Hypoxia Controlled by Hydrodynamics

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Abstract
In summer dissolved oxygen is often depleted in the lower and bottom layers in many coastal basins all over the world. This phenomena is called hypoxia. When the oxygen consumption exceeds oxygen supply, the water becomes hypoxic. The oxygen is consumed by decomposing organic matter by bacteria (biochemical processes), while the oxygen is supplied by physical processes such as convection, advection and diffusion. The primary cause of hypoxia is the consumption of oxygen in the water column, but physical processes mainly control its generation, distribution and configuration. In addition to the vertical supply of oxygen by mixing, horizontal transport by estuarine circulation plays the major role in the formation of hypoxia in regions of freshwater influence. As the hypoxic water contains a lot of nutrients, it plays an important role for primary production, producing middle layer chlorophyll maximum in summer and inducing bloom of phytoplankton in autumn.

1. General introduction
Oxygen is essential for almost all marine biota, including fishes and invertebrates, to maintain their life. However, the amount of oxygen diluted in the water is limited. The saturation rate of oxygen in the water is only 5.2 mL L\(^{-1}\) (=7.4 mg L\(^{-1}\)) under 1 atom at 20°C. This concentration is significantly lower than that in the air (210 mL L\(^{-1}\)). Therefore, marine animals develop advanced gills to take in oxygen efficiently from the seawater. In spite of the advanced organ intrinsic to marine animals, significant decrease of dissolved oxygen (DO) in the water damages them. DO concentration sometimes reduces seriously to the level which has harmful effects on marine animals especially in summer. This water mass is called hypoxia or hypoxic water. The water including nearly zero amount of oxygen is called anoxia.

Oxygen depletion exerts a serious impact on marine ecosystems, although the tolerability of marine animals is different among the species. For instance, fishes such as red sea bream and yellowtail are going to die within a few days by exposure to the water with 3 mg L\(^{-1}\) of DO (Ishioka 1982; Yamamoto et al. 1990). In general, oxygen deficiency lower than 4 mg L\(^{-1}\) exerts a baneful influence upon cultured fish (Inoue 1998). On the contrary, benthic animals tend to be tolerant to low oxygen. Starfish and brittle stars are stressed when DO drops below 1.5 mg L\(^{-1}\), and are found dead when there is less than 1 mg L\(^{-1}\) (Simpson and Sharples 2012). Some bivalves can survive over one week even if DO concentration is less than 1.5 mg L\(^{-1}\). However, even those low oxygen tolerant animals decrease their activities in hypoxia, and all marine animals becomes debilitated and cannot survive in anoxic water.

Animals that have sufficient swimming ability to control their position vertically within the water column (e.g., fish, demersal invertebrates such as shrimp and crab) attempt to leave the hypoxic region in the lower layer. Animals that have less ability to leave the region of low DO gradually become stressed and, if concentrations of oxygen drop low enough, weaken and finally die. It has been reported that hypoxia often causes problems in many eutrophic estuaries and coastal areas (e.g. Borsuk et al. 2001; Hagy et al. 2004; Gilbert et al. 2005). More than 400 hypoxic waters have been reported in many coastal areas, including Baltic Sea, Mexico Bay and various estuaries. Ise Bay is highly eutrophic and is famous for its bottom hypoxia in summer (Kuno 1996), and the oxygen depletion severely damages the ecosystem and fisheries in the bay (Hossain and Sekiguchi 1996). In Mikawa Bay, the number of benthic animals seriously decreases in the DO concentration under 3 mg L\(^{-1}\) (Suzuki 1998). In...
the Seto Inland Sea, there are some regions including Osaka Bay, Harima-Nada, Hiuchi-Nada, Hiroshima Bay, Suo-Nada and Boppu Bay, where hypoxia occurs every summer (Fig. 1a). The diversity of species of macro benthic animals is low in the hypoxic regions (Fig. 1b). One of the most wide-spread hypoxia is observed off the coast of Louisiana and Texas, USA (Rabalais et al. 2002). The survey of the region shows an area of about 17,000 km² experiencing hypoxia, which leads to large changes in bottom water marine life. This region is called the “dead zone” because of the failure to catch demersal fish and benthic animals.

Not only with the objective of environments, but also fisheries are of course damaged by hypoxia, especially in enclosed euphotic bays and lakes. Demersal fishes, crabs and shellfishes are rarely observed in the bay head of Mikawa Bay in summer. Shijimi clam fisheries in Lake Shinji and Lake Ogawara, both of which are foremost Shijimi fisheries brackish lakes in Japan, are restricted in the narrow coastal areas which are shallower than 5 m depth, because the deep central areas become hypoxic in summer and shijimi clams cannot survive.

Fig. 1. (a) The position of hypoxic water in the bottom layer in summer, (b) diversity index of macro benthos and (c) the distribution of $\log_{10}(H/U^3)$, where $H$ is the depth of water and $U$ is the amplitude of M2 tidal current (after Yanagi 1990) in the Seto Inland Sea. Reprinted from Yanagi, Science of tidal front, 169 pp., © 1990, with permission from Koseisha-Koseikaku.
there. It is therefore crucially important to clarify the formation mechanism of hypoxia and to develop a scheme to reduce it.

2. Formation mechanism of hypoxia

In general, both physical and biochemical processes control the generation of hypoxia. The oxygen supply is mainly determined by the physical processes such as water exchange between the water mass and the surrounding water. DO is usually saturated in the sea surface, because oxygen is supplied into seawater from the air in the process of turbulence and mixing by windswell and/or strong currents. As phytoplankton releases oxygen through photosynthetic activities, the oxygen in the surface layer sometimes over saturates in the surface and subsurface layers, especially in blooming seasons. In the deep layer, on the other hand, oxygen is not supplied from the air and photosynthesis would be weak because of insufficient irradiation. Both physical and biological processes therefore cannot supply oxygen in the deep layer.

Oxygen consumption is mainly determined by biochemical processes concerning bacterial activity. Carcass of phytoplankton produced in the euphotic layer gradually sinks, as well as other organic matter such as terrestrial matter. Oxygen is consumed as bacteria decompose organic matter during sinking. Not only small organic particles, which sink slowly, but also the larger particles such as carcass and feces of zooplankton sink faster and accumulates at the bottom. Oxygen is subsequently consumed when the accumulated organic matter is decomposed and/or consumed by other biota. The oxygen consumption exceeding the supply reduces DO concentration in the water, and accordingly leads the water hypoxic.

The vertical supply of oxygen has been mainly focused for the formation and demise of hypoxia in the conventional theory. In spring and summer, strong irradiation warms surface water and makes the water density smaller in the surface than in the bottom (stratification, Fig. 2a). The mixing between the oxygen rich upper water and bottom water is thus limited when the water is stratified. The bottom water becomes hypoxic, because oxygen consumption is larger than the supply (Fig. 2b). In autumn, in contrast, the air cools sea surface, making the surface water denser. This leads to mixing between the surface and lower layer, supply- ing oxygen to the lower layer from the surface. The hypoxia disappears when the whole water column is well mixed in the late autumn or winter. This process is often observed in fjords and lakes, where horizontal currents are weak in the bottom layer.

On the other hand, recent studies have shown that horizontal currents play an important role in the variation of oxygen concentration in coastal areas (Takahashi et al. 2000). Within the regions where freshwater flows in, buoyancy input into the bay head is responsible for producing a physical regime that is radically different from lakes and open ocean (Simpson 1997). It is well known that less saline lighter water flows seaward in the upper layer, while more saline heavier water flows landward in the lower layer (Hansen and Rattray 1966). This flow pattern is called “estuarine circulation” and often observed in estuaries and bays, as the water exchange between coastal basins and adjacent ocean has long been a topic of great interest to oceanographers (e.g., Geyer and Cannon 1982; LeBlond et al. 1991; Allen and Simpson 1998).

If the estuarine circulation is dominant, bottom water is readily renewed by oceanic water. This indicates that oxygen would be supplied to the bottom layer and hypoxia hardly happens. The hypoxia is, nonetheless, ubiquitous in many bays all over the world.

Recent surveys have shown that the circulation does not always follow the classical estuarine circulation pattern in regions where the stratified area (wide bay area) located next to the mixed area (narrow strait). The flow pattern in those regions has a bimodal character: bottom inflow of the mixed water in the cooling season, while mid-layer inflow in the heating season (Kasai et al. 2000, 2002; Takahashi et al. 2000; Fujiwara et al. 2002). In winter, the temperature is reversed and weak stratification is maintained only by
freshwater buoyancy input inside the bay (Fig. 3a). Sea surface cooling causes the freshwater to sink through the mixing process, so that the water in the bay is lighter than that at the strait. This density difference leads to bottom inflow in the cooling season, following the conventional estuarine circulation pattern. During summer, on the other hand, the effect of surface heating is restricted to the upper layer by the strong thermocline and halocline in the stratified area (inside the bay), while it extends to the bottom due to the strong tidal currents in the well-mixed area (strait). This process causes the density of the bottom bay water to be greater than that of the strait water. The well-mixed strait water thus has a density equivalent to the middle layer inside the bay, and then tends to flow not through the bottom layer but through the middle layer, when it intrudes into the bay in the heating season (Fig. 3b, Takahashi et al. 2000; Kasai et al. 2002). The bottom water in the stratified area is therefore a relic spring water, which temperature is lower than the surrounding water. This process limits oxygen supply to the relict water, making it hypoxic in summer.

