thus, DSi concentrations display a large-scale north-south gradient in the Baltic Sea in concert with the salinity gradient.

The DIN concentrations in the Kattegat are about twice as high as those of the Baltic Sea proper, while TN concentrations are similar (Fig. 2.23a, c). This is related to the higher CDOM levels in the Baltic Sea proper compared to the Kattegat (Fig. 2.22). CDOM contains dissolved organic nitrogen (DON), which photochemically can be made available to both heterotrophic and autotrophic plankton (Vähätalo and Järvinen 2007). Other processes that influence nitrogen-cycling, and which vary among the Baltic Sea subbasins, are nitrification and denitrification (*cf.* Sect. 3.6).

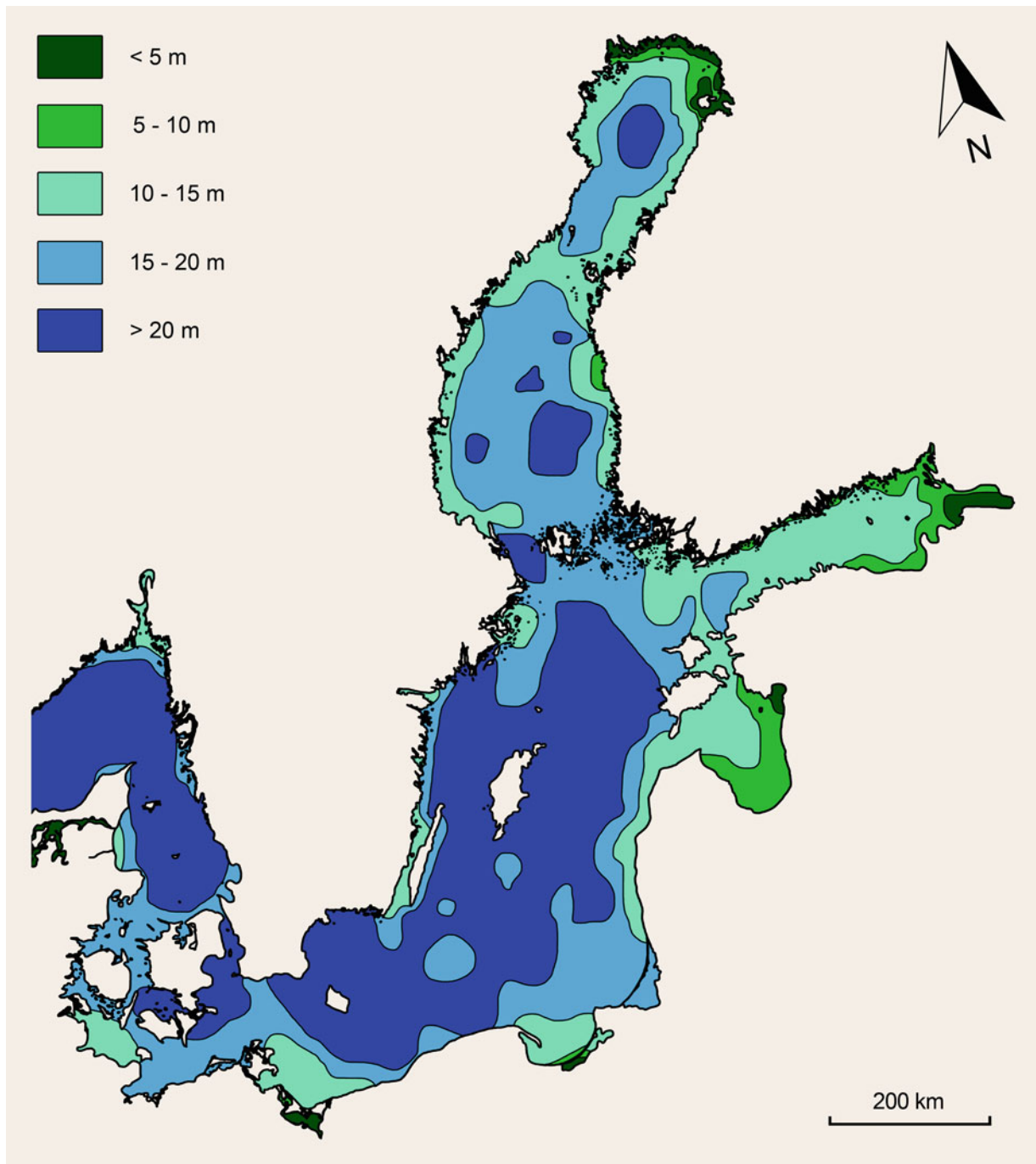


Fig. 2.21 Distribution of the photic zone depth (derived from Secchi depth data) in the Baltic Sea Area. Figure modified from Cameron and Askew (2011)

2.4.8 Basin-specific patterns of nutrient stoichiometry

Redfield (1934, 1958) discovered that the composition of marine particulate matter is relatively uniform and matching the metabolic demands of “average” phytoplankton. In due course, these findings were generalised to the universally accepted rule that the C:N:P molar ratio in phytoplankton is

106:16:1 (Redfield et al. 1963), known as the “Redfield ratio” for optimal phytoplankton growth (*cf.* Sect. 3.2.3). Later on this ratio was complemented with silicate to C:N:Si:P = 106:16:15:1 (Brzezinski 1985), known as the “Redfield-Brzezinski ratio” for optimal diatom growth. These ratios are widely used as reference levels for the assessment of nutrient depletion and nutrient repletion in primary producers and for the assessment of nutrient availability in aquatic

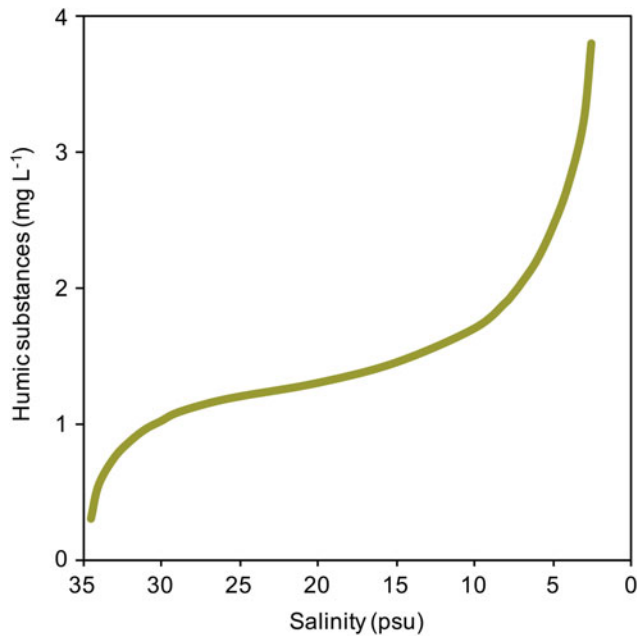


Fig. 2.22 The relationship between salinity and humic substances (CDOM) along the large-scale Baltic Sea gradient from salinity ~ 34 in the Skagerrak to ~ 2 in the Bothnian Bay. Figure modified from salinity Fonselius (1995)

environments, although the ratios may vary quite significantly in nature, especially in lakes (Hecky et al. 1993; Deutsch and Weber 2012).

Besides typical nutrient concentrations, the subbasins of the Baltic Sea also have their own characteristic nutrient stoichiometry, and – together with the absolute nutrient concentrations – this results in the chlorophyll *a* concentration being lower in the Gulf of Bothnia and higher in the Baltic Sea proper (Fig. 2.24). Eutrophication due to

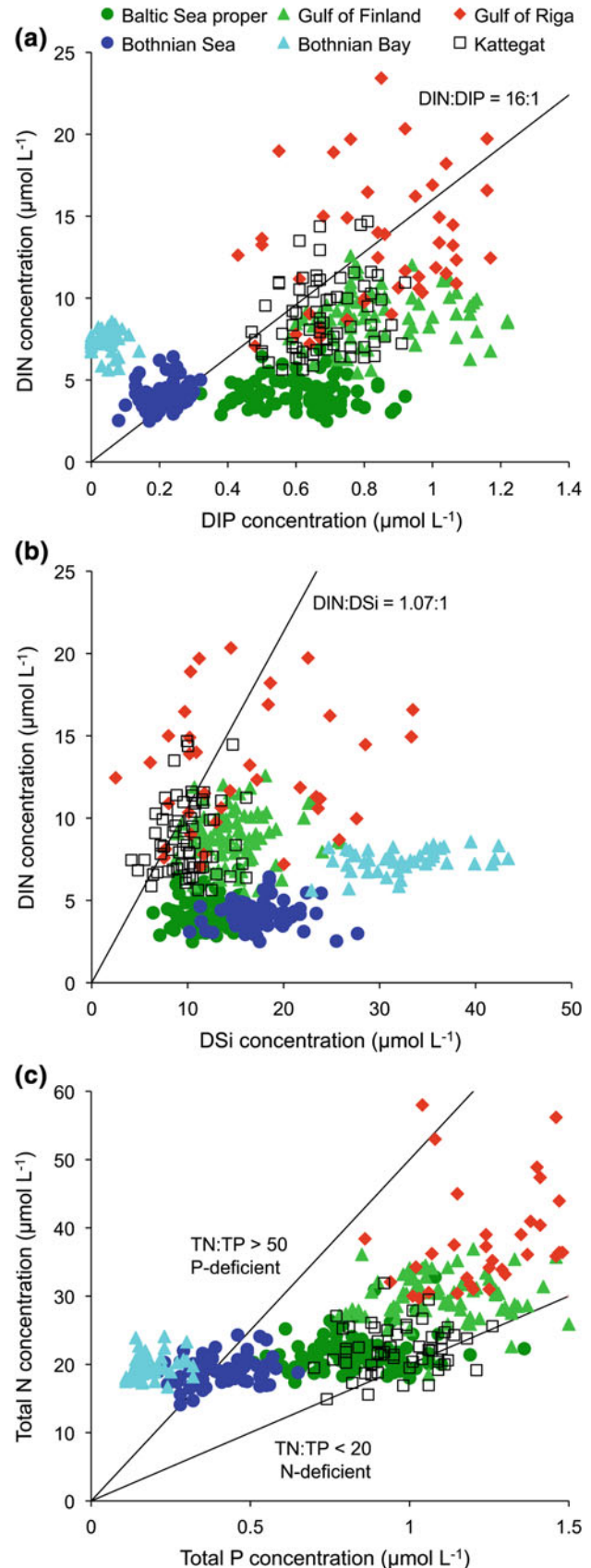


Fig. 2.23 The relationship between winter surface-water nutrient concentrations in the different subbasins of the Baltic Sea Area in 1980–2012. Note that these concentrations change throughout the year. Each score represents the highest reported winter concentration (January, February or March) for one year. For each subbasin, representative monitoring stations are included (Kattegat: GF4, GF8, Fladen; Baltic Sea proper: BY1, BY15, BY20; Gulf of Riga: G1, 119, 123; Gulf of Finland: F1, LL5, LL7; Bothnian Sea: F18, F26, EB1; Bothnian Bay: F3, B03). (a) Dissolved inorganic nitrogen (DIN) and dissolved inorganic phosphorus (DIP). The line represents the Redfield ratio (Redfield et al. 1963). (b) Dissolved inorganic nitrogen (DIN) and dissolved inorganic silica (DSi). The line represents the Redfield-Brzezinski ratio for optimal diatom growth (Brzezinski 1985). (c) Total inorganic and organic nitrogen, including particulate forms (TN) and total inorganic and organic phosphorus, including particulate forms (TP). The lower line represents $TN:TP = 20$, the upper line represents $TN:TP = 50$ indicating N- and P-deficiency limits, respectively, according to Guildford and Hecky (2000). Figure based on measurements accessed from the major databases around the Baltic Sea with the decision support system Baltic Nest (<http://nest.su.se>; Wulff et al. 2013). Figure: © Pauline Snoeijs-Leijonmalm

