

Hypsographic Control on the Diapycnal Flow in the Eastern Gotland Basin of the Baltic Sea

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ABSTRACT: The implications of bottom intensified mixing together with an insulating boundary condition at the ocean floor for the ocean overturning has been investigated recently by many authors in many different settings. Here, we apply similar ideas to a brackish coastal sea, the eastern Gotland Basin in the Baltic Sea. The deep overturning in the basin is estimated using high-resolution temperature, salinity, and bathymetry data for a number of assumed vertical buoyancy flux and diffusivity profiles. The overturning structure in the basin is, in many ways, found to be similar to that inferred for the global ocean in recent studies, having rising water in a bottom boundary layer and sinking waters in the stratified mixing layer above. Moreover, the vertical structure of the overturning is found to be controlled by the hypsographic curve, with strong transformation occurring in certain depth ranges. These transformation hotspots are, however, not visible when water mass transformation is averaged in buoyancy space indicating that the effect of these hotspots is spread naturally over a large range of water masses.

KEYWORDS: Buoyancy; Upwelling/downwelling; Vertical motion; Thermohaline circulation; Water masses/storage

1. Introduction

A new paradigm describing the ocean overturning has emerged in which the deep overturning in the global ocean is a small residual of fast upwelling in narrow boundary layers along the sloping seafloor and slower downwelling occurring in the much larger stratified interior above the boundary layers (Ferrari et al. 2016; McDougall and Ferrari 2017). Two assumptions, both supported by observations, are central to the theory. First, turbulent fluxes owing to small-scale mixing decrease in magnitude with distance from the ocean floor, and second, there is an insulating boundary condition at the ocean floor (Polzin et al. 1997; St. Laurent et al. 2012). The first of these assumptions leads to the conclusion that small-scale mixing drives densification and consequently sinking rather than rising of deep-water masses. The second assumption imposes the condition that close to the bottom there must be an area of flux convergence, where water masses are made less dense. As an alternative to an insulating bottom boundary condition, one may also use a Neumann condition allowing for geothermal heating, which gives rise to an even stronger flux convergence (McDougall and Ferrari 2017).

Diagnostic overturning calculations have been done in many different settings, from idealized ocean geometries (e.g., cones, bowls, and inverted cones) to more realistic ocean basin geometries and have helped explain many features of the global ocean overturning (de Lavergne et al. 2017; Holmes et al. 2018; Drake et al. 2020). Most of these applications make use of the water mass transformation framework first introduced by Walin (1977). The framework combines tracer and volume budgets in tracer space and makes it possible to calculate the volume flux

across an isotracer surface from estimates of the nonadvective fluxes of said tracer. The framework was first applied using salinity as tracer but has since been extended to temperature, density, and multitracer applications (Walin 1982; Marshall et al. 1999; Iudicone et al. 2008; Hieronymus et al. 2014).

The first usage of the water mass transformation framework by Walin (1977) was an application to the Baltic Sea, which incidentally is also the area of interest for our application. The Baltic Sea is a semienclosed brackish and strongly salt stratified water body. The sea consists of a number of subbasins separated by sills. Here, we focus on the eastern Gotland Basin (EGB), see Fig. 1 (upper panel). To a first approximation, the stratification in the basin can be thought of as a two-layer system with relatively freshwater above sill depth separated from much more saline water below by a strong halocline (Lehmann et al. 2022). The main ventilation of the deep water comes with intermittent inflows of high saline waters from Kattegat (Mohrholz 2018), a bay in the North Sea that constitutes the Baltic Sea's only connection to the open ocean. A plot of salinity against time for the lower layer therefore looks somewhat like a sawtooth wave, with a very irregular period, see Fig. 1 (lower panel). During the short inflows, the salinity quickly goes up, and then the salinity is slowly lost to the upper layer during the much longer stagnation periods. The poor ventilation during the often multiyear long stagnation periods in combination with eutrophication leads to many environmental problems (Kuliński et al. 2022). Most prominently, wide spread oxygen deficiency in the bottom waters (Almroth-Rosell et al. 2021). The aim of our endeavors here is to derive a model for salinity changes during stagnation periods and to see how those are affected by different assumptions that can be made about the vertical nonadvective buoyancy flux profile. For this purpose, we lean heavily on earlier work, especially McDougall and Ferrari (2017).

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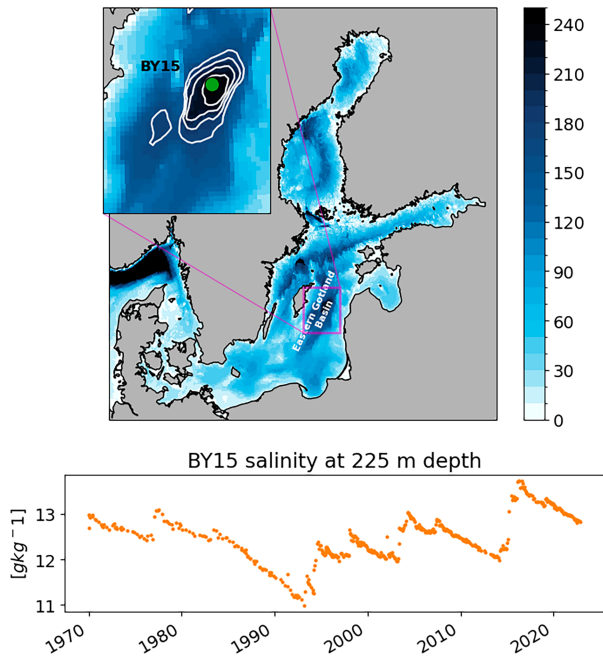


FIG. 1. (top) Bathymetric map of the area and (bottom) deep salinity as a function of time from the station BY15. Isolines are drawn at 170-, 190-, 210-, 230-, and 250-m depth. The station BY15 is also the station where the temperature and salinity data used to infer the basin stratification in our analyses are taken from.