3. Three-dimensional circulation and hypoxia in a coastal embayment

3-1. Introduction

Ise Bay is one of the major coastal embayments on the Pacific coast of Japan (Fig. 4). It is ~30 km wide and 60 km long, and has a mean depth of ~20 m with the deepest longitudinal depression of about 35 m depth in the middle. Three large rivers (Ibi, Nagara, Kiso Rivers) flow into the bay head in the north, while in the south the bay opens to the Pacific Ocean via the narrow Irago Strait, which has a width of ~10 km. Nutrient and organic loading from the rivers results in serious eutrophication, especially at the bay head. Inside the bay, the buoyancy input by freshwater and weak tidal currents (~10−1 m s−1) cause the water to stratify. Based on the stratification-circulation diagram by Hansen and Rattray (1966), the water condition is classified as a strongly stratified regime in summer, and a weakly stratified regime in winter. Both large freshwater discharge and sea surface heating reinforce the stratification in summer. On the other hand, stirring by strong tidal currents in the strait (≥1 m s−1) promotes strong vertical mixing of the oceanic heavy water with relatively lighter water from the bay to produce intermediate density water. The coexistence of stratified and mixed regions complicates the circulation in this system (Kasai et al. 2002).

The oxygen in the bottom water is often depleted in the stratified season, although the bay has no sill in its mouth (Fig. 4). It is reported that shellfish fisheries are strongly damaged by the hypoxia in summer especially in the western basin of the bay. An example of the longitudinal distribution of hydrographic and DO conditions is shown in Fig. 5. A significant thermocline between 10 and 15 m separates the cold water from the upper layer in the bay. Some of the isotherms (T = 16–18°C) bend down and create a sharp bottom front at the bay mouth, while other (T = 20–21°C), along with isohalines of 30–32, are curved upwards to the surface. In contrast to the stratification inside the bay, the water column in the Irago Strait is strongly mixed. The distribution of DO was similar to that of temperature. There is a strong DO front at the bay mouth and the DO level in the cold dome is substantially reduced with a minimum <10% of saturation. This strong depletion of DO within the dome indicates that the bottom water was isolated from the surrounding water masses, in a similar way to the domes on the shelf regions or fjords. These hydrographic and DO features resemble the schematic views shown in Fig. 3b.

However, the formation mechanism of hypoxic water, namely the cold dome, in Ise Bay is different from that on the shelf or fjords because we are dealing with a system in which buoyancy induced by freshwater input is an important part of the problem.
Conventional theory on regions of freshwater influences, two-dimensional (vertical-longitudinal) estuarine circulation develops with upper layer freshwater flowing seaward while lower layer salty water flows landward. If this estuarine circulation were dominant, the inflow water would replace the bottom water so that the hypoxia would not persist. The dynamics are further complicated by the influence of the Earth’s rotation if the horizontal scale of a bay is large compared with the internal Rossby radius of deformation (Kasai et al. 2000). Ise Bay, for example, has a width of ~30 km comparing with 2--7 km of the radius. In addition, many drowned river valley estuaries and bays have complicated topography, which changes the circulation pattern (Wong 1994; Wong and Münchow 1995; Valle-Levinson and Lwiza 1995), so that the current structure in bays is essentially three-dimensional. Figure 5 indeed indicates that the prominent bottom front would prevent the intrusion of the oceanic water into the lower layer, and that circulation in Ise Bay would be different from the classical estuarine circulation. The movement of the intruding oceanic water is thus essentially important for the formation of the cold dome.

In this section, therefore, we show the mechanism responsible for the formation of the cold dome and hypoxia. Focusing on the cold dome and the intruding oceanic water, detailed observational results of three-dimensional temperature, salinity and velocity fields in Ise Bay are presented. Based on the observational results, the principles which govern the formation of hypoxia are demonstrated using a numerical model applied to an idealized condition. The key point for understanding the mechanism is the existence of a region of strong vertical mixing which maintains a well mixed condition next to a stratified bay.

3-2. Cold hypoxic dome in Ise Bay

The detailed observations were carried out on August 28--30, 1995, along three east-west lines (Fig. 4). Observed temperature, salinity, density and residual current distributions across the section are shown in Fig. 6. The panels show views looking up-estuary, with east being to the right. In the figure, the distributions indicate that all transects were strongly stratified. Temperature predominantly controls density in the lower layer, although both salinity and temperature contribute to the density structure in the upper layer. Figure 7 shows the $T$--$S$ diagram, which supports the dependence of density on temperature in the lower layer, as the salinity changed little under 20 m depth. Since the lower layer is substantially important for the cold dome, temperature structure rather than salinity is mainly focused on hereafter. In the all sections, the thermocline was around 10 m deep west of the section, while it bifurcated in the eastern part of the bay (Fig. 6). The upper half of the thermocline ($T > 23^\circ C$) was still lying horizontally in the eastern part, but the lower half ($T = 20--21^\circ C$) bent down and generated a prominent bottom front. Under the thermocline there was a cold ($T < 20^\circ C$) water mass, which extended from the middle to the western lower (deeper than 15 m) basin in each line.

Vertical profiles of temperature in the eastern and western part of the line C and at the Irago Strait are presented in Fig. 8. In the western area of the bay (CS), a sharp thermocline between 8 m and 13 m depth separated the upper and lower layers. The temperature dif-
ference reached 1.5°C m⁻¹. A weaker but prominent thermocline (ΔT ~ 0.7°C m⁻¹) was also found at the same depth in the eastern part (C13), although the western upper (lower) layer was much warmer (colder) than the eastern. In both areas, the upper and lower layers were relatively uniform. On the other hand, no explicit thermocline was observed at the Irago Strait, indicating the water column was mixed. The temperature at the Irago Strait, especially deeper than 13 m, was similar to that at C13, although the thermocline was indistinct compared with that at C13. This feature is also revealed in the T–S diagram (Fig. 7). Temperature and salinity of the water shallower than 20 m at the Irago Strait were between 21°C and 24°C and between 32.5 and 33.8, respectively. In the eastern part of each line (A12, B17 and C13), the water located between the upper and lower thermocline had the same character as the Irago Strait water, indicating that the shallow water at the Irago Strait intruded into the eastern middle layer of the bay. The deeper Irago water belonged to another group, which was more saline (S > 33.8) and dense (σt > 23.6) water. Notice that the bottom water from the middle to the western basin (C5) had an explicitly different character; the temperature was lower than 20°C, which was explicitly colder than the eastern middle layer (C13). In addition, it is indicated that this cold water mass extended from lines A to C at least, as the bottom waters in the all lines occupy the same position in the diagram (Fig. 7).

In Fig. 6, the temperature from 20°C to 24°C, which was the same as that at the Irago Strait, and inflowing areas faster than 10 cm s⁻¹ are shaded in the temperature and velocity panels, respectively. The area between 20°C and 24°C was thick in the eastern part. As the bifurcation of the thermocline, it corresponded to the inflowing area. In line C, the strongest middle layer inflow was ~17 cm s⁻¹ just above the tilting bottom front, while near the seabed the flow was ~4 cm s⁻¹ to the southeast. The difference Δu ~ 21 cm s⁻¹ is approximately consistent with the thermal wind relation, as

![Fig. 6. Transversal distributions of temperature, salinity, density and north (Line A) or northwest (Lines B and C) components of residual currents along the three lines. Positive and negative values of residual currents indicate inflow and outflow respectively, in the velocity panels. Reprinted from Cont. Shelf Res., 22, Kasai et al., Circulation and cold dome in a gulf-type ROFI, 1579–1590, © 2002, with permission from Elsevier.](image-url)
$\Delta \rho \approx 0.4 \text{ kg cm}^{-3}$ in a distance of 5 km with a depth of 20 m would lead $\Delta u \approx 22 \text{ cm s}^{-1}$. The inflowing velocities in the other two lines were also observed above the front and consistent with the relation. This indicates that the well-mixed strait water intruded along the eastern coast as the geostrophic flow. In contrast, the flow speed was weak in the western lower layer. A prominent bottom front separated the eastern brisk intrusion water from the western stagnant cold dome. This flow pattern, namely the inflow in the eastern side of the bay with the maximum in the middle layer, is considerably different from the conventional two-layer estuarine circulation. Since the low temperature ($<20^\circ\text{C}$) was observed nowhere else in the bay in this season (Figs. 6, 7) but existed in spring, the water of the cold dome would be a relict of cold spring water, which would be trapped beneath the thermocline.

### 3-3. Numerical model

From the detailed observations there are strong indications that the strait water intrudes along the eastern coast and relict water in the western lower layer forms the cold dome. The formation of the cold dome by this process can be demonstrated using a suitable model with two different density structures; a stratified and a mixed region (Kasai et al. 2002).