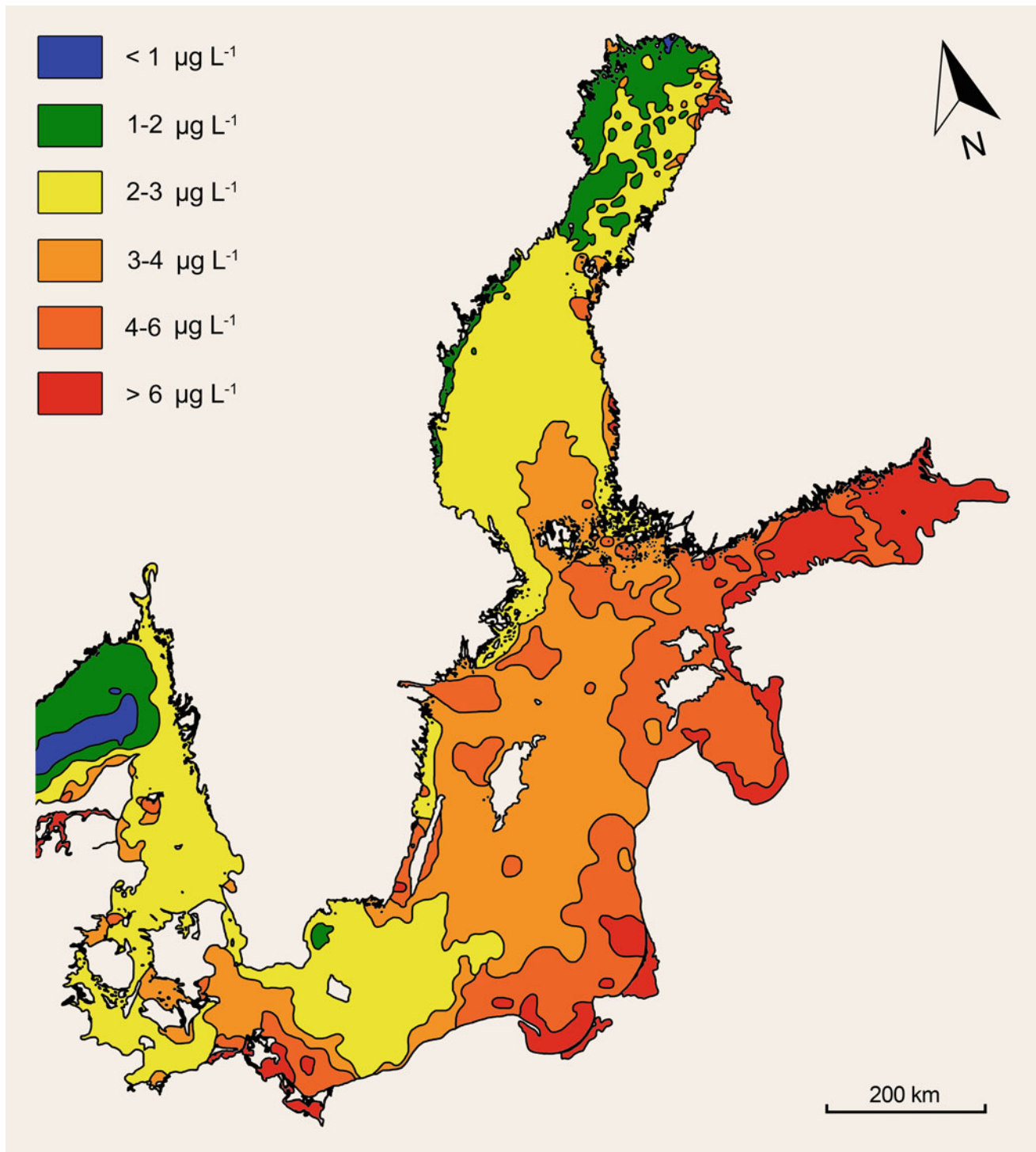


Fig. 2.24 Average chlorophyll *a* concentrations in the Baltic Sea Area in July–August 2004. Figure modified from chlorophyll *a* concentrations from the SeaWiFS satellite (<http://oceancolor.gsfc.nasa.gov>)

enrichment with nitrogen and phosphorus, which is superimposed on the natural nutrient dynamics, is most pronounced in the Gulf of Finland and the Gulf of Riga. Low winter surface-water DIN:DIP ratios of 7 and 11 indicate

nitrogen limitation in the Baltic Sea proper and the Gulf of Finland, respectively (Table 2.7, Fig. 2.23a). In contrast, the average winter DIN:DIP ratio of 203 in the Bothnian Bay indicates a strong phosphorus limitation and in this subbasin

Table 2.7 Average winter nutrient concentrations and stoichiometry in the surface waters in six subregions of the Baltic Sea Area based on the data shown in Fig. 2.23

Subregion	DIN \pm SD ($\mu\text{mol L}^{-1}$)	DIP \pm SD ($\mu\text{mol L}^{-1}$)	DSi \pm SD ($\mu\text{mol L}^{-1}$)	TN \pm SD ($\mu\text{mol L}^{-1}$)	TP \pm SD ($\mu\text{mol L}^{-1}$)	DIN:DIP	DIN:DSi	TN:TP
Kattegat	8.9 \pm 2.3	0.69 \pm 0.11	10 \pm 3	22 \pm 3	0.99 \pm 0.21	13 \pm 3	0.94 \pm 0.32	23 \pm 5
Baltic Sea proper	4.2 \pm 0.8	0.61 \pm 0.13	12 \pm 2	22 \pm 3	0.83 \pm 0.16	7 \pm 2	0.36 \pm 0.10	27 \pm 6
Gulf of Riga	14.7 \pm 7.5	0.86 \pm 0.21	17 \pm 9	39 \pm 11	1.25 \pm 0.26	18 \pm 8	1.08 \pm 0.80	31 \pm 8
Gulf of Finland	8.9 \pm 1.6	0.85 \pm 0.21	15 \pm 4	30 \pm 3	1.13 \pm 0.21	11 \pm 3	0.62 \pm 0.17	27 \pm 5
Bothnian Sea	4.1 \pm 0.8	0.21 \pm 0.05	18 \pm 3	19 \pm 2	0.41 \pm 0.10	20 \pm 6	0.24 \pm 0.06	50 \pm 12
Bothnian Bay	7.4 \pm 0.7	0.05 \pm 0.03	32 \pm 5	20 \pm 2	0.19 \pm 0.06	203 \pm 111	0.23 \pm 0.04	116 \pm 30

DIN is underutilised. However, estuarine systems in the Bothnian Bay are more complex than the open sea and display seasonal switches in nutrient limitation with in general phosphorus limitation in spring and nitrogen limitation in summer (Conley 2000). The average DIN:DIP ratios of the Gulf of Riga, Kattegat and Bothnian Sea lie closest to the Redfield ratio. The most variable of these three subbasins is the Gulf of Riga with a more or less equal distribution of DIN and DIP limitation, while the Kattegat is more often DIN-limited and the Bothnian Sea is more often DIP-limited.

The winter DSi values show that, compared with DIN, there is usually enough DSi for diatom growth in the Baltic Sea, since nearly all DIN:DSi ratios are <1.07 (Fig. 2.23b). DSi shows the same seasonal variation as the other nutrients with low values in summer and high values in winter, except in the Bothnian Bay, where low primary production and large DSi inputs via freshwater cannot alter the high concentrations over the year. In the southern part of the Baltic Sea proper, the situation is the opposite. Here, the diatom spring bloom can consume the whole winter period DSi pool (Wulff et al. 1990).

2.4.9 Stoichiometry of total nitrogen and total phosphorus

The total nutrient reservoir in the water column is often expressed as total nitrogen (TN) and total phosphorus (TP), which includes the dissolved inorganic, dissolved organic and the particulate forms of N and P, respectively. The particulate forms include the N and P bound in plankton organisms. A large meta-analysis comparing the TN and TP ratios in lakes and oceans with respect to nutrient limitation found that N-deficient growth is apparent at TN:TP <20 and P-deficient growth at TN:TP >50 , while at intermediate TN:TP ratios either N or P can become deficient (Guildford and Hecky 2000).

The winter TN and TP concentrations in the Baltic Sea subbasins fall mostly within the intermediate range, except for

the whole Bothnian Bay and partly also for the Bothnian Sea with TN:TP >50 (Fig. 2.23c). However, these observations should be interpreted with some caution. There are two aspects in particular that need to be considered: (1) the measurements were made by different laboratories in several countries around the Baltic Sea (using different methods for nutrient analyses), and (2) in general, TP measurements are more reliable than TN measurements (Hansen and Koroleff 1999).

2.4.10 Phosphate and iron

Altogether, the TN:TP ratio of the Baltic Sea proper is most similar to that of the Kattegat, while the TN:TP ratio of the Bothnian Bay deviates the most from the rest of the Baltic Sea and resembles that of a lake (Fig. 2.23c). In the temperate zone, near-neutral freshwater lakes tend to be phosphorus-limited and coastal seas tend to be nitrogen-limited, although the nutritional requirements of the phytoplankton are similar in both environments (Hecky and Kilham 1988).

This difference can partly be explained by the high sulphate content of sea salt (Blomqvist et al. 2004). In the oxidative hydrolysis of iron (Fe) and the concomitant precipitation of PO_4^{3-} , at least two Fe atoms are needed to precipitate one PO_4^{3-} ion. In anoxic marine bottom waters, Fe:P <2 predominates and some PO_4^{3-} is left in solution after oxygenation due to a shortage of dissolved Fe for PO_4^{3-} co-precipitation by iron oxyhydroxide [$\text{FeO}(\text{OH})$]. In contrast, anoxic bottom waters in most freshwater lakes have Fe:P >2 , which allows an almost complete PO_4^{3-} removal upon oxygenation. Thus, the general bottom-water chemistry in the habitat gradient from limnic to marine shows higher phosphorus availability in marine waters, primarily because of an enhanced iron sequestration by sulphides.

The regional variation in phosphate concentrations between the Baltic Sea subbasins can be further explained by eutrophication-driven microbial processes in sediments. In the nutrient-loaded Gulf of Finland and Baltic Sea proper, the sediments appear to have reached a state in which sulphate reduction is the dominant mineralisation pathway (Lehtoranta

et al. 2008). This implies that the capacity of the sediments to retain phosphorus is limited and high amounts of bioavailable PO_4^{3-} occur in the water column (cf. Sect. 3.6.7). The bottom sediments of the Gulf of Bothnia are still in a state in which iron reduction and coupled cycling of iron and phosphorus prevail in the surface sediments.

2.4.11 Patterns of primary production

Even if it partly depends on the light climate in the water, the concentration of the photosynthetic light-harvesting pigment chlorophyll *a* (Chl *a*) is often used as a proxy for the total biomass of primary producers. The phytoplankton Chl *a* concentration can be estimated from the colour of the sea by satellites, and can thus be monitored over large geographical areas at relatively short time intervals (cf. Sect. 15.1). As a result of changes in primary production, the Chl *a* concentration varies between seasons and years (cf. Sect. 8.2), and it can even vary significantly on a weekly basis.

The Baltic Sea displays an obvious spatial pattern of Chl *a* concentration, which is mainly shaped by natural nutrient conditions and anthropogenic nutrient inputs. In the Gulf of Bothnia, the Chl *a* concentration is generally low, while it is high in the Gulf of Finland, the Gulf of Riga, the Curonian Lagoon, the Gdańsk Bay and the northeastern coast of Germany (Fig. 2.24). Usually, the Chl *a* concentration is higher in coastal waters than in the open sea because of nutrient emissions from point sources (e.g. river discharges, cities) and diffuse sources (e.g. land erosion, agriculture). Another factor that influences the spatial distribution of the Chl *a* concentration is the water circulation pattern in the Baltic Sea (Fig. 2.10). For example, along the western coast of the Bothnian Sea the concentration is normally lower than along its eastern coast. This is partly caused by the transport of nutrient-rich water northwards along the Finnish coast by average currents, while nutrient-poor water is transported southwards along the Swedish coast (Fig. 2.24).

Given the climatic gradient of the Baltic Sea, the growing season is much shorter in the Bothnian Bay than in the southern Baltic Sea. This is, next to low nutrient concentrations (Fig. 2.23), a major factor that limits the annual phytoplankton primary production in the Bothnian Bay, which is as low as $17\text{--}28\text{ g C m}^{-2}\text{ year}^{-1}$ (Fig. 2.25). The values for the Baltic Sea proper ($67\text{--}163\text{ g C m}^{-2}\text{ year}^{-1}$) overlap with the average net primary production in the ocean, which is estimated at $140\text{ g C m}^{-2}\text{ year}^{-1}$ (Field et al. 1998). In comparison with estuarine-coastal systems worldwide, the average net phytoplankton primary production in the Baltic Sea region ($112\text{ g C m}^{-2}\text{ year}^{-1}$) is lower than the average in estuarine-coastal systems ($252\text{ g C m}^{-2}\text{ year}^{-1}$, Cloern et al. 2014). Using the classification of Nixon (1995), the Bothnian Bay and the Bothnian Sea are

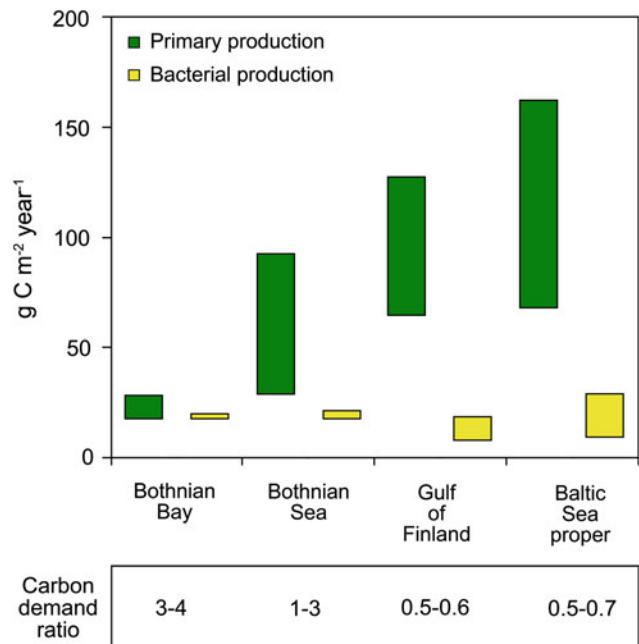


Fig. 2.25 Ranges of Primary production, bacterial production and carbon demand ratio (bacterial carbon demand:primary production, calculated with a bacterial growth efficiency of 25 %) in the pelagic zones of the four major basins of the Baltic Sea. Figure based on data calculated from Table 7.2 in Hagström et al. (2001). Figure: © Pauline Snoeijs-Leijonmalm

oligotrophic ($<100\text{ g C m}^{-2}\text{ year}^{-1}$) while the rest of the Baltic Sea is mesotrophic ($100\text{--}300\text{ g C m}^{-2}\text{ year}^{-1}$), although eutrophic ($300\text{--}500\text{ g C m}^{-2}\text{ year}^{-1}$) and hypertrophic ($>500\text{ g C m}^{-2}\text{ year}^{-1}$) conditions may occur locally in coastal areas with a high nutrient load.