The Baltic Sea has many features in common with the global ocean, like an overturning circulation, a permanent stratification, meso- and submesoscale eddies, and bottom intensified small-scale mixing (Reissmann et al. 2009). However, there are also some notable differences such as very small tidal amplitudes and no bottom water production. A major difference between the Baltic Sea and the global ocean is that the availability of observations is much higher in the former. Here, we utilize the availability of long-term observations of salinity and temperature as well as high-resolution bathymetry in our aim to better understand the deep circulation of the EGB.

A main motivation for our investigation is that earlier tracer-release experiments in the EGB have demonstrated the leading order importance of near-boundary mixing, for the spreading of tracers during stagnation periods (Holtermann et al. 2012; Holtermann and Umlauf 2012). Moreover, these tracer-release experiments did not only give evidence of rapid rising velocities close to the boundary but also of sinking in the stratified interior (Holtermann et al. 2012; Holtermann and Umlauf 2012). It should be noted that the observed sinking could have other causes, some of which were discussed by Holtermann et al. (2012). Nevertheless, these observations give evidence of a deep circulation in the EGB that appears to share interesting characteristics with that proposed to govern the deep overturning in the global ocean in the emergent paradigm discussed in the first few paragraphs. Moreover, evidence of a similarly structured overturning circulation in the Baltic Sea was very recently found in a 2-yr-long integration of a regional ocean model (Henell et al. 2023). These observations, as well as those

of bottom intensified mixing in the Baltic Sea (Reissmann et al. 2009), constitute a strong motivation for making more detailed calculations.

2. Theory and method

a. Transformation of basin deep water

Figure 2 shows a schematic of our domain of interest, a deep-water mass bounded by the ocean floor and an isopycnal surface, as well as the advective and nonadvective fluxes affecting this water mass. The volume of deep water V is defined by having a buoyancy that is less than b . This volume consists of water masses both in the stratified mixing layer (SML) and in the well-mixed bottom boundary layer (BBL). The boundary between these two regions is sketched with a dashed line. The volume is bounded by the seafloor and the isopycnal surface where the buoyancy is equal to b . Through the isopycnal surface goes fluxes of volume $[G\,(\text{m}^3\,\text{s}^{-1})]$ and nonadvective fluxes of buoyancy $[F\,(\text{m}^4\,\text{s}^{-3})]$. Near the bottom, there is a well-mixed bottom boundary layer, shown by the dashed line. Through the top of the boundary layer, there are fluxes of volume $[M\,(\text{m}^3\,\text{s}^{-1})]$ and nonadvective fluxes of buoyancy $[F_{\text{BBL}}\,(\text{m}^4\,\text{s}^{-3})]$. Our aim is to derive expressions for G separated into a component representing the transformation in the BBL (G_{BBL}) and one representing that in the SML (G_{SML}). The idea of calculating these transformations separately comes from McDougall and Ferrari (2017), but our derivations and designations are more in line with those used originally by Walin (1977, 1982).

Consider the volume and buoyancy budgets of the whole volume (i.e., the BBL + SML regions). Volume conservation gives

$$\frac{\partial V}{\partial t} = -G, \quad (1)$$

and conservation of buoyancy gives

$$\frac{\partial}{\partial t} \int_{-\infty}^b b' \frac{\partial V}{\partial b} db' = -Gb + F, \quad (2)$$

where we note that G contains both the contribution to G from the BBL and the SML, that is G_{BBL} and G_{SML} . Substituting $b\partial/\partial b$ of Eq. (1) into $\partial/\partial b$ of Eq. (2) yields

$$G = \frac{\partial F}{\partial b}. \quad (3)$$

Let us now consider those same budgets for the volume within the BBL, denoted as V_{BBL} . Volume conservation gives

$$\frac{\partial V_{\text{BBL}}}{\partial t} = M_{\text{BBL}} - G_{\text{BBL}}, \quad (4)$$

and conservation of buoyancy gives

$$\frac{\partial}{\partial t} \int_{-\infty}^b b' \frac{\partial V_{\text{BBL}}}{\partial b} db' = \int_{-\infty}^b b' \frac{\partial M_{\text{BBL}}}{\partial b} db' - G_{\text{BBL}}b + F_{\text{BBL}}, \quad (5)$$

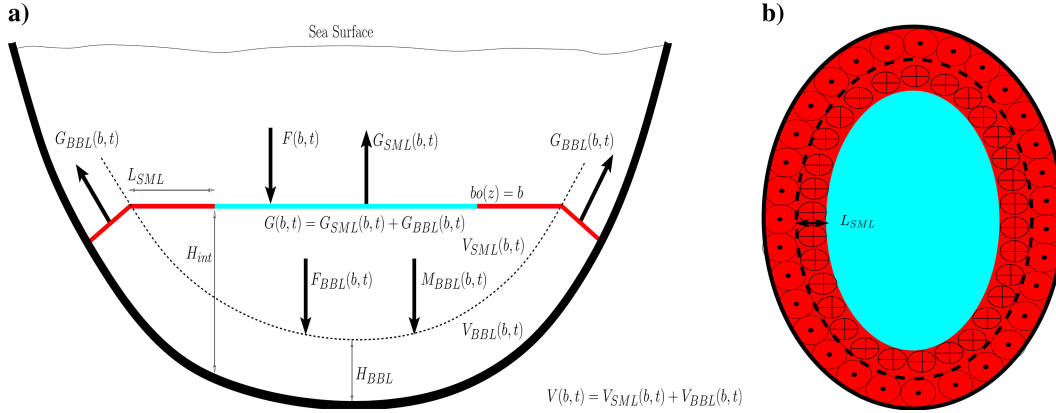


FIG. 2. Sketch of our domain with fluxes of buoyancy and volume. Seen (a) from the side and (b) from the top of the isopycnal. The V is the volume where the buoyancy is less than b , which is bounded by the isopycnal surface where the buoyancy equals b (sketched in red and magenta) and the ocean floor. The red parts of the isopycnal surface marks the area of the isopycnal surface where the mixing is intensified, while the magenta part marks the area of the isopycnal that is far enough away from the bottom that water mass transformations are very weak. The G_{SML} ($\text{m}^3 \text{s}^{-1}$) is the volume flux across the isopycnal surface that occurs outside the BBL. The G_{BBL} is the part of G that occurs in the BBL. The dashed line marks the separation of the BBL and the SML and is also where the nonadvective buoyancy flux f ($\text{m}^2 \text{s}^{-3}$) reaches its maxima. The different F ($\text{m}^4 \text{s}^{-3}$) terms denote laterally integrated nonadvective buoyancy fluxes, and M_{BBL} ($\text{m}^3 \text{s}^{-1}$) denotes the volume flux through the top of the BBL. The top view shows the areas where water is rising with arrowheads and sinking with fletchings.