The model is applied to a rectangular bay of 38 km width and 64 km length, corresponding to Ise Bay (Fig. 9). To reproduce circulation and density structures in a gulf next to a well-mixed area in summer situation, the water is stratified with low vertical eddy viscosity and diffusivity ($=10^{-5} \text{ m}^{2} \text{ s}^{-1}$) in the bay ($x < 50 \text{ km}$), while the water is homogeneous with high viscosity and diffusivity ($=10^{-3} \text{ m}^{2} \text{ s}^{-1}$) in the mixed region ($x > 50 \text{ km}$), which corresponds to the Irago Strait. The water in the bay is stratified in the initial stage but no freshwater input is induced, because the movement of surface layer fresh water is dispensable. Typical temperature and salinity values in summer, with reference to those shown in Fig. 5, were chosen as the initial values; surface temperature and salinity inside the bay are 22°C and 30 and those at the bottom are 15°C and 33.5, respectively. Initial temperature and salinity in the mixed region are 18.0°C and 31.7, which are the same as those in the middle layer of the bay. Both tem-

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**Fig. 6.** (continued).
Temperature and salinity are uniform in the $y$-direction. Velocity is set to be zero in the whole region.

Figure 10 shows the model results indicating the horizontal distribution of temperature and velocity in the middle layer (23 m depth) after 25 days, when the field reaches nearly steady state. The mixed water flows into the stratified region in the center ($y = 14–24$ km) and is forced towards the right wall facing landward as it intrudes into the bay. This inflow is ~$15$ cm s$^{-1}$ and parallel to the isotherms (namely isohalines and isopycnals), suggesting that it is in the geostrophic balance. At the warmer side of the front ($x = 30–50$ km, $y > 20$ km) temperature is $-17.5^\circ$C, which is close to that in the mixed region ($T = 17.5–18^\circ$C). On the other side of the front, the water is cold with a minimum of $16^\circ$C. This contrast of temperature is consist-

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{fig10.png}
\caption{Simulated horizontal distributions of temperature and velocity at 23 m depth after 25 days. Reprinted from Cont. Shelf Res., 22, Kasai et al., Circulation and cold dome in a gulf-type ROFI, 1579–1590, © 2002, with permission from Elsevier.}
\end{figure}
ent with the observational results (Fig. 6). The velocity in this cold water mass is significantly weak ($u \sim 1$ cm s$^{-1}$) except for the vicinity of the wall, where there is return flow.

Figure 11 shows the cross-sectional views of temperature and velocity, facing landward. The water is stratified in the left half of the model basin. The dome-like water in the left deep layer is the coldest, while the surface water is warmest in the section. On the other hand, the homogeneous water, which has the same temperature ($T = 17–18^\circ C$) as that in the mixed region, exists in the right. The velocity panel clearly shows that inflow area concentrates in the right middle and upper regions ($y = 17–30$ km, $z > 30$ m), while the outflow is detected in the upper layer near the left wall. The velocity adjacent to the coldest region is appreciably weak. Both the temperature and velocity sections resemble observational results (Fig. 6) in combination of the stagnant cold dome and intruding mixed water. The model seems effective in producing both temperature distributions and current structures, and satisfactorily demonstrates the key physical principles detected by the observation.

3-4. Cold dome and hypoxia

The key point to control the mechanism is the coexistence of stratified and mixed areas. Figure 12 illustrates a schematic view of the density and flow structure in the bay. In the stratified season, the mixed strait water does not intrude along the bottom but enters through the middle layer of the bay. The Earth’s rotation forces the inflow to the right shore side looking landward in the northern hemisphere. Therefore, the water outside the intrusion is isolated because of the insufficient water exchange. The heating effect cannot reach the bottom water in the stratified area, so that spring water has been left cold.

The mechanism responsible for the formation of the seasonal dome is the summer-time production of horizontal density gradients between the stratified and mixed waters. According to the stratification-circulation diagram proposed by Hansen and Rattray

![Fig. 11. Cross-sectional view of simulated temperature and velocity at $x = 40$ km. Note that arrows directed upward in the velocity panel represent the component in x direction (inflow), and directed right represent the component in y direction. Reprinted from Cont. Shelf Res., 22, Kasai et al., Circulation and cold dome in a gulf-type ROFI, 1579–1590, © 2002, with permission from Elsevier.]

![Fig. 12. A schematic view of the intrusion process and the formation of the cold and hypoxic dome. Reprinted from Cont. Shelf Res., 22, Kasai et al., Circulation and cold dome in a gulf-type ROFI, 1579–1590, © 2002, with permission from Elsevier.]

![Fig. 13. Seasonal changes in density in the middle of Ise Bay and at the Irago Strait (20 m depth). “B-1 m” indicates “1 m above the bottom”. Reprinted from Cont. Shelf Res., 22, Kasai et al., Circulation and cold dome in a gulf-type ROFI, 1579–1590, © 2002, with permission from Elsevier.]

Fig. 13. Seasonal changes in density in the middle of Ise Bay and at the Irago Strait (20 m depth). “B-1 m” indicates “1 m above the bottom”. Reprinted from Cont. Shelf Res., 22, Kasai et al., Circulation and cold dome in a gulf-type ROFI, 1579–1590, © 2002, with permission from Elsevier.
(1966). Ise Bay is classified in the strongly stratified regime from April to October, while during the rest of the period it is in the weakly stratified regime. Figure 13 shows the seasonal change in water density at five depths in the middle of the bay and at 20 m depth in the Irago Strait. In autumn and winter (November–March) temperature is reversed and weak stratification is maintained by the only freshwater buoyancy input. Sea surface cooling sinks the freshwater through the mixing process, so that the water in the bay is lighter than that in the strait. During the heating season (April–October), on the other hand, both large freshwater discharge and sea surface heating makes the water stratified inside the bay. The effect of the heating from the surface extends to the bottom in the well-mixed area (Irago Strait), while it is restricted to the surface layer by the strong thermo- and haloclines in the stratified area (inside the bay). This procedure makes the density of the bottom bay water larger than that of the strait water. The strait water is, therefore, expected to go into the bottom layer in winter but into the middle layer in summer, when it intrudes into the bay (Fig. 3). In the transition season, the external factors such as freshwater discharge and wind stress change the strength of the stratification and could lead the alternation of the intrusion depth (Takahashi et al. 2000).

The cold dome is strongly related to the hypoxia as shown in Fig. 5. By inhibiting replacement of the water, DO is not supplied to the dome. Therefore, the oxygen consumption exceeds the supply, leading to hypoxia in the cold dome. Hypoxia is observed at several points in the Seto Inland Sea every summer (Fig. 1a). The tidal currents are weak and the water tends to be stratified in those areas (Fig. 1c). However, the water is well mixed in the straits, which are next to the stratified areas. This indicates the same physical mechanism controls DO concentration and causes hypoxia in the Seto Inland Sea.

4. Physical vs. biochemical processes controlling the formation of hypoxia

4-1. Introduction

As shown in the previous sections, physical processes are important for the formation of hypoxia. In this section, we present experimental results on the oxygen consumption over the bottom sediments and hydrographic observations both in hypoxic and non-hypoxic water. It will be clarified going through the process, that the contribution of physical processes outweighs that of biochemical processes in the formation of hypoxia in a coastal basin.

Hiuchi-nada is situated in the central part of the Seto Inland Sea, Japan (Fig. 14). It is well known that in the eastern part of the Hiuchi-nada the oxygen in the bottom layer is depleted every summer (Fig. 1a). The serious hypoxia has often damaged fisheries since 1960s in this area (Ochi and Takeoka 1986). The historical view on the formation mechanism of hypoxia is described as follows. Several pulp factories were constructed in 1960s–70s in Kawanoe and Iyomishima, which are on the southeastern coast of Hiuchi-nada (Fig. 14). The pulp factories loaded a large amount of organic and inorganic matter into the southern area of Hiuchi-nada. In the eastern part of Hiuchi-nada, there

Fig. 14. Bathymetry of the study area. Hydrographic observations were carried out at the solid circles. Velocities were measured along the solid line with an ADCP. Water samples for the Oxygen consumption rate experiments were taken from Hiuchi-nada (Stns. H7, H12 and H15), Harima-nada (Stn. H) and Bisan-seto (Stn. B). BS and SP denote Bisan-seto and the Shonai Peninsula, respectively. Reprinted from Coastal and Shelf Science, 71, Kasai et al., Flow structure and hypoxia in Hiuchi-nada, Seto Inland Sea, Japan, 210–217, © 2007, with permission from Elsevier.
is a cyclonic circulation (6th Regional Coast Guard Headquarters 1973), which transports the allochthonous and autochthonous organic matter to the northeast. En route this organic matter sinks and, therefore, accumulates at the bottom. Oxygen is thus consumed during the decomposition of this organic matter at the bottom (Ochi 1992).

For this reason, various measures have been taken to reduce the organic matter load from the factories since the early 1970s. These antipollution measures had an effect on water quality such that chemical oxygen demand (COD) in the region rapidly decreased in the late 1970s (Ukita 1998). However, the DO condition has unsatisfactorily remained.

Figure 15 shows year-to-year variation in the minimum DO concentration in the eastern part of Hiuchi-nada. Unlike the other chemical components, DO concentration in the bottom layer has stayed at a low level. It has increased gradually in recent years, but is still lower than 4 mg l−1. This situation indicates that another mechanism could affect the oxygen depletion in Hiuchi-nada.