2.4.12 Symptoms of eutrophication

Increases in primary production strongly influence ecosystem functioning (Nixon and Buckley 2002). For example, the eutrophication of the Baltic Sea that started in the 1950s (Box 2.2) seems to be the main reason for an average 15-fold (median 4-fold) increase of benthic animal biomass above the halocline (Cederwall and Elmgren 1980) and an eight-fold increase of fish biomass in the Baltic Sea with peak values in the 1970s–1980s (Thurow 1997), in addition to decreased benthic animal biomass (Cederwall and Elmgren 1980), increased supply of organic matter and a decline in oxygen concentrations below the halocline (Carstensen et al. 2014). Simultaneously, the frequency and intensity of potentially toxic cyanobacterial blooms (cf. Box 16.4) in summer have increased since the 1960s (Finni et al. 2001).

In the case of the Baltic Sea, the nutrient inputs exceed the natural processing capacity of the ecosystem with an accumulation of nutrients, while it is naturally susceptible to nutrient enrichment due to a combination of long retention

times and stratification restricting ventilation of deep waters (Diaz and Rosenberg 2008; Andersen et al. 2015).

2.4.13 Heterotrophy dominates in the Gulf of Bothnia

Organic matter of terrestrial origin may serve to compensate for the low primary production in the northern Baltic Sea, *i.e.* in the Bothnian Bay and the coastal Bothnian Sea. The carbon budgets for these areas show a higher bacterial carbon demand than can be supported by primary production (Kuparinen et al. 1994; Zweifel et al. 1995). In the Bothnian Bay, the bacterial carbon demand is four times that of the available carbon produced by autotrophs on an annual basis (Fig. 2.25). In fact, the Bothnian Bay resembles a subarctic lake more than a marine environment and should be considered as net heterotrophic (Hagström et al. 2001). In the Bothnian Sea, the carbon demand ratio is lower than in the Bothnian Bay, and in the Baltic Sea proper the primary production supports the bacterial carbon demand.

2.5 Geological and climatic background

2.5.1 Geology, climate and the distribution of organisms

The geological development of the Baltic basin, along with profound climate changes during the Holocene, has left its traces in the biota of the Baltic Sea in the form of evolutionary adaptations. This is especially apparent for the “glacial relicts”, which are descendants of species that survived in the area after glaciation events (*cf.* Sect. 4.6.2). Since ~20,000 years before present (BP), the Baltic basin has experienced total glacial coverage and several alternating freshwater and brackish-water stages thereafter, including a brackish sea with higher salinity than the present Baltic Sea.

Using a variety of methods (Box 2.3), geologists have been able to reconstruct the environment during the different geological stages of the Baltic basin. Throughout the geological development, organisms invaded and became extinct or adapted to the environmental conditions characteristic of the different stages.

During the last 3,000 years, the environmental gradients in the Baltic basin have been rather stable, although some geological processes are still ongoing (notably the isostatic rebound) and some natural climatic fluctuations have been rather prominent (*e.g.* the “Medieval Climate Anomaly” and the “Little Ice Age”). Today’s large-scale changes in the Baltic Sea environment caused by anthropogenic activities (*cf.* Sect. 17.2) occur incredibly fast compared to most natural geological, climatic and evolutionary changes.

2.5.2 Eustatic sea level rise and isostatic rebound

The late glacial and post-glacial development that shaped the present Baltic Sea was governed by interactions between the eustatic sea level rise and isostatic rebound. Eustasy refers to changes in the amount of water in the oceans. With increasing temperature, the water volume of the oceans increases by the melting of ice and snow as well as by thermal expansion, and *vice versa*. Isostasy refers to the gravitational equilibrium between the lithosphere and the viscous mantle of the Earth on which the tectonic plates are “floating”. When a heavy glacier (ice sheet) presses down the lithosphere by its weight, it takes a long time for the lithosphere to rebound to isostasy (the land rising) when the glacier melts, and even after it has melted.

As a consequence of the processes of eustatic sea level rise and isostatic rebound, the Baltic basin went through different stages during a geologically and evolutionarily short period of time (Björck 2008). When the climate became warmer ~15,000 years BP, the eustatic sea level rise and isostatic rebound followed the gradual melting of the Weichselian ice sheet that covered northern Europe during the latest ice age. Melting of the Earth’s ice sheets caused a 120-m sea level rise (Fig. 2.26a). The isostatic rebound after the heavy load of the ice sheet on Scandinavia is still ongoing, by up to ~9 mm per year in the northern part of the Baltic Sea (Fig. 2.26b). The forces in the lithosphere that cause the land uplift in the northern part of the Baltic basin result at the same time in a land subduction of ~1 mm per year in the southern part of the basin.

Since the sea level rise and land uplift opened and closed the connection with the ocean several times, the salinity of the water in the Baltic basin fluctuated between that of freshwater and brackish water during the last 12,000 years (Fig. 2.26c). The shoreline displacement, caused by an interaction between the isostatic rebound and eustatic sea level rise, in southern Sweden over the last 14,000 years BP has been estimated at ~70 m, while in the north the highest shoreline is situated 285 m above the sea level (Fig. 2.26d).

2.5.3 The Eemian interglacial

During the Eemian interglacial (~130,000–115,000 years BP), the Baltic basin developed in a manner comparable to that during the present interglacial Holocene (~11,600–0 years BP). However, the Saalian ice sheet (preceding the Eemian interglacial) was thicker and heavier than the Weichselian ice sheet (succeeding the Eemian interglacial), and the subsequent isostasy was larger. This resulted in a predecessor of the Baltic Sea, the Eemian Sea, which had a short-lived connection to the Barents Sea via the White Sea

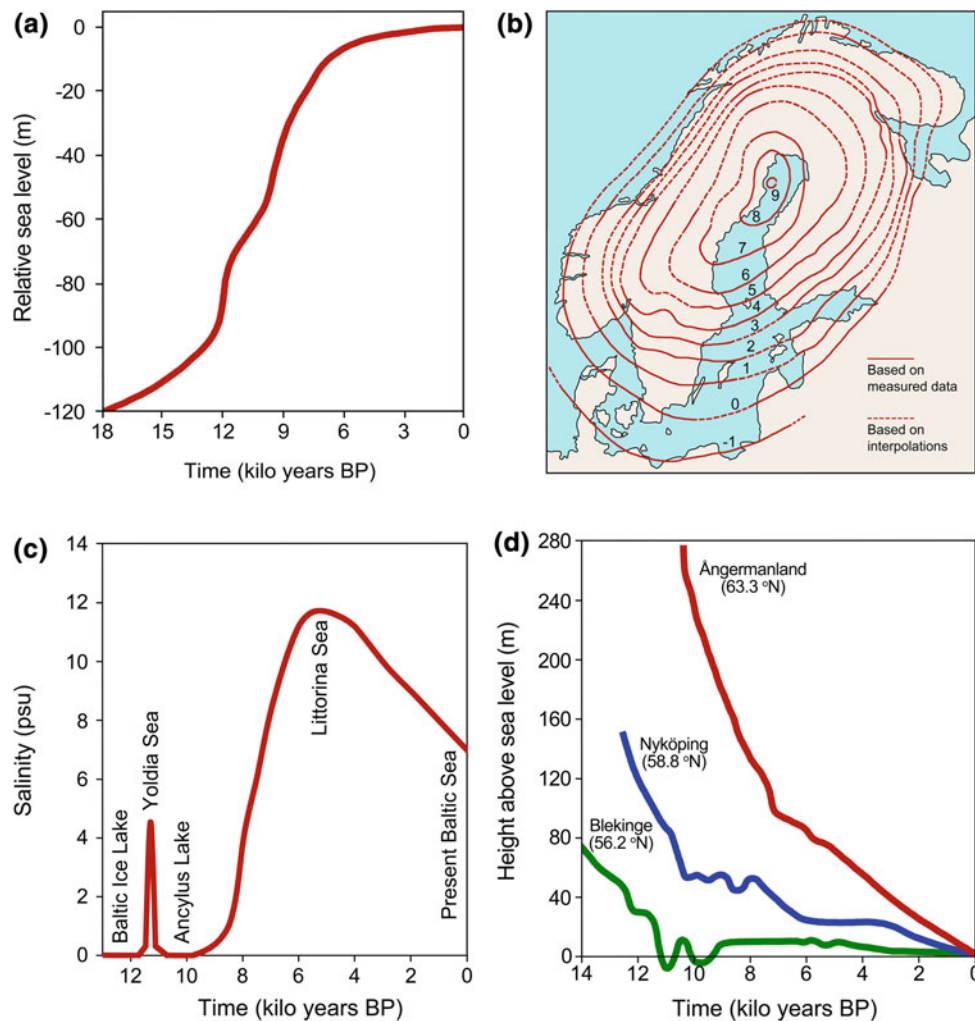


Fig. 2.26 Late-glacial and post-glacial changes in environmental conditions. **(a)** The eustatic sea level rise measured at Barbados. **(b)** The apparent post-glacial uplift of Fennoscandia, here shown for the period 1892–1991 in mm per year, based on sea level records, lake-level records and repeated high-precision levellings. **(c)** Tentative reconstructed surface-water palaeosalinity in the open Baltic Sea proper during the last 13,000 years, based on diatom frustules (data: Elinor Andrén) and different published proxy data, *e.g.* diatom, mollusc, silicoflagellate and cyanobacteria distributions, geochemical measurements (Gustafsson and Westman 2002) and strontium isotopes (Widerlund and Andersson 2006). **(d)** Shoreline displacement over time at three places in Sweden at different latitudes. Figure **(a)** modified from Fairbanks (1989), **(b)** modified from Ekman (1996), **(c)** Figure: © Elinor Andrén, **(d)** modified from Fredén (1994)

in addition to the westerly connection with the Atlantic Ocean. At that time, the Baltic basin experienced more marine conditions than today (Andrén et al. 2011). Since then, the Baltic basin has not been connected to the Arctic Ocean, but there have been several connections with the Atlantic Ocean in the west.

2.5.4 The Baltic Ice Lake: a dammed melt-water lake

The Weichselian ice sheet had its maximum areal extension around 20,000 years BP and covered at that time the whole Baltic basin and surrounding lands, from the northern half of

today's Great Britain and Ireland up to Svalbard, Novaya Zemlya and Franz Josef Land (Svendsen et al. 2004). When the warming of the global climate resulted in deglaciation, a melt-water lake, the Baltic Ice Lake, was formed in the Baltic basin in front of the receding ice. The Baltic Ice Lake (Fig. 2.27a) lasted from ~16,000 until ~11,700 years BP (Andrén et al. 2011).

The Baltic Ice Lake was a dammed freshwater lake in which glacial clays and silts were deposited. Seasonal cyclicity in the melt-water discharge, with high discharge in spring-summer and no melt water in winter, resulted in the deposition of laminated clays in the areas proximal to the ice sheet. These clays normally have thick, silty light-coloured summer layers and thin, clayey dark-coloured winter layers.