Substituting $b\partial/\partial b$ of Eq. (4) into $\partial/\partial b$ of Eq. (5) yields

$$G_{BBL} = \frac{\partial F_{BBL}}{\partial b}. \quad (6)$$

Moreover, $G = G_{BBL} + G_{SML}$ giving

$$G_{SML} = \frac{\partial F}{\partial b} - \frac{\partial F_{BBL}}{\partial b}. \quad (7)$$

Note that our derivations do not rely on any steady-state assumptions, which is important as the deeper parts of the Baltic Sea are never in a steady state. Instead, they are characterized by having irregular oscillatory behavior, owing to inflow-stagnation period cycles.

In our computations, we approximate both F and F_{BBL} as vertical fluxes occurring across small horizontal area elements. That is, we approximate both the isopycnal surface bounding V and the material surface marking the top of the BBL (the area that separates the SML from the BBL) the same way that the free surface and the ocean floor are represented in z -coordinate ocean models. With this approximation, the area of the isopycnal surface bounding $V(b, t)$ and the area that separates the BBL from the SML are equal in size, as they are given by the same sum of area elements. Here, we call the area of these surfaces $A(b, t)$. Using this designation, $F(b, t)$ can be written as

$$F(b, t) = \int_A f(b, t) dA = A \langle f \rangle, \quad (8)$$

where f is the buoyancy flux per unit area and $\langle \rangle$ signifies an average over the area $A(b, t)$. That is, we use uppercase letters (e.g., F) to denote area-integrated terms and lowercase letters (e.g., f) to denote per unit area terms. The F_{BBL} is given by

$$F_{BBL}(b, t) = \max_b(f) \int_A dA = \max_b(f) A, \quad (9)$$

because f takes its maximum value on the boundary between the BBL and the SML. Note that this means that F_{BBL} is directly proportional to A and is thus a function only of the hypsographic curve and $\max_b(f)$, which is here taken to be uniform in x and y . The f profiles are discussed separately in section 2b.

Substituting Eqs. (8) and (9) into Eq. (7), G_{SML} can be written as

$$G_{SML} = A \frac{\partial \langle f \rangle}{\partial b} + [\langle f \rangle - \max_b(f)] \frac{\partial A}{\partial b} = G - G_{BBL}. \quad (10)$$

The second term of Eq. (10) is always negative as $\langle f \rangle$ cannot exceed $\max_b(f)$, and the first term is negative as long as the averaged buoyancy flux is bottom intensified. Given that G_{BBL} is always positive, well as long as, $\partial A/\partial b > 0$ or for a stably stratified ocean when $\partial A/\partial z > 0$, it stands to reason that the overturning G , in many ocean basins and lakes, is a sum of two opposing contributions, a positive contribution from G_{BBL} and a negative contribution from G_{SML} . In fact, McDougall and Ferrari (2017) found G in the global ocean to be much smaller in magnitude than the two opposing contributions from G_{BBL} and G_{SML} . As a quantitative measure, they introduced the amplification factor:

$$\mathcal{A} = \frac{G_{BBL}}{G}. \quad (11)$$

A definition that works fine as long as $G > 0$. Here, however, we have situations where $G < 0$ (i.e., situations where $\partial F/\partial b < 0$). We believe such situations can occur after inflows if the new

bottom water is more strongly stratified than the water masses immediately above. Leading to situations where F decreases with b at the boundary between the more stratified new bottom water and the weakly stratified old deep water. For this reason, we define the amplification factor as

$$\mathcal{A} = \max\left(\frac{G_{\text{BBL}}}{G}, \frac{G_{\text{SML}}}{G}\right). \quad (12)$$

The amplification factor thus goes to infinity when $|G_{\text{BBL}}/G_{\text{SML}}|$ goes to -1 . The amplification factor can be related to the ratio of the diapycnal velocity in the BBL to that in the SML. To proceed, we set $G_{\text{BBL}} = A_{\text{BBL}}\langle w_{\text{BBL}} \rangle$ and $G_{\text{SML}} = A_{\text{SML}}\langle w_{\text{SML}} \rangle$, where w_{BBL} and w_{SML} are the average diapycnal velocities in the BBL and SML, respectively, and $A_{\text{BBL}} = H_{\text{BBL}} \times C$ and $A_{\text{SML}} = L_{\text{SML}} \times C$ are the areas of the BBL and SML, respectively. Here, C is the circumference of the basin, H_{BBL} is the height of the well-mixed bottom boundary layer, and L_{SML} is the length of the isopycnal surface where intensified mixing occurs, measured from where the surface intersects the BBL toward its center. That is, L_{SML} scales as $H_{\text{int}}/\tan(\alpha)$ (McDougall and Ferrari 2017), where H_{int} is the height above the bottom where the mixing is intensified and α is the bottom slope. It is worth noting that L_{SML} typically is much smaller than the radius of the basin and thus that deep isopycnal surfaces typically have three dynamical regimes: upwelling in the BBL, downwelling in the SML, and a quiescent interior that is far enough away from the bottom to have negligible water mass transformation. Putting these estimates together, we get

$$\frac{\langle w_{\text{BBL}} \rangle}{\langle w_{\text{SML}} \rangle} \approx \frac{\mathcal{A}}{1 - \mathcal{A} \tan(\alpha) H_{\text{BBL}} / H_{\text{int}}}, \quad (13)$$

where $\mathcal{A} = G_{\text{BBL}}/G$ has been used (i.e., we are restricted to $\mathcal{A} > 1$). From Eq. (13) it is clear that the velocity ratio asymptotes to $H_{\text{int}}/[H_{\text{BBL}} \times \tan(\alpha)]$ for very large $\mathcal{A}$. As $\mathcal{A}$ tends to one, the first fraction on the left-hand side tends to infinity and the second tends to zero, so the velocity ratio for very small $\mathcal{A}$ depends on the speed by which the area of and velocity across the SML vanishes when G_{SML} tends to zero. Under typical oceanic conditions, small bottom slopes and $\mathcal{A} \approx 3$, $\langle w_{\text{BBL}} \rangle$ is much larger than $\langle w_{\text{SML}} \rangle$.