Therefore, we measured oxygen consumption rates at the bottom in the eastern part of Hiuchi-nada. Comparing the results with those in the other regions, where hypoxia has never been observed, allows the importance of biochemical processes to be evaluated. To estimate the effect of oxygen supply by the physical processes on the hypoxia in the region, extensive CTD and ADCP observations were also conducted. Water exchange by horizontal advection in the bottom layer was evaluated based on the observational results. Combining the results of oxygen consumption experiments and field observations, it will be discussed whether biochemical or physical processes contribute more to the formation of the hypoxia in Hiuchi-nada.

4-2. Characteristics of Hiuchi-nada

The study area, Hiuchi-nada, is located in the center of the Seto Inland Sea, Japan (Fig. 14). As is shown in Fig. 1, the Seto Inland Sea is characterized by a combination of narrow strait and wide basin. In the western part of Hiuchi-nada the geography is very varied and includes many small islands, while in the eastern part the coastline is smooth (Fig. 14c). The complicated topography leads to stronger tidal currents over 1 m s−1 in the western part of the Hiuchi-nada and on the northern side of the Shonai Peninsula (Bisan-seto), but typical tidal currents are less than 0.1 m s−1 in the eastern part. The strong tidal currents result in relatively mixed water, whereas the weaker current makes the water in the east strongly stratified in summer. Hypoxia has often been observed in the eastern lower layer, but never in the west. No large rivers empty into Hiuchi-nada, so that temperature predominantly controls density. The fragmentary data set by observations is suggestive of cyclonic and anti-cyclonic circulation in the surface eastern and western part of Hiuchi-nada, respectively (6th Regional Coast Guard Headquarters 1973).

4-3. Oxygen consumption rate experiments

Sediment cores were taken from the surface to a depth of 20 cm using acrylic core tubes from three stations in Hiuchi-nada (Fig. 14c). After taking the cores to the laboratory, the water in the cores was replaced with the filtered oxygen-saturated water, taking care not to disturb the surface of the sediment. The top of each core was completely capped to exclude air bubbles, and they were incubated in a 20°C water bath in a dark room. The concentration of dissolved oxygen in the water was measured with an oxygen electrode inserted into each core tube. The water in the cores was gently stirred by the stirrer equipped with the electrode, and DO data were continuously monitored for more than 14 hours. As DO gradually decreased and converged to a certain level within 10 hours, DO consumption rate $R(t)$ was estimated by the decline of DO.
To compare the results from Hiuchi-nada with the other oxygen rich areas, the same experiments were conducted using the waters from Harima-nada and Bisan-seto, where hypoxia has never been observed (Fig. 14b). Measured DO concentration decreased monotonically in the three samples from Hiuchi-nada. The DO consumption rates in Hiuchi-nada were from 0.032 to 0.040 g m\(^{-2}\) h\(^{-1}\). Ochi and Takeoka (1986) and Hoshika et al. (1989) measured in situ DO consumption rates using a bell-jar type chamber in the eastern part of Hiuchi-nada. Their results were 0.02 and 0.03 g m\(^{-2}\) h\(^{-1}\) on average, respectively, which are slightly lower but still comparable to ours.

On the other hand, the Bisan-seto samples showed considerably high consumption rates (>0.05 g m\(^{-2}\) h\(^{-1}\)), although the in situ DO concentration is high. Unfinished feeds from adjoining fish farms accumulate at the bottom in Bisan-seto, where aquaculture is well developed. This large amount of organic matter would result in the highest DO consumption. The rates measured in Harima-nada were from 0.023–0.034 g m\(^{-2}\) h\(^{-1}\), which are comparable to those in Hiuchi-nada.

The DO consumption rates should be correlative to the in situ DO concentration, if the strength of DO consumption controls generation of hypoxia at the bottom. However, the relationship demonstrates that there is no significant correlation between them (Fig. 16, \(r^2 < 0.02\)). This indicates that the main reason for the generation of hypoxia in Hiuchi-nada is not the high DO consumption rate (not biochemical processes), but other processes.

4-4. Hydrographic survey

Field surveys consist of hydrographic observations over a wide range in 2002 and current observations along a line in 2003. The hydrographic observations were carried out in the eastern part of Hiuchi-nada on 21 June, 19 July, 16 August, and 19 September, 2002. Observational stations are shown in Fig. 14c. Vertical profiles of temperature and conductivity were obtained using a CTD profiler. DO was measured using an oxygen meter at 5 m depth intervals, and at every 1 m depth intervals around oxyclines (usually near the bottom). Currents were measured by a shuttling operation with

Fig. 17. Horizontal distributions of temperature and DO observed at the bottom in 2002. Reprinted from Coastal and Shelf Science, 71, Kasai et al., Flow structure and hypoxia in Hiuchi-nada, Seto Inland Sea, Japan, 210–217, © 2007, with permission from Elsevier.
an ADCP for about 12 hours on 3 August 2003 along the east-west line in the eastern part of Hiuchi-nada (Fig. 14c). The raw data was analyzed by a harmonic method to obtain tidal residual currents. Temperature and conductivity were measured once at each station using the CTD profiler. The same CTD observation was conducted at a point (H1) on the northern side of the Shonai Peninsula, as a representative of the mixed region.

The strong hypoxia was detected at the bottom in the eastern part of Hiuchi-nada in 2002. Figure 17 shows the observed horizontal distributions of temperature and DO concentration at the bottom. From June to August, cold (and dense) water was distributed along the depression. Strong temperature fronts (~0.2°C km⁻¹) lay along the western edge of the cold water. The shape of the cold water is similar in each observation, although the temperature increases 4–6°C from June to August. Sea surface cooling destroyed the stratification in September, and thus the cold water disappears and temperature became nearly uniform.

Distributions of DO concentration resemble those of temperature; the minimum DO area was observed from the center to east of the observational area and hypoxia corresponds reasonably well to cold water from June to August. Strong DO fronts (~0.5 mg l⁻¹ km⁻¹) were created along the temperature fronts. The DO concentration decreased until August, and then recovered in September by mixing, in the same way as the cold water mass.

Vertical sections along the east-west line (Fig. 18) show the water was strongly stratified in the center and eastern area (5–18 km from the western end of the observational line). Temperature and density resemble each other in structure, indicating that temperature predominantly controls density. Salinity difference in the observational section was moderate comparing with temperature. A strong thermocline was observed at around 10 m depth and the temperature difference between the surface and the bottom reached 9°C in June. However, the water at the most western point (H5) was relatively well mixed and the temperature difference was less than 3°C. The bottom front (T = 19–20°C) separated the mixed water from the eastern cold water with a temperature 2–3°C lower than the mixed water.

The distribution of DO deeper than 15 m depth was similar to that of temperature in the vertical section, as shown in the horizontal distributions. DO concentration was over 6 mg l⁻¹ in the upper layer, and lower than 5 mg l⁻¹ in the cold dome. On the other hand, the DO concentration was relatively high (>5.5 mg l⁻¹) in the western mixed area even at the bottom. Similar characteristics of temperature and DO were also observed in the north-south section (not shown here).

A scatter diagram of observed temperature and DO (Fig. 19) shows the oxygen was saturated and nearly uniform in the surface layer (between 0 m and 5 m). However, under the thermocline (deeper than 10 m), DO concentration decreased with temperature. There were significant correlations between temperature and DO, since the coefficient of determination (r²) is over 0.7 in each month except for September. The
colder water is considered to be older water (relict spring water) in the heating season (Hill 1993; Kasai et al. 2002). Therefore, the older water contains lower DO in Hiuchi-nada, indicating that the physical processes are important for the formation of hypoxia.

Temperature, DO and residual currents observed on 3 August 2003 are shown in Fig. 20. The pattern of salinity and density (not shown here) is similar to that of temperature. The thermocline was 4–7 m deep in the center, and deeper and gentler in the eastern and western part of the section. A dome-like cold (T < 22°C) and saline (S > 32.3) water existed under the thermocline around the center of the observational section. This dome corresponds to low-oxygen water mass (D.O. < 6 mg l⁻¹). These characteristics are the same as those observed in the hydrographic survey in 2002.

Residual currents from ADCP records show cyclonic circulation over the bottom cold dome, with maxima of 14 cm s⁻¹ southward and 10 cm s⁻¹ northward in the western and eastern part, respectively. This result is consistent with the flow pattern estimated by 6th Regional Coast Guard Headquarters (1973). However, the speed was considerably small within the cold dome with a maximum < 5 cm s⁻¹. The vertical profiles of DO and speed at the center of the observational line corresponded to each other (Fig. 21). In the surface layer, DO and speed were high, while in the bottom water, they were significantly lower. Both properties reduced rapidly in 12–19 m depths under the thermocline. Oxygen was depleted in the stagnant bottom water, suggesting that horizontal advection would play an important role for generation of hypoxia. It is considered that little oxygen-sufficient water would be supplied to the bottom layer.