Box 2.3: Geological methods

Elinor Andrén and Thomas Andrén

Different types of data witness of past changes

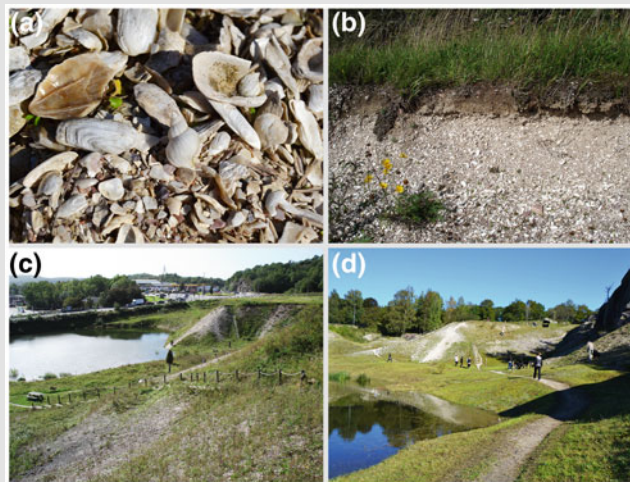
The present understanding of the post-glacial development of the Baltic basin is based on a variety of multiproxy data derived from a wide range of sedimentological, geochemical and biological techniques. The deglaciation of the Weichselian ice sheet was initially investigated from terrestrial traces of the ice sheet's movement and its melting such as glacial erratics, glacial striation, moraine ridges, eskers and the lithological composition of tills. A time scale following the Weichselian deglaciation in Sweden, known as the “Swedish Time Scale”, was constructed by measuring, counting and correlating the clay varves deposited in front of the receding ice sheet. By studying raised beach ridges and varved glacial clays, which contained fossil molluscs (Box Fig. 2.3), an account of the early Baltic Sea stages with shifting salinity was revealed in the late 19th century. As technology developed, it became possible to investigate bedrock and seabed sediments from research vessels by using hydroacoustic and coring equipment (Box Fig. 2.4), and this is the major approach used today.

How to find coring sites

The use of hydroacoustic profiling (*e.g.* with seismics, echo sounder or side scan sonar) enables us to follow the areal extension of lithological units below the seafloor and to identify the most suitable coring locations. Sediment cores serve as archives of past events and are used to study changes in the palaeo-environment and climate. Essential for all stratigraphical research are correct age determination and models. To achieve this, a number of different dating methods are available, ranging from radiometric (*e.g.* radiocarbon, uranium-series, ^{137}Cs , ^{210}Pb , optically stimulated luminescence) and annually laminated records (*e.g.* varved glacial clays) to tephrochronology and palaeo-magnetism.

Sediment properties

Multi-sensor core logging is a non-destructive geophysical method used on sediment cores, enabling continuous measurements of γ -ray attenuation, p-wave velocity, and magnetic susceptibility. Sediment properties such as density, porosity and water content can then be calculated, and this provides information on the origin of the sediments and if sedimentation is continuous. Such data can also be used for making correlations between different sediment cores.



Box Fig. 2.3 Huge amounts of marine shells were deposited in the nutrient-rich and turbulent environment at the narrow threshold area between the Yoldia Sea and the Kattegat ~11,000 years ago. The deposits, now situated outside the town of Uddevalla in western Sweden, were visited by Carl von Linné in 1746 (Linnaeus 1747). He wrote: “Skalbärgen räknas med rätta ibland et af de största Bohus-Läns under, ty de ligga uppå Landet, nästan hela quarten på somliga ställen, ifrån hafwet. Desse Skalbärgen bestå af Snäcke- och Mussle-skal, som här äro samlade i den myckenhet, at man kan undra det så många lif lefwat i werlden”. This roughly translates as: “The shell hills are rightly considered one of the largest wonders of Bohuslän as they are located on land far from the sea. These shell hills consist of sea snails and bivalves, which are gathered in such high abundance that one can wonder if there ever have been so many lives living in the world”. These fossil deposits are today protected in a nature reserve and a small museum informs the public of the site's extraordinary past development. Photo: © Elinor Andrén

Biological and geochemical proxies

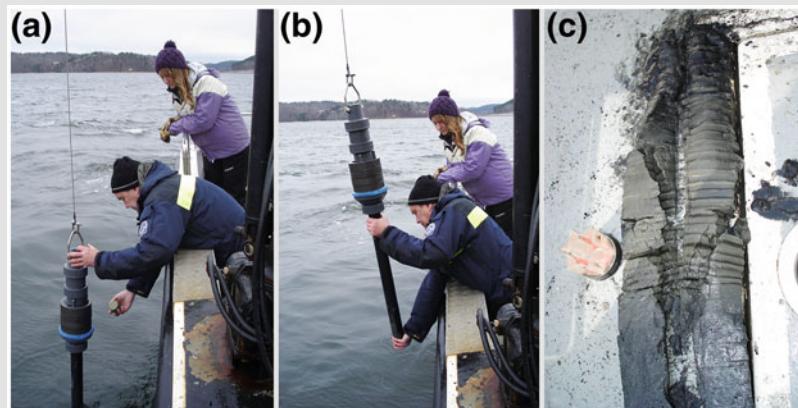
Several biological proxies, fossil remains of organisms preserved in the sediments, are used to assess changing environmental conditions over time, *e.g.* salinity, water depth, primary production and climate. A wide range of fossils can be found preserved in sediments, *e.g.* diatoms, molluscs, ostracods, foraminifers, silicoflagellates, dinoflagellate cysts, chrysophyte cysts, plant remains, fish skeletons and otoliths. The most widely used fossils for palaeo-environmental studies in sediments from the offshore Baltic Sea are diatom frustules. Geochemical measurements, such as the contents of organic carbon, biogenic silica and lipid and pigment biomarkers, are used to reveal *e.g.* changes in vegetation dynamics, primary production, salinity and water temperature.

Radioactive and stable isotopes

Isotopes are atoms of the same element, *i.e.* with the same number of protons in their nucleus, but with a different number of neutrons. Isotopes are radioactive when they have an unstable nucleus with excess energy, and these isotopes are subject to radioactive decay with time. Other isotopes have a stable nucleus and they do not change over time. A number of radioactive and stable isotopes are essential in geological research. Since the 1950s, radiocarbon dating has been used to determine the age of organic materials of up to $\sim 50,000$ calendar years old. Radiocarbon (^{14}C) forms in the upper atmosphere through the interaction between nitrogen and neutrons from cosmic rays. Plants fix all isotopes of atmospheric carbon in photosynthesis and the level of ^{14}C in living matter is in equilibrium with the ^{14}C levels in the atmosphere. However, after death of an organism the radioactive decay of ^{14}C into ^{14}N starts with a half-life of $\sim 5,730$ years and the $^{12}\text{C}:^{14}\text{C}$ ratio increases. By measuring the $^{12}\text{C}:^{14}\text{C}$ ratio in a material with organic origin (*e.g.* fossils, sediment and wood) it is possible to estimate the age of the material. Radiocarbon years need to be calibrated since the concentration of radiocarbon in the atmosphere has varied over time and is expressed as calendar years before present (BP), which corresponds to years before 1 January 1950. This date was agreed upon as a standard since nuclear weapons tests have changed the proportion of carbon isotopes in the atmosphere during the last ~ 60 years.

Isotopic fractionation

The deviating atomic weights of stable isotopes can cause isotopic fractionation. Processes in nature (*e.g.* varying temperature) can affect the relative abundance of isotopes of the same element and the ratio between heavy and light isotopes can be used to trace the process in nature. The stable oxygen ($\delta^{18}\text{O}$ = the $^{18}\text{O}:^{16}\text{O}$ ratio) and carbon ($\delta^{13}\text{C}$ = the $^{13}\text{C}:^{12}\text{C}$ ratio) isotope tracers in foraminifers are among the most important proxies in global palaeo-oceanography and powerful tools for palaeo-climatic reconstructions. Stable oxygen isotopes enable *e.g.* the reconstruction of past ocean temperatures, global ice volume, ocean circulation, river discharge, and surface-water salinity. Stable carbon ($\delta^{13}\text{C}$) and nitrogen ($\delta^{15}\text{N}$ = the $^{15}\text{N}:^{14}\text{N}$ ratio) isotopes are used to trace palaeo-primary production and carbon and nitrogen sources (*e.g.* to assess the proportion of material with terrestrial origin). Several other stable isotopes such as silica, strontium and sulphur are also used and show potential to improve palaeo-environmental interpretations and reconstructions.



Box Fig. 2.4 A gravity corer is used for sampling the most recent soft sediments. The number of weights on the corer can be adjusted to fit the softness of the sediment. When longer sediment sequences are required, a piston corer is used to secure coring of undisturbed sediments. (a) The corer is lifted up from the water. (b) The lower end of the corer is closed with a lid. (c) The sediment core is retrieved from the corer. Photo: © Elinor Andrén

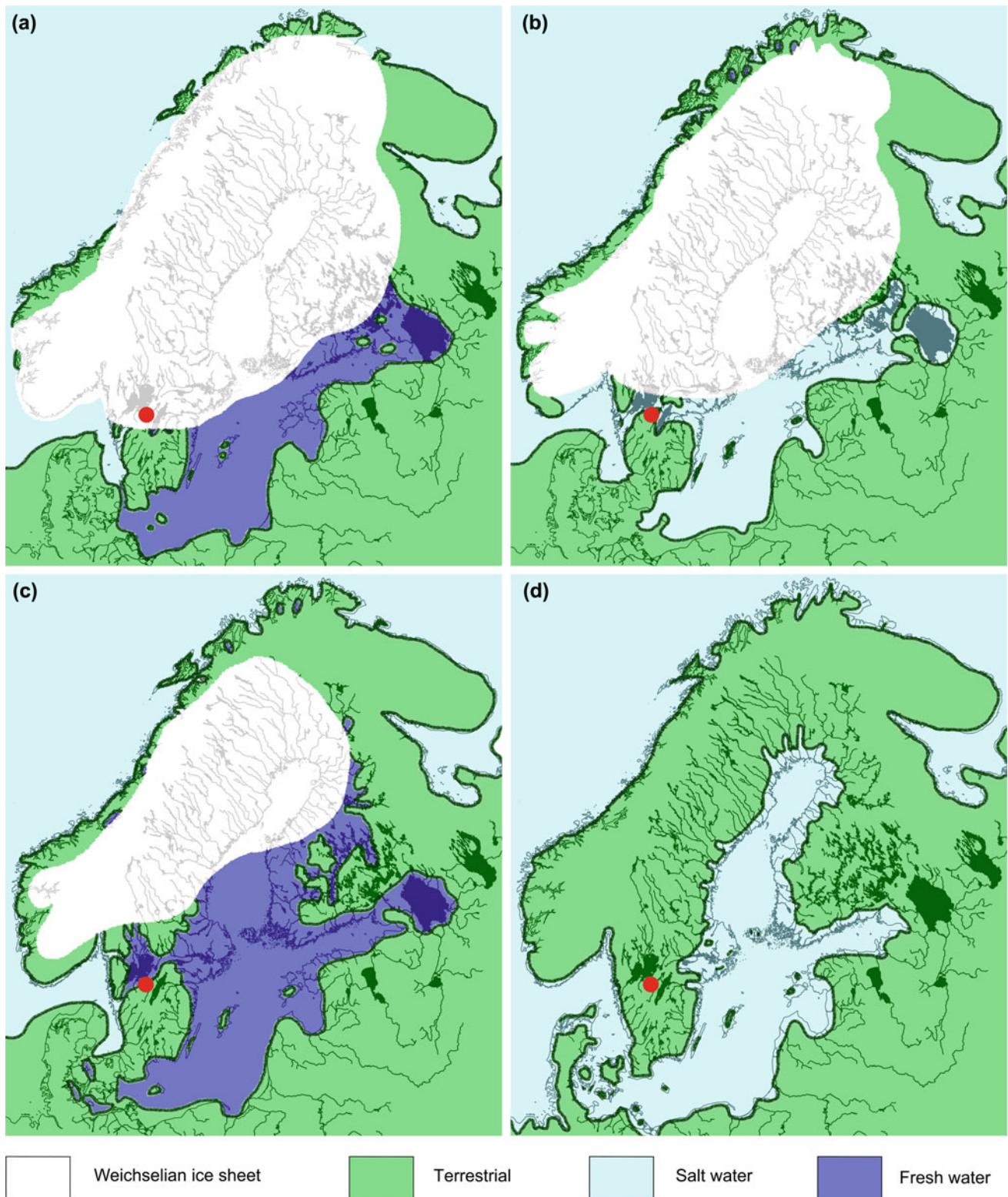


Fig. 2.27 Palaeo-geographic maps of the Baltic basin with the configuration of the present Baltic Sea and the major lakes and rivers in the present drainage area shown as background layer. (a) The Baltic Ice Lake just prior to its maximum extension and final drainage at ~11,700 years BP. (b) The Yoldia Sea at the end of the brackish phase at ~11,100 years BP. (c) The Ancylus Lake during its maximum transgression at ~10,500 years BP. (d) The Littorina Sea during the most saline phase at ~6,500 years BP. The red dot indicates the location of Mount Billingen. Figure modified from Andrén et al. (2011)

One clay varve (a sequence of one summer and one winter layer) reflects the deposition during one year. These varved glacial clays have been used to construct a deglaciation chronology named the “Swedish Time Scale” (De Geer 1912). In areas farther away from the ice margin, the seasonality was less pronounced and homogeneous clays were deposited.