b. Flux profiles

The rate of water mass transformation depends on the divergence of the nonadvective buoyancy flux in buoyancy coordinates as was shown in the preceding subsection. To compute the water mass transformation, we shall here assume that we know the buoyancy flux f , either directly or through knowledge of the turbulent diffusivity κ and the buoyancy frequency N^2 . Measuring the buoyancy flux directly is not practically possible but many types of measurements help constrain it. From microstructure measurements, we can infer the viscous dissipation ϵ , which can be related to f through the Osborn (1980) relation $f = \Gamma \epsilon$, where a constant mixing efficiency, $\Gamma = 0.2$, is typically used. From tracer-release experiments (Holtermann et al. 2012; Holtermann and Umlauf 2012), we know a bit about κ , which relates to f through

$f = \kappa \partial b / \partial z = \kappa N^2$. The N^2 and its variability are known quite well from measurements that have been ongoing with an approximately monthly frequency in the EGB since the 1960s. Estimates of κ derived from observationally based budgets of salinity and heat is another possible constraint on κ (Axell 1998).

Just as in the proper ocean, several lines of evidence suggest that mixing is bottom intensified in the EGB. Moreover, the buoyancy flux f must go to zero at the ocean floor to satisfy the no-normal flux condition on the boundary. Here, we will assume a number of different f profiles that all have a similar functional form, heavily inspired by Ferrari et al. (2016) and McDougall and Ferrari (2017). A difference from these aforementioned works is, however, that we consider both κ and f profiles, while the earlier works considered only f profiles. The functional form we use for f is

$$f = 0.2\{B_1 + B_0 e^{[-z - (H - H_{\text{BBL}})]/h}\}. \quad (14)$$

If $z \geq -(H - H_{\text{BBL}})$ and otherwise

$$f = 0.2(B_1 + B_0) \times (H + z)/H_{\text{BBL}}, \quad (15)$$

where z is the vertical coordinate equal to zero at the ocean surface, H is the maximal depth as a function of x and y , H_{BBL} is the height of the well-mixed bottom boundary layer, h is the e -folding length scale, and B_0 and B_1 set the magnitude of the buoyancy flux curve. Similarly for κ , we use

$$\kappa = \{K_1 + K_0 e^{[-z - (H - H_{\text{BBL}})]/h}\}. \quad (16)$$

If $z \geq -(H - H_{\text{BBL}})$ and otherwise

$$\kappa = (K_1 + K_0) \times (H + z)/H_{\text{BBL}}. \quad (17)$$

The aim of our investigation is not to try and pin down the most basin-appropriate f profile but rather to investigate what difference different profiles make for the water mass transformation in the basin. Therefore, we have done calculations using a number of different parameter values. Our best guess profile, based on Reissmann et al. (2009), Holtermann et al. (2012), Holtermann and Umlauf (2012) and references therein, is called “base,” and the other profiles represent slight variations around the parameter choices used for the base case. However, the simple analytical nature of our functions for κ and f suggests that also the base case should be considered a very coarse representation of reality. The main aim of these functions is to capture three features: 1) that both ϵ and κ have been found to increase by about an order of magnitude in the vicinity of the bottom from 10^{-9} to 10^{-8} and 10^{-6} to 10^{-5} , respectively; 2) no buoyancy flux goes through the ocean floor; and 3) a well-mixed bottom boundary layer of varying thickness (typically some meters) exists over the ocean floor. The first and third features are supported by observations (Reissmann et al. 2009; Holtermann et al. 2012; Holtermann and Umlauf 2012), while the second feature is motivated by the theoretical consideration that the flux curves must satisfy the insulating boundary condition at the bottom. The parameter values used for base and our other profiles are given in Table 1. Vertical

TABLE 1. Parameter values for κ and f in the different cases.

	B_0	B_1	K_0	K_1	H_{BBL}	h
Base	2×10^{-8}	2×10^{-9}	1×10^{-5}	1×10^{-6}	5	5
int/10	2×10^{-9}	2×10^{-9}	1×10^{-6}	1×10^{-6}	5	5
intx10	2×10^{-7}	2×10^{-9}	1×10^{-4}	1×10^{-6}	5	5
h1	2×10^{-8}	2×10^{-9}	1×10^{-5}	1×10^{-6}	5	1
h10	2×10^{-8}	2×10^{-9}	1×10^{-5}	1×10^{-6}	5	10
h20	2×10^{-8}	2×10^{-9}	1×10^{-5}	1×10^{-6}	5	20
BBL2	2×10^{-8}	2×10^{-9}	1×10^{-5}	1×10^{-6}	2	5
BBL10	2×10^{-8}	2×10^{-9}	1×10^{-5}	1×10^{-6}	10	5
BBL20	2×10^{-8}	2×10^{-9}	1×10^{-5}	1×10^{-6}	20	5

profiles of f and κ for the different cases over a 250-m-deep part of the basin are shown in Fig. 3. Each of these cases thus give rise to two different ways of constructing a buoyancy flux profile for the basin either by using a vertical f profile or by using the expression $f = \kappa N^2$ together with an observationally based N^2 and a vertical κ profile. Both approaches are tested in the following.

c. Numerics

Our calculations are done on the horizontal grid of the two nautical mile version of the NEMO-Nordic model (Hordoir et al. 2019). Bathymetry is taken from the General Bathymetric Chart of the Oceans 2020 (GEBCO-2020). The vertical