The horizontal scale of the target phenomena is about 20 km. In contrast, the Rossby internal radius of deformation is estimated to be several kilometers from Fig. 18, which shows the thickness of upper and lower layers is ~ 10 m and density difference between the layers is ~1 kg m⁻³ in summer. Since the former is considerably larger than the latter, the Earth’s rotation is expected to be important for the dynamics of water circulation. Therefore the first-order dynamical balance ought to be geostrophic. The vertical shear in the along front current is related to the cross-front density gradient through the thermal wind relation:

$$\frac{dv}{dz} = -\frac{g}{\rho_f} \frac{\partial \rho}{\partial z}$$

where v is the along frontal velocity (northward velocity for Fig. 20), z is taken to be vertically upwards, g is gravitational acceleration, ρ is density, f is Coriolis parameter and z is the cross frontal co-ordinate (eastward for Fig. 20). Using this equation, geostrophic velocities were calculated from the observed density profiles relative to an assumed level of no motion at the deepest common depth between adjacent stations. The result (Fig. 20) demonstrates the existence of geostrophic circulation in the upper layer of the stratified eastern Hiuchi-nada. Associated with the thermocline slope, northward and southward geostrophic velocities of 10 cm s⁻¹ are detected on the eastern and western side of the dome over the bottom front, respectively. In the cold dome, however, the velocity is appreciably weaker (v ~ 1 cm s⁻¹), implying insufficient water exchange. Estimates of geostrophic currents (Fig. 20) are consistent with the observed flow pattern (Fig. 20c), indicating that the residual currents are in approximate geostrophic balance.

The water in Bisan-seto is well mixed even in summer, because of the strong tidal currents. In addition, freshwater supply from rivers reduces its density. In Fig. 20, the area where the density ranges from 20.3 to 21.5 kg m⁻³, which is same as that in Bisan-seto (H1), is shaded. This area is limited to the upper layer along the ADCP line, and is compatible with the cyclonic circulation. At the most western point of the east-west section in Fig. 20 (H5), the water shallower than 15 m had the same character as that at H1, indicating that comparatively mixed water at H1 flowed into the western side of the cold water. The mixed water contains sufficient DO, so that the DO concentration is relatively high at H5. On the other hand, the cold dome water was explicitly different in character; the temperature was lower than 19°C, which was colder than the water in Bisan-Seto. These features correspond to the flow pattern as shown in Figs. 20c, 20d, 21. The cold bottom water is stagnant and isolated from the surrounding water. The prominent bottom front and the
thermocline separate the western southward flow from the bottom cold dome.

Baroclinic circulation in coastal areas shares many aspects of its dynamics with tidal mixing fronts. The cold domes have been observed under the thermocline in a weak tidal area adjacent to the mixed area (LeFavre 1986; Svendsen et al. 1991; Hill et al. 1994, 1997; Kasai et al. 2002). The formation mechanism of the dome was clarified by Hill (1993) using a simple dynamical model. In a region of weak tidal currents the water stratifies during the heating season, even though the surrounding area is well mixed by strong tides. The water under the thermocline is warmed very slowly by the weak diffusion of heat. This relict winter water is, therefore, isolated from surrounding mixed waters by horizontal bottom fronts, and develops into a cold dome. The water over the dome circulates anticlockwise above the bottom front, based on the thermal wind relation as shown in the aforementioned estimates (Hill et al. 1994, 1997). Isolation of the bottom water from

![Fig. 20. Vertical distributions of (a) temperature, (b) DO, (c) northward components of residual currents and (d) vectors of residual currents observed on 3 August 2003. (e) Vertical distribution of estimated geostrophic current (northward component). The panels show views looking northward, with east being to the right. Triangles indicate observational points. Positive and negative values indicate northward and southward flow, respectively, in (c) and (e). Arrows directed upward represent northward currents, and arrows directed to the right represent eastward currents in (d). Shaded area indicates the area of the temperature from 23.5 to 26.5°C, which was same as that on the northern side of the Shonai Peninsula (H1). Reprinted from Coastal and Shelf Science, 71, Kasai et al., Flow structure and hypoxia in Hiuchi-nada, Seto Inland Sea, Japan, 210–217, © 2007, with permission from Elsevier.](image-url)
the surrounding water is essential: in addition to heat, oxygen is not transported into the cold dome from the surroundings, and thus the cold dome becomes hypoxic.

4-5. Conclusions

In order to clarify the formation mechanism of the hypoxic water mass in Hiuchi-nada, oxygen consumption rate experiments and detailed hydrographic observations were conducted. The oxygen consumption rate experiments showed the rates in Hiuchi-nada are less than or comparable to other oxygen rich areas, indicating that high oxygen consumption is not the main reason for the formation of hypoxia. On the other hand, the detailed hydrographic observations demonstrated that an isolated dome is created under the thermocline in the central and eastern part of the observational area. The water in the dome is significantly colder than the surrounding water and the current speed is weak, indicating it is the relict spring water and stagnant. Therefore, the main reason of the formation of the hypoxia in the eastern part of Hiuchi-nada is not the high oxygen consumption, but rather the insufficient water exchange in the bottom water.

5. Fortnightly shifts of intrusion depth of oceanic water into a bay and hypoxia

5-1. Introduction

As we saw in earlier sections, the strength of mixing mainly caused by tidal stirring has a key to control the bimodal character of the estuarine circulation in a bay. Hypoxia in coastal areas is greatly affected by the intrusion of mixed water (oxygen-rich water). This process controls the transport and dispersal of DO, so an understanding and elucidation of the scale and frequency of the intrusion is important, especially for eutrophic basins. In general, inflow events of oceanic water into a basin occur over a wide range of timescales, from days to years (Farmer and Freeland 1983). The variation of tidal stirring with spring-neap cycles is one of the most plausible phenomena that influence the mixing of strait water (Simpson 1997). In fjord systems, for example, many observations have shown that inflows to the deep water change in relation to tidal amplitudes (Griffin and LeBlond 1990; Allen and Simpson 1998). Some studies have demonstrated enhancement of the inflow by insufficient mixing over the sill during neap tides (Geyer and Cannon 1982). A similar phenomenon may occur in bays, because it is common to both bays and fjords that the water is stratified in the basin while it is mixed in the mouth.

From detailed hydrographic and ADCP observations, Kasai et al. (2000) showed that a strong inflow (~20 cm s⁻¹) toward the north existed just over the deepest depression in Ise Bay. In Section 3, on the other hand, we saw that the water that has a similar character to
the Irago Strait water flows into the bay through the middle layer with a maximum speed of 15 cm s\(^{-1}\). These studies suggest fortnightly shifts of intrusion depth into Ise Bay, because the former observation was conducted during the neap tide and the latter during the spring. It is therefore expected that the strength of tidal stirring (tidal strength) varies with spring-neap cycles at the narrow strait, and that the variation could lead to a switch from the middle layer intrusion to bottom intrusion of the strait water into the basin.

In this section, we present the observational results in Ise Bay, and try to clarify the time dependence of the mixing conditions at the Irago Strait and the time dependence of the depth of the intrusion of the strait water into the basin.

In the first spring tide (2 August), a cold (<21°C) and saline (>33) water pool was observed inside the bay. In the strait, temperature, salinity and DO concentration were observed twice a month from June 2000 to February 2001 (Fig. 22, closed triangles).

Examples of the vertical profile of density at the Irago Strait (Stn. K7) are shown in Fig. 23. A strong pycnocline was always observed in the surface layer (<10 m) in summer. During spring tides (thick lines), the water below the pycnocline was well mixed. Especially below 20 m, it was nearly homogeneous. On the other hand, water was weakly stratified even below the pycnocline during neap tides (thin lines). Density differences between 10 m and 50 m were smaller during the spring tides (0.54 and 0.48 kg m\(^{-3}\)) than during the neap tides (1.1 and 1.5 kg m\(^{-3}\)). It is clear that mixing at the Irago Strait was strongly affected by the tidal strength. This relationship is more readily apparent in the scatter plot shown in Fig. 24. There is a strong negative correlation between the tidal range on each observational day and the density difference between 10 m and 55 m depth at the strait, and the coefficient of determination (\(r^2\)) reaches 0.61 (\(P < 0.01\), 15 June–27 November 2000). The stronger mixing in the middle and lower layers of the strait is associated with the stronger tidal currents.