The water transparency in the Baltic Ice Lake was probably low due to the heavy load of sediments from the melting ice sheet. Light conditions in the water, and possibly also a lack of nutrients, were unfavourable for the growth of photosynthetic organisms. Consequently, the Baltic Ice Lake was a barren water body with extremely low biological production (Winterhalter 1992). This is recorded in the sediments deposited during this time as a low organic carbon content and the absence of fossils.

During the early stage of the Baltic Ice Lake, its water level was the same as that of the ocean, which was ~ 100 m lower than at present due to the enormous amount of water still bound by the global ice sheets (Fig. 2.26a). The Baltic Ice Lake had an early outlet to the ocean in the Öresund area (Fig. 2.27a), where the easily eroded Quaternary deposits were superimposed on chalk. As soon as the erosion of the threshold in Öresund reached the bedrock $\sim 14,000$ years BP, the erosion ceased and the water level in the Baltic Ice Lake started to rise (Björck 2008).

2.5.5 Several drainages of the Baltic Ice Lake

The Baltic Ice Lake was dammed by the Scandinavian ice sheet at the northern point of Mount Billingen in south-central Sweden (Fig. 2.27a). When the ice sheet receded northwards, a connection between the lake, now at an altitude of ~ 10 m above the sea level, and the North Sea was created in south-central Sweden. This first drainage of the Baltic Ice Lake, possibly subglacial, occurred $\sim 13,000$ years BP (Björck 1995). There is no evidence of a marine inflow into the Baltic basin at that time (Andrén et al. 2011). The drainage ceased when a climatic cold event, the Younger Dryas, $\sim 12,800$ years BP, caused a re-advance of the ice sheet and closed the drainage path at Mount Billingen, so that the Baltic Ice Lake became dammed once again.

Because of the isostatic rebound, the threshold area was rising faster than the eustatic sea level, and the Baltic Ice Lake was drained by a waterfall in the Öresund area (Björck 2008). At the very end of the Younger Dryas cold event at $\sim 11,700$ years BP, a second, much more dramatic, drainage of the Baltic Ice Lake took place at the northern point of Mount Billingen, and lowered the water level of the Baltic Ice Lake by 25 m down to the oceanic water level within 1–2 years (Andrén et al. 2011).

2.5.6 The Yoldia Sea: at the level of the world oceans

The next stage in the development of the Baltic basin is named the Yoldia Sea ($\sim 11,700$ – $10,700$ years BP) (Fig. 2.27b) after the Arctic nutclam *Yoldia arctica* (syn. *Portlandia arctica*, Box Fig. 2.5), which is a common fossil in the Yoldia-stage sediments. After the sudden drainage of the Baltic Ice Lake, a passage between the Baltic basin (including Lake Vänern in western Sweden) and the Skagerrak was created in south-central Sweden. The passage included the area from today’s Göta älv river valley to Uddevalla and the Otteid/Steinselva straits at the border between Sweden and Norway. Even though there was an open connection between the Baltic basin and the ocean, it was initially only the cold melt water from the still receding ice sheet that passed through it. It took 300 years before the melt-water outflow had reduced enough for marine water to enter the Baltic basin.

The water exchange with the ocean resulted in a three-phase Yoldia Sea, consisting of a short (~ 150 – 350 years long) brackish-water phase, which coincided with the cold Preboreal climate oscillation, in-between two freshwater phases (Andrén et al. 2011). It has been suggested that the incoming marine water caused flocculation and subsequent rapid sedimentation of clay particles, which had previously prevented the sunlight from penetrating the water body (Winterhalter 1992). Thus, conditions for phytoplankton growth improved with the increasing depth of the photic zone, and the nutrient-rich marine water entering the Baltic basin served as a fertiliser.

High abundances of fossil diatom silica frustules deposited during this time provide evidence of an increased primary production in the pelagic zone during the short brackish-water phase. Brackish-water diatom species, such as *Thalassiosira baltica* (Box Fig. 2.7), indicate slightly brackish conditions in the open basin (Andrén et al. 2000). In the narrow threshold straits in south-central Sweden, fossils of ostracods, such as *Cytheropteron montrosiense* and *Paracyprideis fennica*, and the foraminifer *Criboelphidium excavatum* (syn. *Elphidium excavatum*) (Box Fig. 2.6), indicate a weak marine influence (Schoning and Wastegård 1999). The inflow of marine water can also be traced from sulphide bands or stains and weakly developed varved clays in the sediment, which suggests the development of a weak halocline in the Baltic basin.

2.5.7 The end of the Yoldia Sea

The isostatic rebound continued when the melting ice sheet retreated northwards. The threshold straits in south-central

Box 2.4: Shells used in palaeo-ecological studies of the Baltic basin

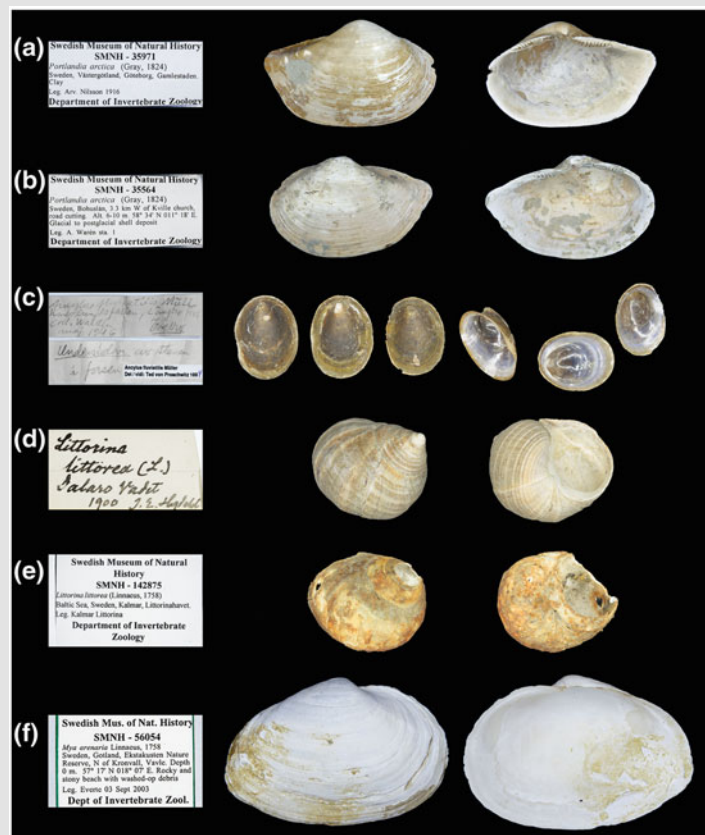
The molluscs that gave the geological stages of the Baltic basin their names

The Yoldia Sea is named after the Arctic nutclam *Portlandia arctica* (syn. *Yoldia arctica*), a marine infaunal detritivore, which lives mainly in the sublittoral zone on silty sediments in the Arctic Ocean (Holte and Guliksen 1998). Adults of this species, which does not occur alive in the present Baltic Sea, are usually 15–18 cm in length. The shell of *Portlandia arctica* (Box Fig. 2.5a, b) is a common fossil in the Yoldia-stage sediments and indicates cold marine conditions. The Ancylus Lake is named after the freshwater limpet *Ancylus fluviatilis*, a pulmonate gastropod, which is widespread throughout Europe. It typically occurs on stone surfaces in running waters with high dissolved oxygen concentrations. It is ~4–8 mm in size and seems to prefer diatoms as food (Callow 1973). The shell of *Ancylus fluviatilis* (Box Fig. 2.5c) is a common fossil in the Ancylus-stage sediments and indicates freshwater conditions. The Littorina Sea is named after the common periwinkle *Littorina littorea*, which occurs on marine intertidal rocky shores and is native to the northeastern Atlantic Ocean. It occurs in the Kattegat and the Belt Sea, but in this microtidal area it is mainly found in the sublittoral zone (Lauckner 1984). The food preference of *Littorina littorea* are foliose green ephemeral algae such as *Ulva* spp., and adults are usually 2–4 cm in size (Watson and Norton 1985). The snail itself is considered a delicacy in e.g. Scotland, Ireland and Belgium. The shell of *Littorina littorea* (Box Fig. 2.5d, e) is a common fossil in the Littorina-stage sediments and indicates marine influence. The geological stage of the present Baltic Sea is called the Mya Sea after the sand gaper *Mya arenaria* (Box Fig. 2.5f), which is a very common cryptogenic species in the Baltic Sea (cf. Box 5.2).

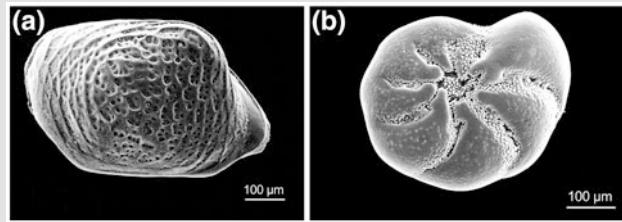
Small fossils in Yoldia-stage sediments

The Arctic ostracod (seed shrimp) *Cytheropteron montrosiense* is a small crustacean (~0.5 mm in body size) with a bivalve-like calcareous perforate shell (carapace) that protects its body. In Yoldia-stage sediments, *Cytheropteron montrosiense* carapaces (Box Fig. 2.5a) indicate cold marine influence (Schoning 2001). The foraminifer *Cibicides excavatum* is a common infaunal herbivore and detritivore of marine coasts and occurs in the Belt Sea (Schönfeld and Numberger 2007). This ~0.5 mm-sized amoeboid protozoan has a hard external skeleton (test) made

Box Fig. 2.5 Shells of the four mollusc species after which the geological stages of the Baltic basin have been named. (a) *Portlandia arctica* from Yoldia Sea deposits, 23 × 15 mm. (b) *Portlandia arctica* from Yoldia Sea deposits, 22 × 14 mm. (c) *Ancylus fluviatilis* from freshwater in central Sweden, 7 × 5 mm (recent material). (d) *Littorina littorea* from Littorina Sea deposits, 13 × 11 mm. (e) *Littorina littorea* from Littorina Sea deposits, 12 × 10 mm. (f) *Mya arenaria* from the Baltic Sea proper, 33 × 22 mm (recent material). Images photographed from the shell collection at the Museum of Natural History, Stockholm. Photo: © Pauline Snoeijs-Leijonmalm



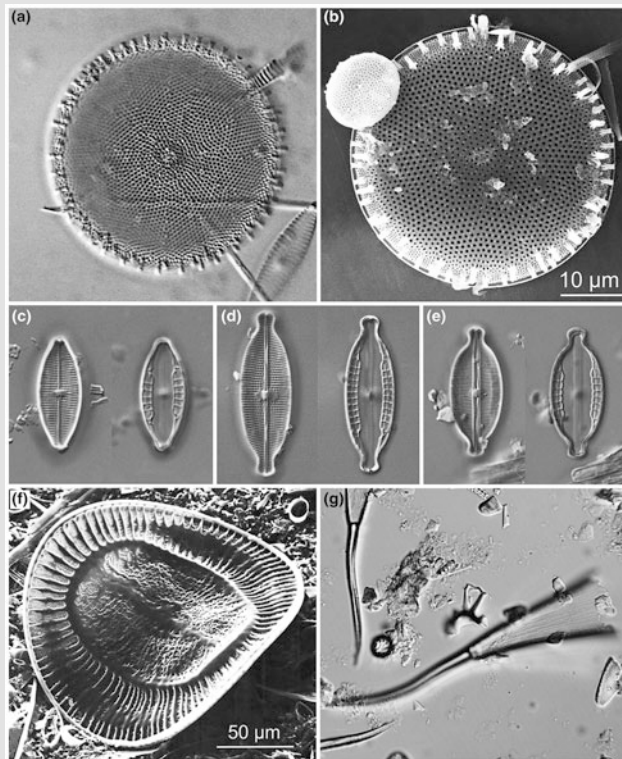
of calcium carbonate which is preserved in sediments. In brackish-phase Yoldia-stage sediments, *Criboelphidium excavatum* tests (Box Fig. 2.6b) indicate marine influence (Schoning 2001).