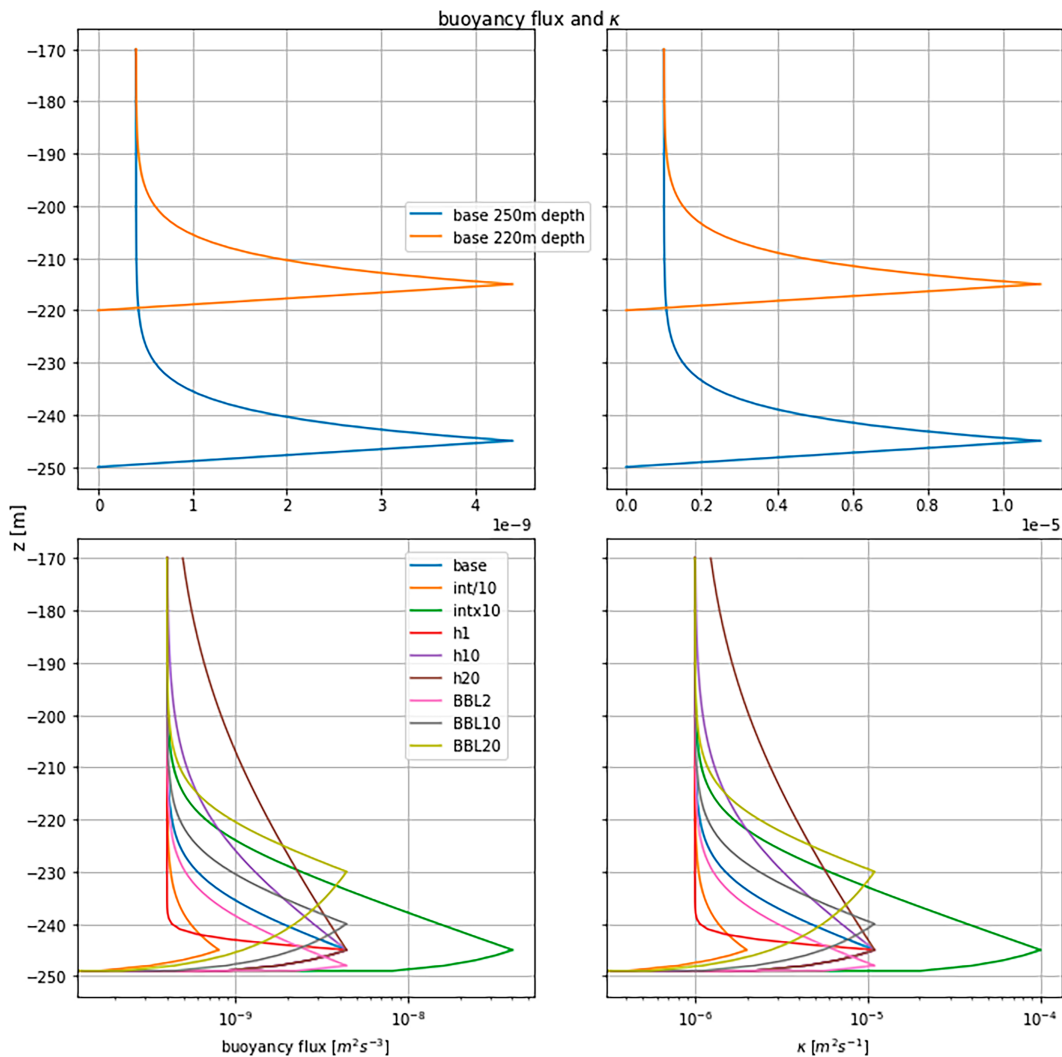


FIG. 3. Vertical profiles of κ and f for the different cases. (top) The profiles for the base case parameters for two different depths. (bottom) All different cases for a depth of 250 m, with a logarithmic x axis.