Three consecutive vertical distributions of temperature, salinity, density and DO concentration along the longitudinal section are shown in Fig. 25. The water in the basin was always strongly stratified by both surface heating and river discharge. Well-defined thermoclines and haloclines around 10 m depth separated the upper and lower layers at all observations. Below the thermocline, however, the water condition changed completely at every observation. In the first spring tide (2 August), a cold (<21°C) and saline (>33) water pool was observed inside the bay. In the strait,
on the other hand, the water was mixed under the thermocline and halocline, and temperature and salinity were higher than 21°C and less than 33, respectively. A prominent bottom front created at the bay mouth (45–55 km from the bay head) separated the strait water from the dense bottom water pool inside the bay. The density deeper than 15 m at the Irago Strait was between 22 and 22.3σt and equivalent to that of the middle layer, but explicitly less than the bottom layer inside the bay, indicating that the strait water did not intrude through the bottom, but through the middle layer. In contrast, during neap tide (31 August), the densities deeper than 15 m at the Irago Strait and inside the bay were similar to each other. Cold (<20°C) and saline (>33.5) water was distributed widely from the bay mouth to the middle of the bay with a depth of 20 m in the lower layer. The oceanic water apparently intruded through the bottom layer. It appears from the salinity distribution that the oceanic water reached 30 km from the bay head. The bottom cold water in the north (15–30 km) would be the water that was widely distributed in the basin on 2 August (20–40 km) and be forced to the bay head by the oceanic water, because their temperature (18–19°C), salinity (33.3–33.6) and DO (<2 mg l⁻¹) were similar. The weaker tidal current insufficiently mixed the strait water, as shown in Fig. 23, and thus the bottom water at the bay mouth became heavier than that inside the bay. It is considered that the density difference at the bay mouth leads to the downward shift of the intrusion depth of strait water. The hydrographic condition returned to its initial state during the second spring tide (22 August). Water at the strait was well mixed again. Saline (>34) oceanic water disappeared from the observational section and the bottom front with a density difference of more than 0.3 × 10⁻³ kg m⁻³ was recreated at the bay mouth (50–65 km). The densest water did not exist in the strait but at the bottom inside the bay, indicating

Fig. 25. Vertical distributions of temperature, salinity, density and dissolved oxygen along the longitudinal section on 2, 11 and 22 August 2000. Darker areas indicate cold (<21°C), saline (>33.5), dense (>23) and hypoxic water (<4 mg l⁻¹). With kind permission from Springer Science + Business Media: J. Oceanogr., Fortnightly shifts of intrusion depth of oceanic water into Ise Bay, 60, 2004, 817–824, Kasai et al., Fig. 5.
that strait water would not intrude through the bottom layer. Modulated by the tidal strength, the hydrographic condition changed in only 10 days.

This feature is also revealed in the temperature-salinity diagram (Fig. 26). Since the strait water was vertically homogeneous on 22 August (spring tide), the marks (triangles) under the thermocline gather in a small area in the figure, compared to the rather scattered marks on 11 August (neap tide) when the water was weakly stratified. On the other hand, the marks for the central basin (circles) spread in a long line on both dates. The water under the thermocline in the strait had a similar character to that in the depth range from 8 m depth to the bottom and from 12 m to 20 m depth in the central basin on 11 and 22 August, respectively. The bottom bay water on 22 August was explicitly more saline and colder than the strait water. The features of this T–S diagram tell us that the strait water intrudes into the basin in the middle and bottom layers during spring and neap tides, respectively.

5-3. Estimates of intrusion depth

The tidal range was used to evaluate the strength of mechanical stirring by tidal currents, because a larger tidal range naturally leads to stronger tidal currents. The tidal range is defined as the difference between the predicted higher high water and the lower low water in a day. Each constituent tide that is recorded in “Tidal constituents of the coast of Japan” (Japan Coast Guard 1983) is used for the prediction of tide. Estimates of the depth of inflow were made by comparing the density of the strait water to that at the centre of the bay. It is reasonable to assume that the incoming water inflows to the depth of equal density (Allen and Simpson 1998; Takahashi et al. 2000). From their analysis of the driving force of density current, Fujiwara and Yamada (2002) showed that the oceanic water flows into the Tokyo Bay through the layer where the inflow water has same density as the ambient water. They explained that the driving force has a maximum at a depth where there is no longitudinal density gradient between the mouth and the inside of the bay. As is shown in Section 3, temperature and salinity in the inflow area are equivalent to those in the Irago Strait under the thermocline. In addition, the intrusion process of the Irago water through the middle layer, which has same density as the Irago water, is successfully demonstrated by a numerical model in Section 3.
The intrusion depth was defined as follows (Fig. 27). First, the densities at three depths (beneath the pycnocline, at the bottom, and their average as the representative densities of the intrusion water) were selected in the Irago Strait (Stns. K6 or M5) on each observation. Secondly, the top, bottom, and the centre of the intrusion depths in the middle of the bay (Stn. K4) were defined as the depths that had the same densities as those beneath the pycnocline, at the bottom, and their average in the Irago Strait, respectively. The strait water was assumed to intrude through the bottom layer when the density below the pycnocline in the strait was larger than that at the bottom in the middle of the bay.

Figure 28 shows time changes in the intrusion depth and the tidal range at Nagoya. Open and closed circles indicate middle layer and bottom intrusion, respectively. Error bars in the upper panel indicate the top and bottom depth of the intrusion layer. With kind permission from Springer Science + Business Media: <J. Oceanogr., Fortnightly shifts of intrusion depth of oceanic water into Ise Bay, 60, 2004, 817–824, Kasai et al., Fig. 7>.

5-4. Distribution of DO affected by the intrusion depth

The lowest panels in Fig. 25 show the distributions of DO concentration. DO is always depleted (<2 mg l⁻¹) in the bottom water near the bay head (<25 km). During spring tides, the hypoxic water, which is defined as water with a DO concentration less than 4 mg l⁻¹, was widely distributed in the lower layer under the estimated intrusion depths. In addition, the distribution of hypoxia was closely related to the hydrographic conditions. The contour lines of 4 mg l⁻¹ of DO concentration corresponded well with the 21°C isotherm in both spring tides. DO concentration was relatively high in the well-mixed strait water and there was a strong DO front at the bay mouth, as is shown in the temperature front. This indicates that the physical processes strongly affect the distribution of hypoxic water. During the neap tide, on the other hand, the hypoxic water was forced to the bay head and/or uplifted to the middle layer by the bottom intrusion of the oxygen-rich oceanic water from the strait. The DO pattern was complicated and fragments of hypoxic water scattered around the front of oceanic water (25 km and 15–25 m depth). It appears that the oceanic water intruded landward, mixing with the bay water. The distribution of hypoxic water changed frequently due to the shifts in the intrusion of strait water. When waters from the strait intrude through the middle layer (spring tides) hypoxia develops in the lower layer below the intrusion depth, while the bottom intrusion reduces the scale of bottom hypoxia during neap tides.

5-5. Conclusions

From repeated observations, we have clarified the shift of intrusion depth of oceanic water into Ise Bay.
The key point that controls the mechanism is the different mixing condition between the stratified and mixed area. The conclusions of this section are schematically illustrated in Fig. 30. In the heating season, the water in the Irago Strait is well mixed during the spring tides and its density is equivalent to that in the middle layer inside the bay. The strait water therefore intrudes through the middle layer and the water under the intrusion is excluded from water exchange and becomes hypoxic (Fig. 30a). During the neap tides, the strait water is insufficiently mixed so that the density in the lower layer at the strait is greater than that inside the bay. The oceanic water therefore intrudes through the bottom layer and hypoxic water is pushed to the bay head and/or uplifted (Fig. 30b). The intrusion process of oxygen-rich strait water moves frequently according to the tidal strength. The distribution and scale of hypoxia changes according to the intrusion process of oxygen-rich strait water. The hypoxic water can be uplifted and the oxygen minimum can be observed in the middle layer due to the intrusion of oceanic water to the bottom according to the shift of the intrusion depth.

6. Nutrient release from hypoxic water

6-1. Introduction

In coastal environments, the dissolved oxygen concentration changes seasonally in response to stratification of the water column. Hypoxia or anoxia develops in the lower part of the water column in summer (e.g., Jensen et al. 1990; Kasai et al. 2002; Naqvi et al. 2006). The changing concentration of DO in the water just above the sediments alters the depth to which oxygen penetrates into the sediments, and limits the NO$_3^-$ supply from nitrification to denitrification within the sediment (Kemp et al. 1990; Rysgaard et al. 1994; Caffrey et al. 2003). Jensen et al. (1990) suggested that nitrification in the sediments may temporarily cease after rapid sedimentation events and that in such cases nutrient rich bottom waters may be the chief NO$_3^-$ source for sedimentary denitrification. Therefore, it is necessary to consider temporal and spatial variations in these pathways both within the sediments and at the sediment-water interfaces with changing concentrations of oxygen.

NO$_3^-$ concentrations in aquatic systems are affected by both physical and biogeochemical processes. High primary production increases the availability of organic matter in marine environments. Increased organic matter loading enhances O$_2$ consumption, sedimentary NH$_4^+$ regeneration and consequent NH$_4^+$ fluxes from the sediment into the water column (Jensen et al. 1990). This NH$_4^+$ may be oxidized to NO$_3^-$ by nitrifying bacteria within the water column. Direct measurements of nitrification rates by $^{15}$N tracer techniques have con-
sistently shown that nitrification occurs in the lower portion of the euphotic zone in coastal environments (Ward 2005). Even if the nitrification rates in a unit volume of the water column are lower than those in the sediments, the total volumetric rates would be significant when integrated over the water column (Pauer and Auer 2000). In shallow coastal environments, however, the quantitative contribution of each process to the total NO₃⁻ generation is insufficiently understood.

Here we report NO₃⁻ isotope data from the water column in Ise Bay, which is one of the most eutrophic estuaries in Japan. In spring and summer the water is strongly stratified as a result of heating and large freshwater run-off (Kasai et al. 2004). Bottom water becomes hypoxic and regenerated dissolved inorganic nitrogen (DIN) accumulates in the lower layer of the hypoxic water. In addition, there could be a seasonal transition in the dominant DIN composition from NH₄⁺ in spring to NO₃⁻ in summer, but the quantitative contribution of water-column nitrification to the NO₃⁻ maximum in summer is little understood. We have thus investigated the seasonal change in DIN composition in the central part of the bay. The main objective of this study is to ascertain the occurrence of nitrification in the water column and to elucidate the effects of nitrification on the δ¹⁵NNO₃.