Box Fig. 2.6 Scanning electron micrographs of small shells from Yoldia-stage sediments. **a** Carapace of the ostracod *Cytheroapteron montrosiense*. **b** Test of the foraminifer *Criboelphidium excavatum* (syn. *Elphidium excavatum*). Photo: © Kristian Schoning

Diatom silica frustules

The centric diatom *Thalassiosira baltica* is a common phytoplankton species in the Baltic Sea with a diameter of $\sim 55 \mu\text{m}$. Diatoms have a silicified cell wall (frustule), which is preserved in sediments. In Yoldia-stage sediments, *Thalassiosira baltica* frustules (Box Fig. 2.7a, b) indicate slightly brackish surface water conditions. Benthic pennate diatom species belonging to the genus *Mastogloia* occur along the coasts of the Baltic Sea, mainly in places less exposed to wave action. These species are $\sim 20\text{--}40 \mu\text{m}$ long and they often live in a gelatinous matrix together with other diatom species. High abundances of *Mastogloia* frustules (Box Fig. 2.7c–e) are typical of the Initial Littorina Sea (originally named the “*Mastogloia* Sea”) and indicate slightly brackish conditions in coastal environments. Typical of *Mastogloia* are the distinct marginal chambers of their silica frustules. The epipellic diatom *Campylodiscus clypeus* occurs on soft bottoms in nutrient-rich brackish-marine lagoonal areas of the Baltic Sea. Its frustule (Box Fig. 2.7f) indicates such conditions in Littorina-stage sediments as suitable environments arose when lagoons were formed in uplift areas. The species is extant today, but may have been more common in subfossil times (Poulicková and Jahn 2007). The centric diatom *Pseudosolenia calcar-avis* is a marine phytoplankton species. Its frustule (Box Fig. 2.7g) is an indicator of the warmer and more marine conditions in Littorina-stage sediments during the Holocene Thermal Maximum ($\sim 8,000\text{--}4,000$ years BP) and the Medieval Climate Anomaly ($\sim 1,000$ years BP).



Box Fig. 2.7 Diatom silica frustules that indicate different environmental conditions in the geological stages of the Baltic basin. **(a)** Light micrograph of a *Thalassiosira baltica* frustule. **(b)** Scanning electron micrograph of a *Thalassiosira baltica* frustule, with in the upper left a frustule of the smaller species *Thalassiosira levanderi*. **(c)** Light micrographs of a *Mastogloia smithii* frustule with a different focus. The valve chambers are visible in the micrograph to the right. **(d)** Light micrographs of a *Mastogloia smithii* var. *amphicephala* frustule with a different focus. The valve chambers are visible in the micrograph to the right. **(e)** Light micrographs of a *Mastogloia baltica* frustule with a different focus. The valve chambers are visible in the micrograph to the right. **(f)** Scanning electron micrograph of *Campylodiscus clypeus*. **(g)** Remnants of *Pseudosolenia calcar-avis* frustules in sediment from the Gotland Sea. Photo: **(a–f)** © Pauline Snoeijers-Leijonmalm, **(g)** © Elinor Andrén

Sweden became more and more shallow and, due to the high outflow of melt water, no marine water could enter the Baltic basin anymore at the end of the Yoldia Sea stage. However, the very short (only a few hundred years long) brackish influence had a strong impact on the immigration of marine organisms, so-called “glacial relicts”, to the Baltic basin (De Geer 1932), e.g. the harp seal *Pagophilus groenlandicus* (syn. *Phoca groenlandica*), the ringed seal *Pusa hispida* (syn. *Phoca hispida*, cf. Box 4.13), the whiting *Merlangius merlangus*, the Arctic char *Salvelinus alpinus* and the benthic isopod *Saduria entomon* (cf. Fig. 4.25d), some of which still occur in the Baltic Sea. Furthermore, the land uplift and the still low sea level of the ocean resulted in a large land bridge between Scandinavia and the rest of the European continent, which facilitated the immigration of terrestrial plants and animals, including humans, following the retreat of the ice sheet.

2.5.8 The Ancylus Lake: dammed once again

Eventually, the threshold areas west of Lake Vänern rose, and a transgression (rising water level) occurred in the southern part of the Baltic basin, south of the line between Stockholm and Helsinki, causing a flood. North of this line, however, the isostasy was so large that a regression (decreasing water level) occurred. These conditions created a freshwater lake in the Baltic basin (including the basin of Lake Vänern), which is named the Ancylus Lake (Fig. 2.27c) after the freshwater limpet *Ancylus fluviatilis* (Box Fig. 2.5). The Ancylus Lake lasted from ~10,700 until ~9,800 years BP.

Sediments deposited during this stage consist of homogeneous clays, which suggests a fairly oligotrophic environment with low organic carbon accumulation and a diatom flora characterised by species typical of large clear-water lakes, such as *Aulacoseira islandica* and *Stephanodiscus neoastraea* in the pelagic zone and *Ellerbeckia arenaria* in the littoral zone. The Ancylus clays sometimes contain sulphide bands or stains, but this does not indicate the presence of a halocline. These sulphide traces are rather a secondary effect due to hydrogen sulphide diffusion from the overlying younger organic sediments (Sohlenius et al. 2001). In sediments from the open basin, the transition to the Ancylus Lake is visible as a small increase in the organic carbon content, an effect of the damming when old reworked carbon from the transgressed coasts was discharged and redeposited in the deeper areas (Andrén et al. 2000).

The Ancylus transgression, estimated to have raised the water level by 10 m above that of the ocean (Björck et al.

2008), is today visible as raised beach ridges along the coasts of Estonia, Latvia, the Swedish mainland and the island of Gotland. It is also documented by the simultaneous flooding of pine forest along the coasts of the southern Baltic basin. A new outlet to the ocean, called the Dana river, was created in the Storebælt area. This was a complex river system with various channels and lakes (Björck et al. 2008). After a few hundred years, the Ancylus Lake was at the level of the ocean as a result of the worldwide eustatic sea level rise caused by climatic warming and the subsequent melting of ice sheets. The end of the Ancylus Lake stage is defined by records of the first weak inflows of marine water into the Baltic basin.

2.5.9 The Initial Littorina Sea: between fresh and brackish

The next transitional phase with slightly brackish water conditions was originally named the “Mastogloia Sea” after the diatom genus *Mastogloia* (Box Fig. 2.7), but today it is known as the “Initial Littorina Sea” or the “Early Littorina Sea”. This stage lasted from ~9,800 until ~7,500 years BP. The transformation from the Ancylus Lake into a brackish sea did not proceed simultaneously in different parts of the Baltic basin. It started close to the narrow inlet in the south and gradually spread northwards during a period of ~2,000 years. This implies that while the southern part of the Baltic basin was already experiencing slightly brackish Initial Littorina Sea conditions, the central part still had an Ancylus Lake setting (Andrén 1999).

The inflowing marine water contributed to the development of a halocline and brought nutrients into the system, which resulted in pronounced cyanobacterial blooms as early as ~8,000 years ago (Bianchi et al. 2000). Distinctive of the Initial Littorina Sea stage is the combination of high cyanobacterial production and low diatom abundance in the open basin. However, in the coastal zones a diverse brackish diatom flora, the so-called “*Mastogloia* flora”, was established during this transitional phase (Miettinen 2002).

2.5.10 Hypoxia in the Initial Littorina Sea

The salinity stratification, together with the decreased oxygen saturation of the warmer water and increased primary production, initiated periods of deep-water hypoxia in the open Baltic basin. This is evident from findings of extended areas of laminated gyttja-clay deposits in the sediment record (Zillén et al. 2008). Increased upward transport of

phosphorus, released from the anoxic bottoms to the photic zone, has been suggested as an explanation of the enhanced primary production observed at the Ancyclus/Littorina transition (Sohlenius et al. 2001).

Diazotrophic cyanobacterial blooms have been proposed to have played a role as a trigger of eutrophication during this period. The stable nitrogen isotope is indicative of the origin of nitrogen (Box 2.3) and $\delta^{15}\text{N}$ measurements indicate that the bloom-forming cyanobacteria of the Initial Littorina Sea were actually fixing nitrogen (Borgendahl and Westman 2007). The sediments deposited during the Initial Littorina Sea show a sudden increase in the total organic carbon content from $\sim 1\%$ to 4–8 %. This is reflected in a characteristic ecosystem-wide change from the homogeneous Ancyclus clay to the homogeneous or laminated Littorina gyttja-clay (Winterhalter 1992).

2.5.11 The Littorina Sea: the most saline stage

The Littorina Sea (Fig. 2.27d), named after the common periwinkle *Littorina littorea* (Box Fig. 2.5), is the stage during the Holocene development of the Baltic basin with the largest marine influence. It is time-transgressive, *i.e.* it occurred at different times in different parts of the Baltic basin, and is estimated to have lasted from $\sim 7,500$ until $\sim 3,000$ years BP in the central part of the basin.

The current knowledge about the Baltic palaeo-salinity is mainly based on fossilised organisms such as diatoms and molluscs. Early attempts to reconstruct Baltic palaeo-salinities by using the minimum and maximum salinity tolerances of the mollusc fauna found in raised beach ridges date back to the late 19th century (*e.g.* Lindström 1886). The classical view of the palaeo-salinity in the Baltic basin during the Littorina Sea stage is that brackish conditions were established $\sim 7,500$ years BP and that the salinity increased until it reached a maximum level during $\sim 6,000$ – $4,000$ years BP, after which it gradually decreased down to the present level (Fig. 2.26c).

Attempts to quantify palaeo-salinity by using species distribution optima (*e.g.* molluscs, diatoms), species morphology (spine length of dinoflagellate cysts) or measurements of pore water and stable isotopes (oxygen, carbon, strontium, Box 2.3) have produced varying results. At present, there is no consensus as to the exact magnitude of the past salinity changes. However, we do know that the horizontal salinity gradient in the Baltic basin was not as pronounced during the Littorina Sea as it is today because the sills separating the subbasins were situated deeper

(Gustafsson and Westman 2002). The halocline was probably found at a similar water depth as it is today.

2.5.12 The Holocene Thermal Maximum and the Littorina Sea

The Holocene Thermal Maximum in northwestern Europe during $\sim 8,000$ – $4,000$ years BP was characterised by high temperatures and low humidity (Seppä et al. 2009). The warm climate led to a high global eustatic sea level, which increased the depth of the sills between the ocean and the Baltic basin and enhanced the inflow of marine water (Gustafsson and Westman 2002). The high salinity strengthened the halocline, which together with declining oxygen saturation in a warmer sea created hypoxic bottoms that released phosphorus. This in turn increased the abundance of diazotrophic cyanobacterial blooms (Sohlenius et al. 2001).

Models of the fluctuating salinity in the Littorina Sea until present show that freshwater supply has been the main driver, and that sill-depth changes in the transition zone only partly contributed to the variations in salinity (Gustafsson and Westman 2002). The low humidity during the thermal maximum resulted in higher evaporation and decreased freshwater discharge, resulting in higher salinity in the Baltic basin.

2.5.13 Shoreline displacement in the Littorina Sea

In the coastal zone, the shoreline displacement (Fig. 2.26d), and consequently the configuration of the Baltic basin, was affected by both eustatic and isostatic components. The Littorina stage is characterised by a fluctuating water level, but the number and magnitude of transgressions on the Baltic coasts varied from one location to another due to a higher isostatic rebound in the northern part of the basin where the ice sheet had been thicker and the duration of glaciation longer (Pirazzoli 1991).

As a result of shoreline oscillations, the sediments deposited in the coastal zone were resuspended and thereby they contributed to nutrient cycling. This made the coastal zone of the Littorina Sea a highly dynamic environment, which was naturally rich in nutrients. In near-shore areas, the diatom flora was very diverse and consisted of the so-called “*Clypeus* flora”, named after the diatom *Campylodiscus clypeus* (Box Fig. 2.7), also known as “the lagoonal flora”. This diatom flora is indicative of shallow nutrient-rich bays, which were gradually lifted up due to the isostatic rebound,

and is common along the Finnish and Swedish coasts (Miettinen 2002).

Slowly, the Littorina Sea turned into the Baltic Sea of today, also known as the “Mya stage”, named after the bivalve *Mya arenaria* (Box Fig. 2.5), whose appearance is considered as an early example of human-mediated species introductions to the North Sea-Baltic Sea Area (Strasser 1999).

2.5.14 Laminated sediments, palaeo-production and hypoxia

Sediment cores that cover the whole time-span from the Littorina Sea until the present Baltic Sea are characterised by homogeneous gyttja-clays interlayered by laminated gyttja-clays. The homogeneous layers are indicative of oxygenated bottoms with burrowing animals, which perturb the seasonal deposits. Laminated sediments are formed when seasonal sedimentation is not disturbed by animal life since animals are absent due to oxygen stress. Therefore, laminated sediments are used to trace past bottom-water hypoxia (Zillén et al. 2008).

The laminated sediments of the Baltic Sea have a high organic carbon content, which could be interpreted as a sign of a high primary production. However, it has been debated whether a high organic carbon content is exclusively coupled to changes in primary production or whether it may also partly be due to a better preservation of carbon under anoxic conditions (Sohlenius et al. 2001). Enhanced preservation of carbon could explain why variations in the geographical distribution of laminae are correlated with water depth (Zillén et al. 2008). Thus, the formation of laminae could partly be triggered by increased stratification due to increased salinity in a warmer climate, and not exclusively by increased primary production.

Apart from the laminated sediments that are deposited today, the sediment record of the open Baltic Sea shows the presence of laminated sediments deposited during two previous periods. The first period occurred ~8,000–4,000 years BP and correlates with the Holocene Thermal Maximum (Zillén et al. 2008). The pelagic microfossils in these sediment layers show a high abundance of resting stages of the diatom genus *Chaetoceros* and the silicoflagellate *Dictyocha speculum*, both of which are indicators of high nutrient concentrations. Simultaneously, diatom taxa indicative of warmer and more marine conditions occur as well, e.g. *Pseudosolenia calcar-avis* (Box Fig. 2.7), *Thalassionema nitzschioides*, *Thalassiosira oestrupii* and *Chaetoceros mitra* (Andrén et al. 2000).