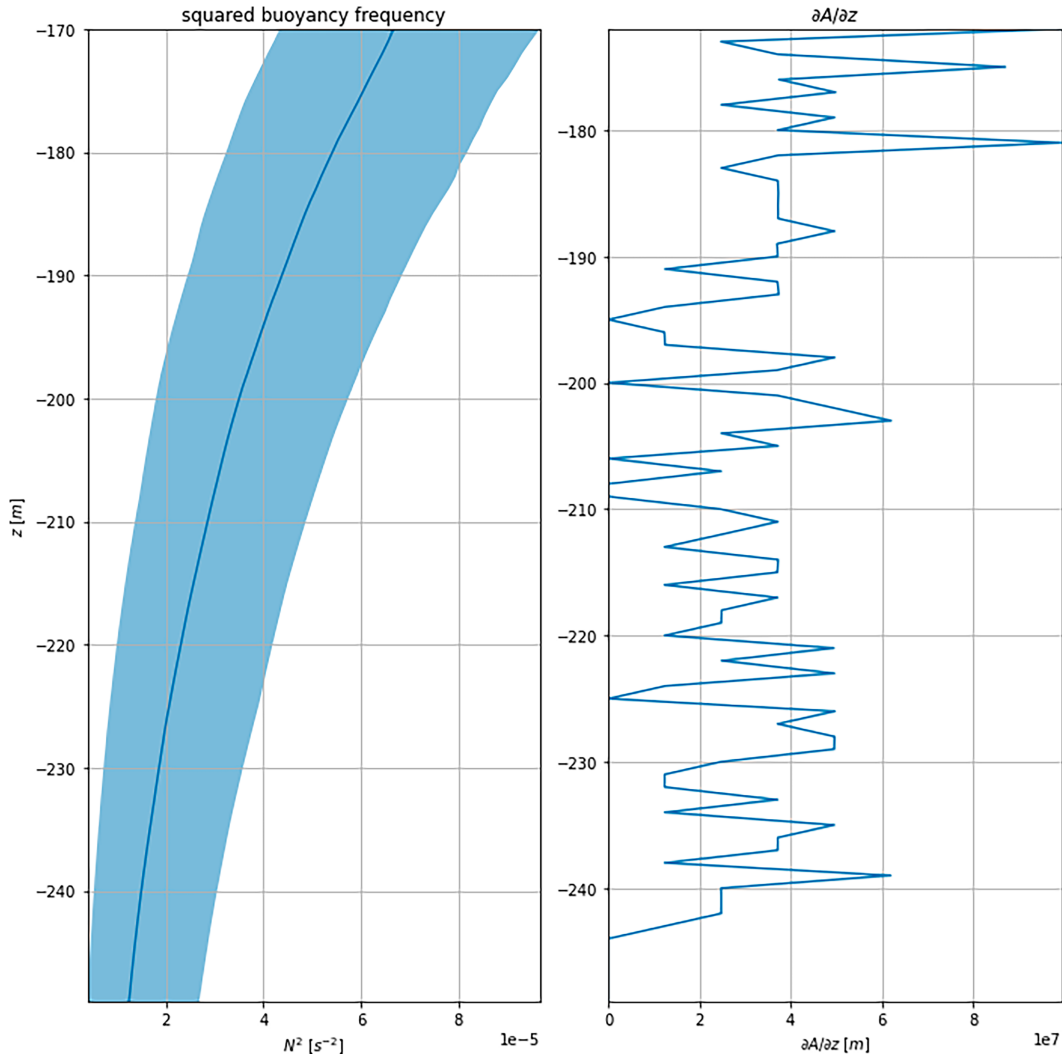


FIG. 4. Vertical profiles of N^2 and $\partial A/\partial z$. The shaded area shows the 5th–95th percentile range for N^2 derived from the 322 potential density profiles, while the solid line shows the median value.

grid used has a 1-m resolution. The N^2 is calculated from measured T and S profiles from the station BY15 using the Thermodynamic Equation of Seawater 2010 (TEOS-10). The T and S profiles are available since the early 60s with an approximate monthly resolution. In total, we have 322 such profiles. From those profiles, we calculate potential density referenced to 170-m depth using TEOS-10. Each of the 322 potential density profiles is then used to fit a potential density profile of the form $\sigma = a_1 + a_2 e^{a_3 z}$. From these, N^2 can then be computed analytically. A large number of measured profiles give us an opportunity to evaluate not only a mean state but also the range of variability in water mass transformation that occurs as a consequence of stratification changes. Figure 4 shows N^2 and a measure of its variability together with the vertical derivative of A . The latter is included as it has a strong control on the transformation as can be seen from Eq. (10).

It should be noted that using potential density as a coordinate is a mere approximation. We are treating potential density here

as if was a materially conserved quantity like T and S . That is, we ignore effects of the nonlinearities of the equation of state that we know to be important in the buoyancy budget of the global ocean (Klocker and McDougall 2010; Hieronymus and Nycander 2013b) and that should contribute to the buoyancy budget also in the EGB. Ignoring these effects is common also in the global ocean, but here in the Baltic Sea it is much more justifiable as the stratification at depth is strongly S dominated. This means that water mass transformations owing to the cabbeling and thermobaricity nonlinearities (Hieronymus and Nycander 2013b; Groeskamp et al. 2016) are small, and that similar results should be expected from b - and S -based frameworks.

To calculate $\partial F/\partial b$ from our depth-based profiles, we use

$$G = \frac{\partial F}{\partial b} = \frac{\partial F}{\partial z} \frac{\partial z}{\partial b} = \frac{1}{N^2} \frac{\partial F}{\partial z}, \quad (18)$$

where we used the inverse function theorem, and that $\partial b/\partial z = N^2$. The $\partial F/\partial z$ is estimated using a five-point stencil giving an error

of order dz^4 , while N^2 is calculated analytically from the fitted density profiles. The G_{SML} and G_{BBL} are evaluated the same way using five-point stencils for the respective F derivatives and an analytically derived N^2 .

These assumptions form the basis for our computations. Calculations start at the bottom of the EGB, which is approximately at 250-m depth, and we focus our interest in the depth range between 170- and 250-m depth. Horizontal integration is done using a criterion that all grid points at a given depth must be connected. That is, other Baltic Sea basins that are deeper than z' but that are connected to the EGB at depths shallower than z' are not contributing to $F(z')$. We also assume that surfaces of constant b coincide with surfaces of constant z except for where the surface of constant b intersects the BBL. In the BBL, we instead assume surfaces of constant b to be normal to the seafloor, see Fig. 2.

3. Overturning implied by different buoyancy flux profiles

Here, we investigate the overturning implied by the different κ and f curves for the different stratifications. Figure 5 shows G , G_{BBL} , and G_{SML} as a function of depth for the different f profiles given in Table 1. Shown are the 5th, 50th, and 95th percentiles of the different G values. The spread owes to them being calculated for our 322 different stratifications. The usage of a depth rather than a more customary buoyancy coordinate is done because our interest here is not in how different water masses change over time but in how the hypsographic curve, $A(z)$, affects the overturning in the basin. The different G curves in the figure thus illustrate the diapycnal transport across the buoyancy surface that happen to be at the depth seen on the y axis for the given stratification.

A main finding in this paper is that the vertical structure of G as a function of z changes very little with the stratification and is qualitatively similar for all parameter choices for f given in Table 1. Note that this applies only to the structure, not to the magnitude of G which changes similarly to f in the int/10 and intx10 experiments. In other words, the structure of the overturning in the basin is very strongly constrained by the hypsographic curve, which can be seen in the consistent vertical structure of G for all the different f curves, and through the similarity between $\partial A/\partial z$ shown in Fig. 4 and the different G curves shown here. The variation with stratification is weaker for G than for G_{SML} and G_{BBL} . The reason is simple to see from Eq. (18). Since $\partial F/\partial z$ is independent of b for a vertical f profile, the spread comes from the division with N^2 . Moreover, G is smaller than both $|G_{\text{SML}}|$ and $|G_{\text{BBL}}|$, and the opposing sign on G_{SML} and G_{BBL} means that a stratification that gives a large positive G_{BBL} also gives a large negative G_{SML} and vice versa. That is, the net diapycnal flow at a fixed z level is rather unaffected by stratification changes because of compensatory changes in the diapycnal flow in the SML and the BBL.