6-2. Nutrient concentrations in hypoxic water

Oxygen concentrations in the bottom water at the deepest point of the bay (Stn. K4 in Fig. 22) change seasonally; a hypoxic condition develops with thermal stratification during summer (Fig. 31). The variation in the dissolved oxygen concentrations is associated with seasonal changes in the concentrations of DIN (=NO₃⁻ + NO₂⁻ + NH₄⁺) in the bottom water (Fig. 31). Although DIN concentrations increased with decreasing DO concentrations until June, the loss of DIN was prominent in summer. The decreasing oxygen concentrations may have induced the temporal variation in DIN composition. There were lags of few months between the NH₄⁺ and NO₃⁻ peaks, the former occurring in spring, while the latter were found in summer (Fig. 31).

Vertical profiles of NH₄⁺ concentrations in the lower layer showed different trends from those of NO₃⁻ concentrations (Fig. 32). In May, high concentrations of NH₄⁺, probably originating from mineralization, were found in the oxic (DO > 100 µM) lower water column. NO₃⁻ was vertically homogeneous at low concentrations (2–3 µM) and had δ¹⁵NNO₃ values of −6.3 to −10.4‰ with an average of −8.5 ± 2.0‰ below 10 m depth. On the other hand, NO₃⁻ concentrations increased significantly in the lower water column with
decreasing NH$_4^+$ and DO concentrations by July. NH$_4^+$ fall to nearly zero at all depths. Moreover, $\delta^{15}$NNO$_3$ increased to 8.4 ± 0.7‰ in the hypoxic water (DO < 100 µM) deeper than 20 m. Although concentrations of NO$_3^-$ and DO decreased simultaneously with depth in the deepest layer between 30 and 35 m, $\delta^{15}$NNO$_3$ was almost constant with depth (Fig. 32b). A discrepancy between stoichiometric variation in NO$_3^-$ with DO and isotopic variation in NO$_3^-$ was also observed in the deepest layer at the other stations. NO$_3^-$ concentrations therefore showed a strong negative correlation with DO concentrations in the bottom waters in July ($r^2 = 0.90$). NO$_3^-$ in the water column was significantly 15N-depleted in May (Fig. 32a). Nitrification has been proposed as the main cause of low $\delta^{15}$NNO$_3$ in several marine environments (Ostrom et al. 1997; Sutka et al. 2004; Reinhardt et al. 2006). NO$_3^-$ newly generated by nitrification is significantly depleted in $^{15}$N with respect to the substrate NH$_4^+$, because partial nitrification occurs, with a marked isotope effect. The extremely $^{15}$N-depleted NO$_3^-$ found in Ise Bay in May is ascribed to newly produced NO$_3^-$ arising from nitrification. By contrast, $\delta^{15}$NNO$_3$ had increased significantly within the lower water column by July. The inverse relationship between the concentrations of NO$_3^-$ and NH$_4^+$ at the two observation times, in conjunction with the decrease in DO concentrations over this period, indicates that NH$_4^+$ was completely oxidized and turned into NO$_3^-$ by July.

6-3. Contribution of nutrients in hypoxic water to primary production

Nutrients are key elements in phytoplankton production and subsequent ecosystem processes in the sea. Various nutrient sources, including riverine, oceanic, and regenerated nitrogen, maintain the phytoplankton production in coastal ecosystems. Of these major nitrogen sources, riverine input affected by anthropogenically perturbed nitrogen flux from land areas to the coast has long been considered primarily responsible for eutrophication in coastal zones. On the other hand, we showed in the previous Subsection 6-2 that rich organic matter in the sediments and the water column in eutrophic coastal ecosystems supply a large amount of regenerated nutrients, which accumulate in the oxygen-depleted water mass. Moreover, recent studies have revealed that a large amount of nitrogen is supplied from the adjacent marginal sea to the coastal region. Intrusion of nutrient-rich shelf water would stimulate phytoplankton production in coastal ecosystems (e.g., Yoder et al. 1981; Sugimoto et al. 2009).

The quantitative characterization of DIN sources to phytoplankton production is essential for the management of coastal areas. Since these complicated processes are strongly non-linearly coupled, studies on the relationships between nutrient behavior and phytoplankton production have usually relied on physical-ecosystem models that include inorganic and...
Japan: Use of nitrogen isotopes to identify dissolved inorganic nitrogen sources, 450–466, Elsevier.

The ecosystem model has five compartments, namely ocean and regeneration) to phytoplankton production.

The ecosystem model has five compartments, namely

\( \text{NH}_4^+ \), \( \text{NO}_3^- \), phytoplankton (PHY), zooplankton (ZOO) and detritus (DET), as the prognostic variables being concentrations of nitrogen. The evolutions of these compartments are described with differential equations composed of biological source and sink terms, external loading terms, diffusion terms, and advection terms. Two other parameters, dissolved inorganic phosphorus (\( \text{PO}_4^{3-} \)) and DO, are calculated using stoichiometric methods in the model. The oxygen consumption by remineralization in the sediments is proportional to the oxygen consumption rates based on the observational results. Comparison between physical parameters such as velocities, temperature and salinity calculated by this model and observed parameters confirmed that the model can reproduce physical processes in coastal areas (Kakehi 2006). Details of the model are described by Sugimoto et al. (2010).

Figure 33 shows the seasonal variations in observed and calculated concentrations of Chl-\( \alpha \), \( \text{NH}_4^+ \) and \( \text{NO}_3^- \) concentrations in the middle and bottom layer at the central part of Ise Bay. Calculated concentrations of each compartment show seasonal patterns similar to observed ones. The calculated concentrations of Chl-\( \alpha \) in the middle layer are high in spring (~4.5 \( \mu g L^{-1} \)) and autumn (~3.5 \( \mu g L^{-1} \)), while low in winter (~1 \( \mu g L^{-1} \)). Chl-\( \alpha \) maximum in spring amply reproduces the subsurface chlorophyll maximum in the bay. The concentrations of \( \text{NH}_4^+ \) and \( \text{NO}_3^- \) and their seasonal variations in the middle layer are smaller and lower than those in the bottom layer. The \( \text{NO}_3^- \) concentration in the middle layer decreases to ~1 \( \mu M \) during the spring bloom. In the bottom layer, the \( \text{NH}_4^+ \) concentration peaks in spring (~7 \( \mu M \)), while \( \text{NO}_3^- \) concentration peaks in summer (~12 \( \mu M \)). This temporal shift from \( \text{NH}_4^+ \) to \( \text{NO}_3^- \) in the lower layer clearly displays the effect of nitrification, indicating the accumulation of regenerated \( \text{NO}_3^- \) in the lower layer during the stratified periods. The model can overall reproduce the observed conditions.

Figure 34a shows the seasonal variation in DIN (=\( \text{NH}_4^+ + \text{NO}_3^- \)) fluxes of the new supply from the river and ocean, and the consumption rates of DIN by phytoplankton within the bay. The riverine flux was calculated as a product of the DIN concentration and riverine circulation. High assimilation continues during summer (~200 T N d\(^{-1} \), peaking in autumn (~220 T N d\(^{-1} \), while low in winter (~20 T N d\(^{-1} \). However, the DIN supply from the river and shelf shows decreasing trends in autumn. In winter, consumption rates rapidly decrease to < 80 T N d\(^{-1} \). Figure 34b presents the ratio of DIN consumption by phytoplankton to the supply from the river and shelf (consumption/supply ratio). The ratio is higher than 1.0 throughout the year, indicating that the external...
supply of DIN is insufficient and that the phytoplankton production largely depends on the regenerated DIN within the bay. The consumption/supply ratio considerably differs in spring and autumn. The lower ratio in spring (<1.5) suggests a larger contribution of the external DIN supply from the river and ocean, while the higher ratio in autumn (~3) suggests the dominant contribution of regenerated DIN. In Ise Bay, physical properties change seasonally as was shown in the previous sections. This seasonal difference strongly influences the relationship between phytoplankton and DIN sources.