The second period with laminated sediments is dated to ~2,000–700 years BP and correlates with the Roman Warm Period around 2,000 years BP and the Medieval Climate Anomaly around 1,000 years BP (Zillén et al. 2008; Seppä et al. 2009). The microfossil record from the Medieval Climate Anomaly contains resting stages of the diatom genus *Chaetoceros* and the ebridian *Ebria tripartita*, both indicating high nutrient concentrations (Andrén et al. 2000). During this warm event, the pelagic diatoms were dominated by *Pseudosolenia calcar-avis*. This species gradually disappeared with the succeeding cold event known as the “Little Ice Age” (~500–100 years BP), and the diatoms shifted to the dominance of *Thalassiosira hyperborea* var. *lacunosa* and varieties of *Actinocyclus octonarius*.

2.6 A changing ecosystem

2.6.1 Fast changes and regime shifts

Over the past 1,000 years the Baltic Sea has changed dramatically from the hypoxic Medieval Climate Anomaly to the oxic Little Ice Age to the present day, again with hypoxia. These changes were primarily climate-driven, without any major interferences by humans. Today the large human population living in the Baltic Sea drainage area exerts strong pressures on the sea and its well-being increasingly depends on how it is managed (cf. Sect. 18.5).

The Baltic Sea ecosystem is not the same as it was 100 years ago, i.e. with the same species composition, food webs and productivity, and this could be expected of a dynamic ecosystem. However, the changes caused by anthropogenic pressures are large and take place very fast. Some of the large-scale changes that are listed as major threats to the diversity of the Baltic Sea ecosystem are multifaceted. While there is a risk with each non-indigenous species invading the Baltic Sea, some have enriched its functional diversity. While eutrophication has many negative effects on the ecosystem, it does increase fish productivity. While climate change compromises the existence of the cold-water species in the Baltic Sea, it increases the length of the growing season and increases productivity. The Baltic Sea is one of the large marine ecosystems with the highest recorded temperature increases during the past century (Belkin 2009). This temperature increase is consistent with the anthropogenic climate change signal (Bhend and von Storch 2009; Rutgersson et al. 2014). The global warming of the Earth’s surface since the 1980s has been highest at latitudes above 50 °N (Fig. 2.28) where the Baltic Sea is situated between latitude 53°55’ N and 65°48’ N.

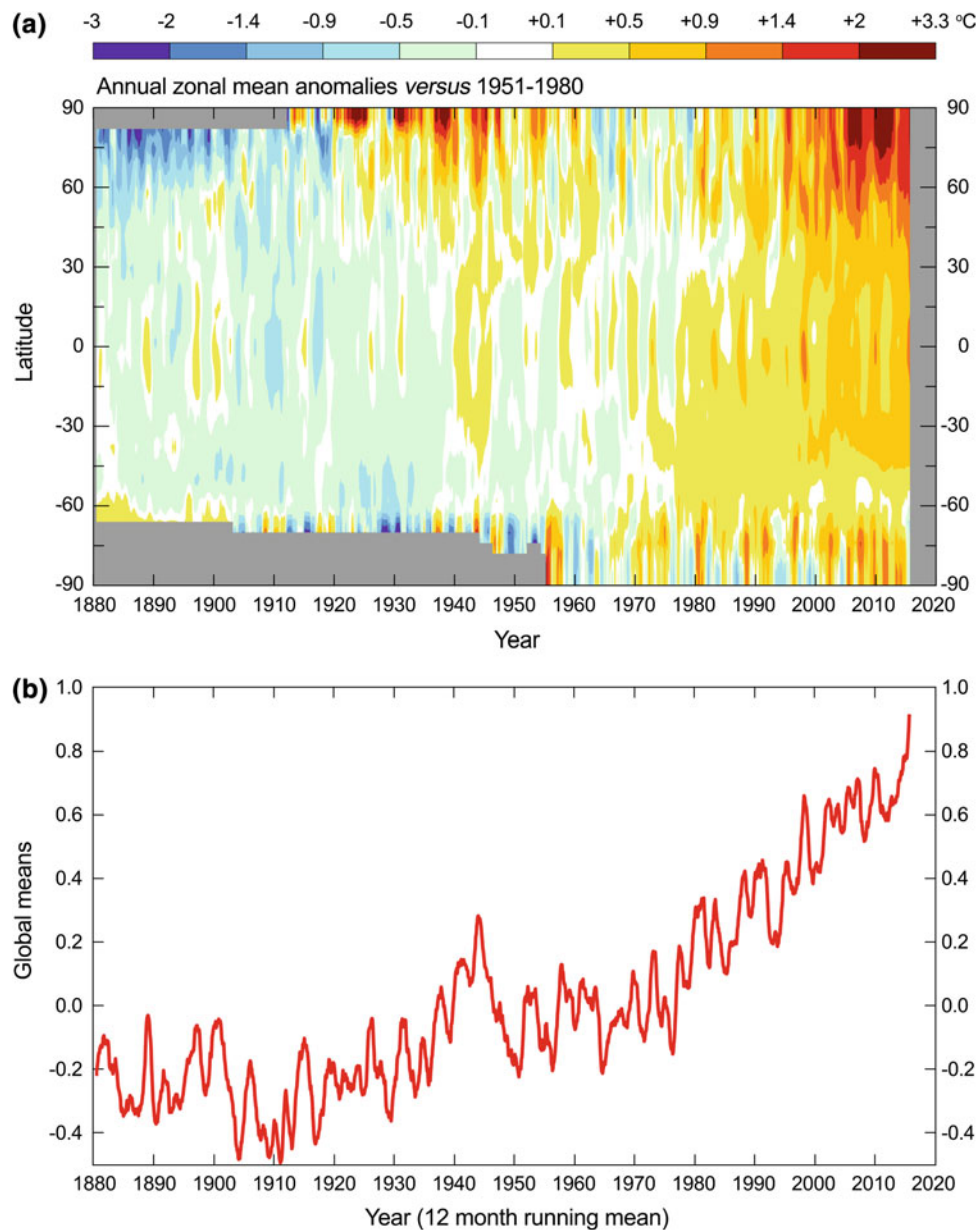


Fig. 2.28 Summary of global warming based on the Earth surface temperature Land-Ocean Temperature index (ERSST v4), whereby the temperature for each year is compared with the mean temperature for the years 1951–1980. (a) Annual zonal mean anomalies from 1951 to 1980. (b) Global mean anomalies from 1951 to 1980. Figures printed with permission from the National Aeronautics and Space Administration (NASA), USA (http://data.giss.nasa.gov/gistemp/time_series.html)

One of the typical features of the Baltic Sea ecosystem is that at least the northern part is ice-covered in winter. With climate change the ice cover extension, duration and thickness will decrease, which is expected to lead to significant changes in the ecosystem. The ice cover affects physical processes, *e.g.* the water masses are less affected by winds and the ice cover changes air-sea heat fluxes. An example of ice cover impacts on biological processes is the timing of the

onset of the pelagic spring bloom and its species composition (*cf.* Sect. 9.6.4).

Many environmental and biological factors interact and it is not possible to understand ecosystem change by monitoring one or a few factors only. An integrated, simultaneous analysis of many factors shows that they change in concert, and that some change more than others (Box 2.5). Powerful instruments for understanding ecosystem functioning and

Box 2.5: Regime shifts

Threshold-like ecological shifts

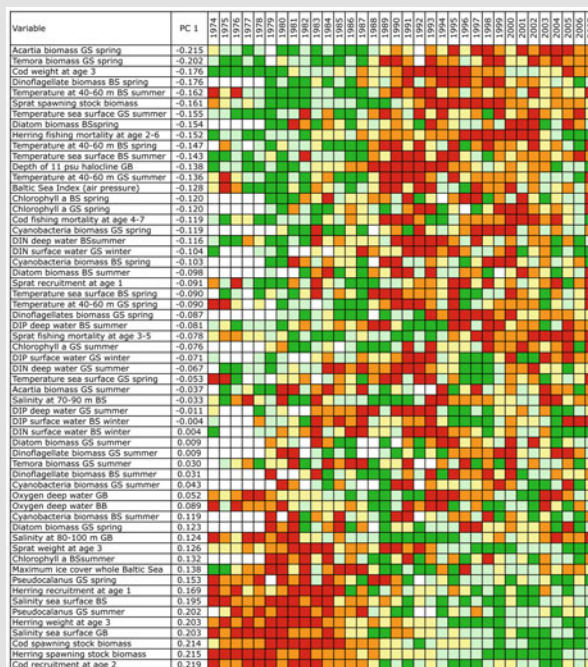
Abrupt and rapid shifts in food web and community structure, so-called “regime shifts”, are increasingly being reported for large marine ecosystems around the world (Kraberg et al. 2011). Such a threshold-like ecological shift occurred in the pelagic system of the Baltic Sea proper at the end of the 1980s and changed the composition of the zooplankton and fish communities (Box Fig. 2.8). Almost synchronous shifts were recorded in the other subbasins of the Baltic Sea (Diekmann and Möllmann 2010), as well as in the North Sea and some other seas in the northern hemisphere (Möllmann and Diekmann 2012). These changes could partly be coupled to the North Atlantic oscillation (NOA) and other climatic phenomena, which modified local temperature regimes. Human-induced trophic cascades, triggered by the removal of predators such as seals and cod (Österblom et al. 2007; Casini et al. 2009), as well as eutrophication and introductions of non-indigenous species, are also coupled to regime shifts as they create possible “tipping points” for food web functioning. Multiple drivers potentially interact in a way that one driver (e.g. overfishing) undermines resilience and another one (e.g. climate change) provides the final impulse for an abrupt change (Möllmann and Diekmann 2012).

Wasp-waist trophic structure

Like many other aquatic ecosystems, the Baltic Sea exhibits a characteristic so-called “wasp-waist” trophic structure (Cury et al. 2000; Bakun 2006), in which one or a few species of small planktivorous fish entirely dominate their trophic level where bottom-up and top-down processes meet. In the Baltic Sea, the cod collapsed during the 1980s in concert with the establishment of the ecosystem’s dominance by the cod’s wasp-waist prey, the sprat.

The future of the Baltic Sea ecosystem is in our hands

The future of the Baltic Sea ecosystem is largely down to how it will be managed even if climate change sets some limits to possible outcomes. Two different future cod fishing and eutrophication scenarios were investigated by combining three regional biogeochemical models with an Ecopath model including the Ecosim food web procedure by Niiranen et al. (2013). The results of this modelling study showed that by the end of the 21st century, the combination of intensive cod fishing and high nutrient loads projected a strongly eutrophicated sprat-dominated sea, whereas low cod fishing in combination with low nutrient loads would result in a cod-dominated system with eutrophication levels close to present.



Box Fig. 2.8 Regime shifts can only be detected in data sets that include relevant measurements and cover a substantial period of time including data from before, during and after a potential regime shift. This figure shows a “traffic-light plot” of the temporal development of environmental and biological variables in the Baltic Sea proper from 1974 to 2007, and documents a sudden “regime shift” in the end of the 1980s. GS = Gotland Sea, BS = Bornholm Sea. Dark green = very low values, light green = low values, yellow = intermediate values, orange = high values, red = very high values. The variables are sorted according to their scores on the first axis (PC1) of a principal components analysis. Figure modified from Diekmann and Möllmann (2010)

predicting future changes in the Baltic Sea ecosystem are being developed, *e.g.* by the coupling of oceanographic, climatic and biological models (Box 2.6).

Several recent studies have suggested that the on-going changes in the Baltic Sea ecosystem form a directional trend with a sudden “regime shift” in the end of the 1980s (Box 2.5, *cf.* Sect. 17.2.4), rather than being the result of normal inter-annual fluctuations (Österblom *et al.* 2007; Möllmann *et al.* 2009). Regime shifts are abrupt, substantial and persistent changes in the state of natural systems. The 1980s regime shift was recorded not only in the Baltic Sea and the North Sea (Alheit *et al.* 2005), but on a planetary scale, and can be explained by changing climatic factors through interactions between major volcanic eruptions and anthropogenic climate change as the main forcing factors (Beaugrand *et al.* 2015; Reid *et al.* 2016). Regionally, other environmental drivers may coincide and interact with global warming. In the case of the Baltic Sea, the 1980s regime shift was also strongly related to an overfishing-induced trophic cascade from a cod-dominated to a sprat-dominated food web in the pelagic zone (Möllmann and Diekmann 2012).