Figure 6 is the same as Fig. 5, but instead of vertical f profiles, we use a vertical κ profiles and hence a buoyancy flux profile given by $f = \kappa N^2$. Here, we find similarly to the earlier example that G is strongly constrained by the hypsographic curve and qualitatively similar for all vertical κ profiles. A

noteworthy difference to the earlier shown f -based G profiles is that these are less-bottom intensified. This is simply a result of N^2 decreasing with depth, as the κ profiles used here have the same depth dependencies as the f profiles used in Fig. 5. Also noteworthy here is that the overall stratification-induced variation in all G terms is very small. Again from Eq. (18), we have

$$G = \frac{1}{N^2} \frac{\partial F}{\partial z} = \frac{1}{N^2} \frac{\partial A \langle \kappa \rangle N^2}{\partial z} = \langle \kappa \rangle \frac{\partial A}{\partial z} + A \left[\frac{\partial \langle \kappa \rangle}{\partial z} + 2 \langle \kappa \rangle \frac{\partial \ln(N)}{\partial z} \right], \quad (19)$$

showing that the dependence on the stratification for the G terms in this case goes as the $\ln(N)$ term. For our stratifications, we get nearly no spread in any of the G terms for the κ -based flux profiles, thus suggesting that basin geometry controls the structure of the diapycnal flow. This is also evidenced by the similarity between $\partial A/\partial z$ shown in Fig. 4 and different G curves. It is worth noting that many parameterizations of κ are N dependent. Nilsson et al. (2003), for example, stated that $\kappa \propto N^{-1}$ and $\kappa \propto N^{-1/2}$ are candidate choices. Our two cases have $f \propto N^0$ and $f \propto N^2$, respectively. Using $\kappa \propto N^{-1}$ and $\kappa \propto N^{-1/2}$ would thus give a stratification-induced G variability that is between that in our two cases. We could thus assume that changes in the stratification should give modest changes in G at fixed depths also for commonly used stratification-dependent κ parameterizations.

Turning our attention now to the diffusive fluxes F rather than the diapycnal transports G . These are shown in Figs. 7 and 8 for the vertical f and κ profiles, respectively. Here, the BBL curve shows F_{BBL} as defined in Fig. 2. That is, it shows the nonadvective buoyancy flux entering the BBL through the control surface at the top of the BBL and thus not the negligible buoyancy flux F that goes through the tiny part of the surface of constant buoyancy that is in the BBL [i.e., to the left (right) of where the dashed and $b = \text{constant}$ lines intersect on the left (right) in Fig. 2]. The total term shows F as defined in Fig. 2, and the SML curve is simply the difference between the total and the BBL. The three curves thus show the net nonadvective buoyancy flux entering the BBL, the SML + BBL, and the SML, respectively.

The situation with respect to the spread is for F more or less the opposite of what it was for G . The result is as expected; vertical f profiles naturally give rise to F profiles that are independent of N^2 , and vertical κ profiles give rise to N^2 -dependent F profiles. For a fixed f profile and a variable N^2 , we thus get constant F profiles with some variability in G . For a fixed κ profile and a variable N^2 , we instead get a nearly fixed G profiles and variable F profiles. Neither of these flux profiles are entirely realistic, and some degree of stratification dependence and time variability on both F and G should be expected in reality. Moreover, realistic flux profiles should depend not only on distance from the bottom but also on bottom roughness and the circulation. That is, they should be functions of latitude, longitude, and time as well as of the distance from the bottom. Nevertheless, all tested, somewhat observation like and not so observation like, buoyancy flux profiles give a net diapycnal transport that is strongly constrained by the hypsographic