As an example, model results in summer are shown in Fig. 35. Water stratification is strengthened by surface heating and high freshwater discharge in summer. In the bottom layer, the cold water mass (<18°C) is isolated from the surrounding waters by the horizontal thermo- and haloclines and the vertical bottom front. This cold water mass becomes hypoxic (DO < 3 mg L⁻¹). The Chl-a concentrations show two prominent maxima; a surface maximum at the bay head and a subsurface maximum at the bay mouth. As both NH₄⁺ and NO₃⁻ are depleted in the upper layer, the δ¹⁵NPHY of the two Chl-a maxima expresses the δ¹⁵N values of assimilated nitrogen, because there is little isotope fractionation. Near the river mouth, the slightly high δ¹⁵NPHY (~7 ‰) and low f-ratio (<0.6) indicate uptakes of riverine NH₄⁺, which has low δ¹⁵N values (<20 ‰) compared to those of regenerated NH₄⁺ (>40 ‰) in the lower layer. Moving from the bay head to the center of the bay, phytoplankton tends to use NO₃⁻ instead of NH₄⁺. On the other hand, at the bay mouth, the phytoplankton forming the subsurface Chl-a maximum takes up NO₃⁻ as the major nitrogen source. However, the NO₃⁻ source (oceanic and/or regenerated) cannot be determined because both types of NO₃⁻ have similar isotope values in summer. The consumption/supply ratio (~2, Fig. 34b) suggests that the contribution...
of regenerated NO$_3^-$ is comparable to that of external DIN supplied from the rivers and shelf. On the other hand, the high (>0.6) f-ratio indicates the weak contribution of NH$_4^+$ to phytoplankton production in the upper layer. The $\delta^{15}N$ values in the upper layer is considerably lower than that in the lower layer. This indicates that residual NH$_4^+$ is left by the nitrification processes in the lower layer, while the new NH$_4^+$ is produced by remineralization in the upper layer. Mino et al. (2002) found that $\delta^{15}N$ values of suspended matter decrease with the uptake of regenerated NH$_4^+$ rather than the uptake of new NO$_3^-$ supplied from the deep layer in the Atlantic Ocean. The relatively low $\delta^{15}N_{PHY}$ values (<10‰) associated with lower f-ratios (<0.6) indicate that phytoplankton do not take up the residual NH$_4^+$ by nitrification. They rather take up the newly generated NH$_4^+$ by remineralization in the upper layer. Fluxes of newly regenerated NH$_4^+$ by remineralization (33.5 tN d$^{-1}$) and excretion (19.8 tN d$^{-1}$) in the upper layer (>12 m) was found to be considerably larger than fluxes of NH$_4^+$ supplied from the lower layer (<12 m) to the upper layer by advection (3.9 tN d$^{-1}$) and diffusion (12.3 tN d$^{-1}$).

Figure 36 shows the annual average nitrogen flux (tN d$^{-1}$) of each process and of standing stocks (tN) in Ise Bay. The nitrogen supply of 32 tN d$^{-1}$ from the river is composed of 24 tN d$^{-1}$ as NO$_3^-$, 5 tN d$^{-1}$ of NH$_4^+$, and 3 tN d$^{-1}$ as detritus. The net nitrogen flux from the ocean reaches 29 tN d$^{-1}$, which is comparable to the riverine nitrogen flux. Dissolved forms, NO$_3^-$ (37 tN d$^{-1}$) and NH$_4^+$ (11 tN d$^{-1}$), are supplied from the ocean to the bay, while particulate forms composed of phytoplankton (12 tN d$^{-1}$), zooplankton (2 tN d$^{-1}$), and detritus (7 tN d$^{-1}$) flow out from the bay to the ocean. A total of 77 tN d$^{-1}$ of DIN is supplied to the inner bay as new nitrogen.

In contrast, the nitrogen flux for each process within the bay is considerably larger than the new nitrogen flux from the outer regions. Phytoplankton takes up 93 tN d$^{-1}$ of NO$_3^-$ and 67 tN d$^{-1}$ of NH$_4^+$. The total flux of phytoplankton assimilation (160 tN d$^{-1}$) is more than twice the total flux from the river and ocean. This result means that more than half the amount of annual phytoplankton production is supported by the regenerated DIN. However, high consumption/supply ratios (>2.0) occur only in periods from September to November (Fig. 34b), suggesting that regenerated DIN, which is accumulated in the hypoxic water mass during the stratified season, contributes significantly to annual phytoplankton production. Because a large amount of phytoplanktonic nitrogen is transferred to detritus (94 tN d$^{-1}$) rather than zooplankton (54 tN d$^{-1}$), phytoplankton mortality accounts for more than 70% of the total detritus supply, which is one of the largest processes in Ise Bay. Sedimentation of detritus is larger than the NH$_4^+$ release from the sediments. Regenerated nitrogen from remineralization (54 tN d$^{-1}$) and release from sediments (55 tN d$^{-1}$) also shows large flux. These two fluxes account for ~70% of the total NH$_4^+$ supply, suggesting that they are the most important sources for phytoplankton production. Moreover, zooplankton excretion accounts for 16% of the total NH$_4^+$ supply. A large supply of NO$_3^-$ is nitrification (80 tN d$^{-1}$), which accounts for ~60% of the total NO$_3^-$ supply. The denitrification loss of 48 tN d$^{-1}$ is twice as large as the NO$_3^-$ supply from the river (24 tN d$^{-1}$).

Fig. 36. Annual averages of nitrogen fluxes (tN d$^{-1}$) and standing stocks (tN) in Ise Bay. The magnitude of flux is visually shown by the thickness of lines. Reprinted from Coastal and Shelf Science, 86, Sugimoto et al., Modeling phytoplankton production in Ise Bay, Japan: Use of nitrogen isotopes to identify dissolved inorganic nitrogen sources, 450-466, © 2010, with permission from Elsevier.
The calculated denitrification rate is comparable to that estimated from the Redfield stoichiometry (42 tN d⁻¹; Kakehi et al., 2005) and larger than the potential denitrification rate (19–28 tN d⁻¹; Sugawara, 2003). These results indicate that phytoplankton production in Ise Bay is mainly dominated by the internal cycle (DIN assimilation by phytoplankton, mortality of phytoplankton, remineralization of detritus, release from the sediments, nitrification, and denitrification) rather than the external supply (fluxes from rivers and the ocean). The former is accompanied by oxygen consumption, and thus strongly related to the hypoxia in the bay.

In Ise Bay, the formation of a cold water mass isolated from surrounding waters during the stratified periods is key to the large contributions of regenerated nitrogen to phytoplankton production, because the cold water mass covers a large volume in the lower layer (e.g., Kasai et al., 2002). The oxygen within the isolated water mass is mainly consumed by remineralization of organic matter in the waters and sediments, and thus a large amount of DIN accumulates in the hypoxic water mass during the stratified periods (Sugimoto et al., 2008). Consequently, the standing stock of DIN in the lower layer is high (~2,500 tN) in summer, but low in winter (~1,400 tN). The consumption/supply ratio differs considerably per season (Fig. 34b). The lower ratio in spring (<1.5) means that the direct DIN supply from the rivers and ocean is more important than that from regeneration, while the higher ratio in autumn (~3) means that regenerated DIN is the major source for phytoplankton production (Fig. 34b).

Although the riverine DIN stimulates phytoplankton production and forms a Chl-α maximum in the surface layer at the bay head, phytoplankton production at the bay mouth is largely controlled by the estuarine circulation. In spring, the intrusion depth of oceanic water changes from the bottom to the middle layer (e.g., Kasai et al., 2007). The oxygen within the isolated water mass is mainly consumed by remineralization of organic matter in the waters and sediments, and thus a large amount of DIN accumulates in the hypoxic water mass during the stratified periods (Sugimoto et al., 2008). Consequently, the standing stock of DIN in the lower layer is high (~2,500 tN) in summer, but low in winter (~1,400 tN). The consumption/supply ratio differs considerably per season (Fig. 34b). The lower ratio in spring (<1.5) means that the direct DIN supply from the rivers and ocean is more important than that from regeneration, while the higher ratio in autumn (~3) means that regenerated DIN is the major source for phytoplankton production (Fig. 34b).

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6.4. Conclusions

An important aspect of the nitrogen cycle in coastal environments concerns the source of the nitrogen used in primary production. Phytoplankton production in Ise Bay, one of the most eutrophic embayments in Japan, is supported by external nitrogen derived from rivers and the ocean, and regenerated nitrogen formed in hypoxic water within the bay. We evaluated the contribution of each source of dissolved inorganic nitrogen (DIN) to phytoplankton production in Ise Bay. The ecosystem model revealed that DIN consumption by phytoplankton exceeds the DIN supply from the rivers and ocean, indicating that a large amount of phytoplankton production in Ise Bay depends on regenerated DIN within the bay rather than on newly supplied DIN from hypoxic water. The intrusion depth of oceanic nitrate changes from the bottom to the middle layer in spring. Oceanic nitrate is transported into the euphotic layer by the middle-layer intrusion and stimulates phytoplankton production at the bay mouth. The subsurface chlorophyll maximum layer then develops. In autumn, however, the intrusion depth of oceanic water changes from the middle layer to the bottom layer. Regenerated NO₃⁻, which is accumulated in the hypoxic water mass, is lifted up and supplied to the euphotic layer. These results imply that phytoplankton production in Ise Bay is mainly dominated by the internal cycle rather than the external supply.

7. General conclusions

In this article, we reviewed recent studies on the formation mechanism of hypoxic water in the bottom layer in summer. Although both biological and physical processes control the generation of hypoxia, the physical processes such as water exchange and convection determine the scale and strength of hypoxia, rather than the biological processes namely oxygen consumption (Section 4). The key point for understanding the formation mechanism is the existence of a region of strong vertical mixing which maintains well mixed condition next to a stratified area with weak currents (Section 2). The bottom water in the stratified area is not renewed and thus tends to be isolated and hypoxic. Therefore hypoxia occurs in specific places, in which spring water has been left as a cold dome (Section 3). The scale of the hypoxia changes according to the relation between the strengths of stratification and mixing determined by tidal condition and water densities (Section 5). The hypoxia has an enormous effect on the marine ecosystem through not only the direct negative impact on the living organisms, but also the nutrient release and subsequent primary production in coastal embayment (Section 6).
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