2.6.2 How to turn negative trends

A scenario with intensive cod fishing and high nutrient loads projects a strongly eutrophicated sprat-dominated sea by the end of the 21st century, whereas a scenario with low cod fishing in combination with low nutrient loads is expected to result in a cod-dominated system with eutrophication levels close to present (Box 2.5). Although ecosystem-based management is the agreed principle today, in practice the various environmental problems are still handled separately, since we still lack both basic ecological knowledge and appropriate governance structures for managing them together, in a true ecosystem approach (Elmgren *et al.* 2015). Proper ecosystem-based management is mandatory for maintaining a well-functioning ecosystem that can provide goods and services to human society in a sustainable way (*cf.* Sect. 18.5). Modelling results have shown that regional management is likely to play a major role in determining the future of the Baltic Sea ecosystem (Box 2.5, Niiranen *et al.* 2013).

Some improvements of the Baltic Sea environment have been achieved as a result of changes in the management of the sea’s resources through actions such as financial investments, legislation and international agreements. As a result of the reduction of atmospheric nitrogen emissions (*e.g.* from traffic) and the reduction of phosphorus emissions by building a large number of modern wastewater treatment plants (especially in the eastern part of the drainage area), nutrient inputs to the Baltic Sea have declined since 1980 (Box Fig. 2.2). Some species that previously were endangered by contaminants and/or hunting, such as the white-tailed eagle *Haliaeetus albicilla* (*cf.* Box Fig. 4.19), the common guillemot *Uria aalge* (*cf.* Fig. 16.2), the grey

seal *Halichoerus grypus* (*cf.* Box Fig. 4.21b) and the ringed seal *Pusa hispida* (*cf.* Box Fig. 4.21c) have recovered thanks to legislation forbidding the use of certain chemicals (*cf.* Sect. 16.1) and protecting them as red-listed species (*cf.* Sect. 18.6.2). Fisheries regulations are the main reason why the eastern Baltic cod stock has shown some signs of recovery after more than two decades of low biomass and productivity (Cardinale and Svedäng 2011). This recovery was mainly driven by a sudden reduction in fishing mortality despite the absence of any exceptionally large year classes.

These examples show that proper management can have positive effects on ecosystem health on a relatively short time scale. Necessary actions include decreasing nutrient inputs, chemical pollution bans, habitat protection and control of fisheries, so that key functions of the ecosystem can be operational and the Baltic Sea can be a fully functional ecosystem delivering ecosystem goods and services to society. Since the Baltic Sea is surrounded by countries that live in peace and are rather wealthy in comparison with many other areas in the world, the Baltic Sea countries have the potential to invest in adequate ecosystem-based management, *i.e.* to set clear operational goals focused on long-term ecological sustainability (Thrush and Dayton 2010). Also the political willingness to improve the Baltic Sea environment seems to be present, *e.g.* through well-organised international meetings at different levels. However, in practice the necessary governance processes are unfortunately remarkably slow while a variety of pressures on the sea, from ship traffic to climate change, are increasing.

2.6.3 Will we overcome the “tragedy of the commons”?

The “tragedy of the commons” (Hardin 1968) is a theory that emerged in the 1960s from the growing concern about the rapid human population growth on Earth. The subtitle of Hardin’s paper in the journal “Science” is “The population problem has no technical solution; it requires a fundamental extension in morality”. According to this theory each human individual acts rationally and independently, conforming to his/her own interests, in contrast to the general long-term interest of society, by depleting common resources. Pollution of the environment is one of the examples that Hardin raised to illustrate his theory.

Today we know even better than 50 years ago that environmental resources, like clean water and clean air, are not endless. However, while public awareness has grown, the morality of the average human individual may not have changed much, as the theory claims. The “tragedy of the commons” is also applicable to countries sharing a common resource, *e.g.* when negotiating about fishing quota and nutrient emissions. The Baltic Sea is such a common resource that has been overexploited and polluted for a long time, mainly by its nine riparian countries.

Box 2.6: Modelling coastal seas

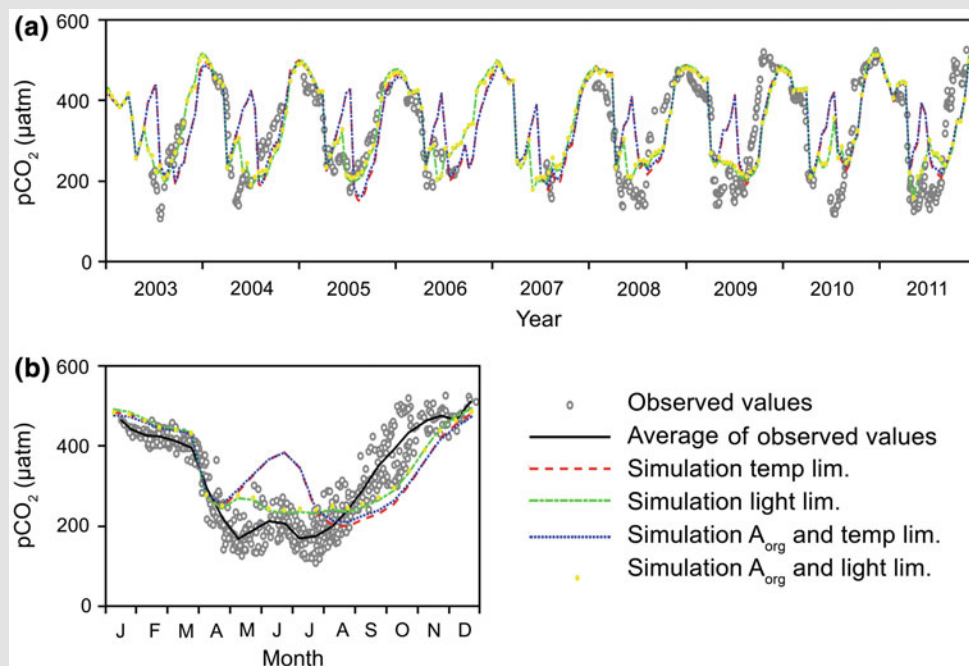
Anders Omstedt

Computational fluid dynamics

The use of computational fluid dynamics (CFD) to analyse and predict environmental changes has increased considerably in recent decades. Mathematical models are now standard tools in research, as well as in a wide range of practical applications. Intensifying concern about human influence on climatic and environmental conditions has increased the need for multidisciplinary modelling efforts, including numerical modelling of oceans, lakes, land surfaces, ice, rivers, and the atmosphere. Scientists have traditionally developed specialised models limited to application within their own disciplines. Today, increasing efforts are being made to develop Earth System models that include major processes that one needs to consider when dealing with climate change and other environmental changes. In general, the models rely on conservation laws, including many processes that are not known in detail.

Parameterisations

These rather unknown processes then need to be parameterised in different ways. For example, turbulence, which is always present in coastal seas, is poorly understood and needs to be parameterised in the models. This is also true for a number of chemical and biological processes. For example, there is no standard parameterisation yet available for biological processes such as plankton growth and ecosystem change, and instead most parameterisations are based on some available observations. However, one can regard the new and updated models as a systematic collection of the present available knowledge. The models can therefore help us to identify gaps in understanding and where new research programmes need to be developed. CFD cannot, however, profess adequately without reference to experimental and field validation. This is also a good reason why models, field experiments and monitoring programmes need to be strongly linked to each other. Coastal sea models not only include codes for solving the conservation laws, they also include initial data (*e.g.* salinity and temperature) and forcing data (*e.g.* weather conditions, riverine runoff, atmosphere and land emissions from nutrients and carbon components).



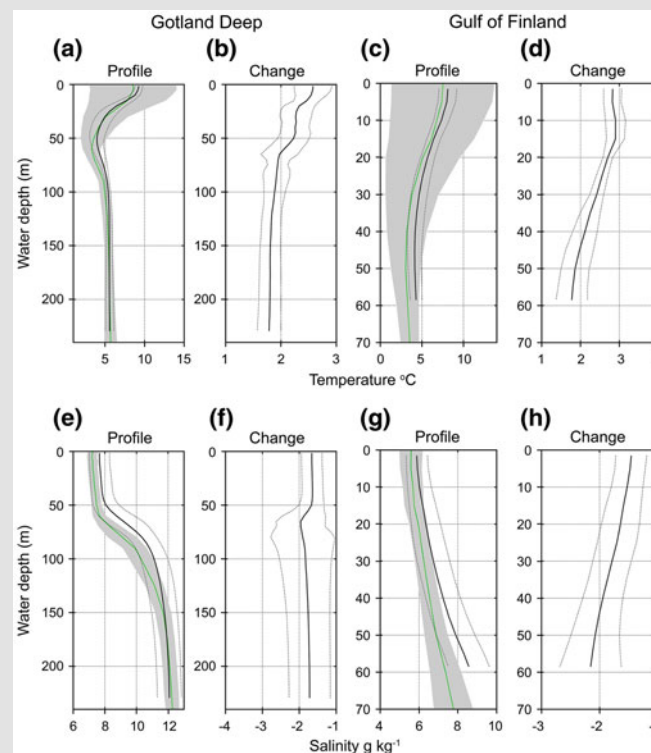
Box Fig. 2.9 Variation of pCO₂ in the surface water of the Gotland Sea (Baltic Sea proper). (a) Observed and simulated pCO₂ during nine years (2003–2011). (b) Observed and simulated seasonal average pCO₂ in 2003–2011. The red lines indicate the results from a model simulation with the original parameterisation for growth of cyanobacteria (“temperature limitation”). The green lines indicate the results from a simulation with the new parameterisation for growth of cyanobacteria (“light limitation”). The blue lines and yellow dots indicate the results from simulations including organic alkalinity (A_{org}) with temperature and light limitation, respectively. Figure modified from Gustafsson et al. (2015)

Different types of Baltic coastal sea models

The available coastal sea models vary from simple box models to coupled three-dimensional atmosphere-land-ocean models. They are often developed for different applications and use different kinds of forcing fields. One class of models is process-based models (*e.g.* Omstedt 2015) and another class is three-dimensional models (*e.g.* Meier et al. 2006). Both types of models have been used in many Baltic Sea applications. The strength of process-based models is here exemplified by a study that analyses CO₂ dynamics in the Baltic Sea (Gustafsson et al. 2015). In this study, the modelling focused on how air-water CO₂ fluxes respond to parameterisations of organic alkalinity, gas transfer, phytoplankton growth, and changes in river loads. The forcing data in the study included the most complete compilation of Baltic river loads for dissolved inorganic and organic carbon (DIC and DOC), as well as total alkalinity. One result demonstrated how air-water CO₂ fluxes depend on the river load of carbon. If the river load of total alkalinity decreases, the CO₂ buffer capacity is reduced, which has the effect that the partial CO₂ pressure (pCO₂) in the water increases and changes the air-water CO₂ fluxes. By analysing different aspects in the modelling, the calculations were getting closer to the observations. For example, a new parameterisation of cyanobacteria (removing strict temperature dependence and instead letting the growth of cyanobacteria be more strongly controlled by light intensity) significantly improved the seasonal development of pCO₂, although the values were overestimated in summer and underestimated in autumn (Box Fig. 2.9).

Model comparisons

The accuracy of Baltic Sea models to produce biogeochemical parameters was compared by Meier et al. (2012). This is one of the first studies that develop the concept of adding the results from different models together in a so-called “ensemble calculation”. In general, the results show that the models were not too far from the observations, and that the best representation of the data was the ensemble mean (Box Fig. 2.10). Thus the community of scientists working with models now enters a new era, as not a single model should be used to give management advice. Instead, a number of different models needs to be used. This is in line with the modelling developments within global climate change, illustrating that management actions in the future need to be closely linked to assessment activities.



Box Fig. 2.10 Vertical profiles and predicted changes in temperature and salinity in the Gotland deep (left panels) and the Gulf of Finland (right panels). (a, c, e, g) Profiles for the control period 1978–2007, showing the average (green line) ± 1 standard deviation (grey shaded area) of observations, and the ensemble average of the results of different models (black line) ± 1 standard deviation (dotted line). (b, d, f, h) Ensemble average changes between 1978–2007 and 2069–2098, showing the predicted increase of temperature and the predicted decrease in salinity for the Baltic Sea by the end of the century. Figure modified from Meier et al. (2012)

Review questions

1. What factors influence the salinity of the Baltic Sea?
2. How does geography affect the sensitivity of the Baltic Sea ecosystem to external impacts?
3. What were the causes of the changes between the different stages of the Baltic basin during its late- and post-glacial environmental history?
4. How and why has the geographical extent of hypoxic sea bottoms varied through times?
5. Summarise the conditions that make the Baltic Sea of today such a special environment for different types of species.

Discussion questions

1. What do the Baltic Sea, the Black Sea and the Caspian Sea have in common, and how is this related to their geography and geological developments?
2. What do the Baltic Sea and the Mediterranean Sea have in common with respect to water exchange with the ocean? What are the differences?
3. In what way is knowledge about the geological development of the Baltic basin relevant when discussing contemporary environmental problems?
4. What do you think the Baltic Sea would be like today if no anthropogenic impact had occurred for the last 10,000 years?
5. Will we overcome the “tragedy of the commons”? How?

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