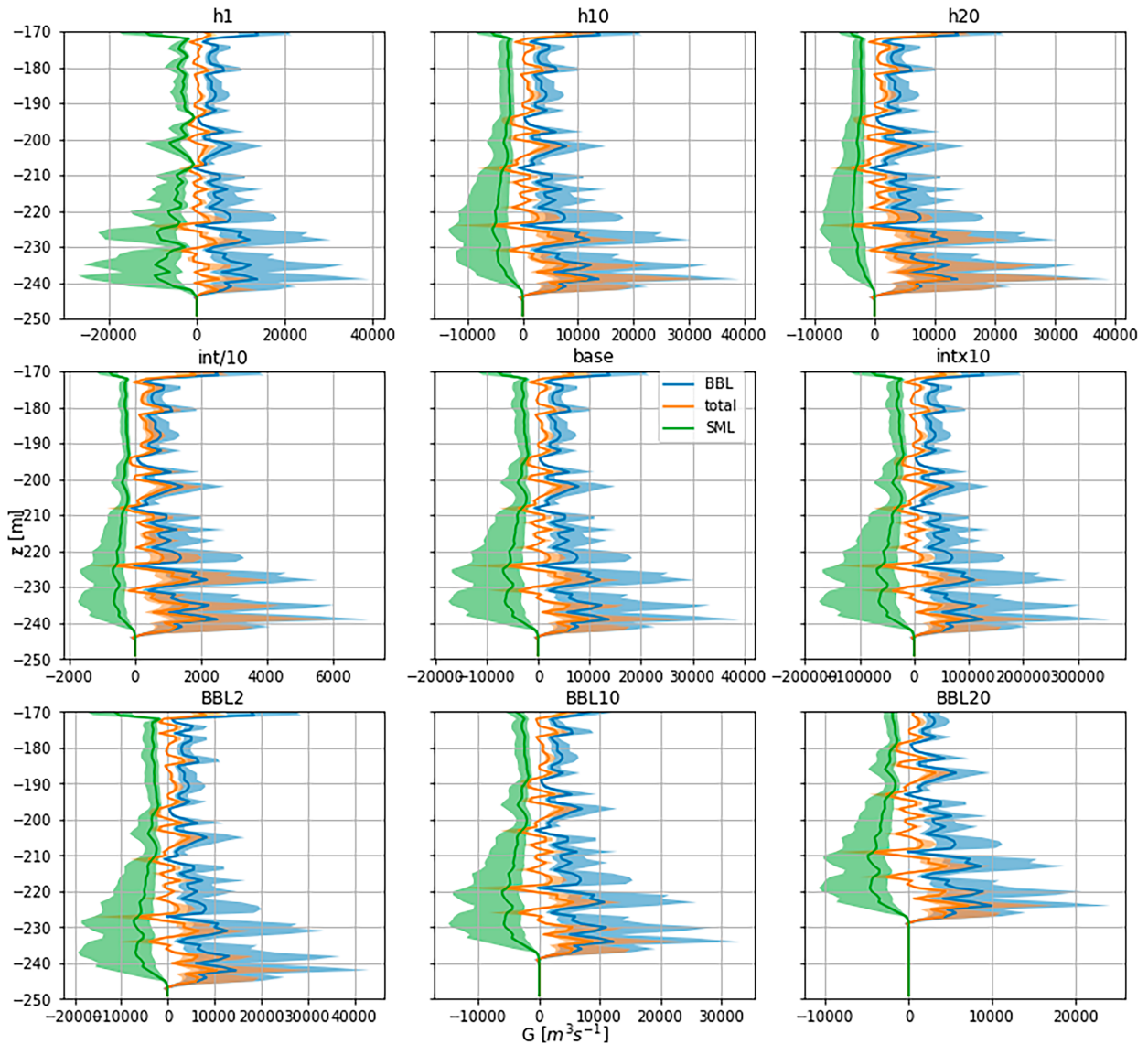


FIG. 5. Vertical profiles of G , G_{BBL} , and G_{SML} for the different vertical f profiles defined by the parameters given in Table 1. The profiles are time averaged in depth space. The lines show the median values of the respective terms, while shading shows the 5th–95th percentile range.

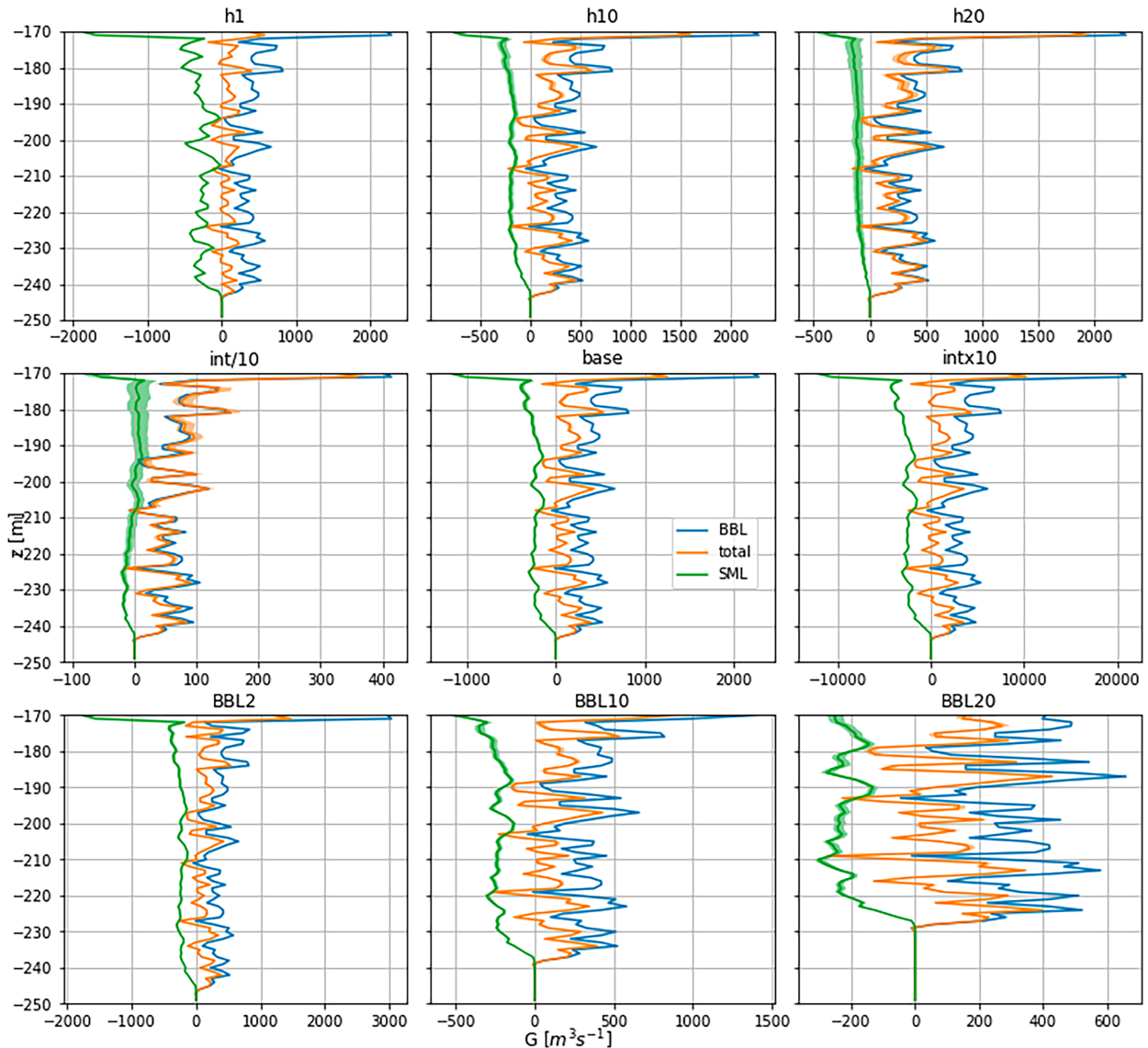
curve, suggesting that this result is likely to stand also for many other flux profiles. Next, we will consider what implications this has for the deep circulation in the basin.

4. Implications for the deep circulation

The G curves in Figs. 5 and 6 show many depth ranges where there is vertical convergence or divergence in the temporal mean. From Eq. (1), it is evident that if $G(b_1, t) < G(b_2, t)$ and $b_1 > b_2$, then the volume $V(b_1, t) - V(b_2, t)$ must increase with time. Given that the volume between two fixed depths is constant in time, such volume changes affect the stratification [i.e., $b(z)$]. For example, ΔG in the limit $\lim_{b_2 \rightarrow b_1}$ represents a stretching or compression of the $b(z)$ curve depending on its

sign. Inferring the temporal evolution of $b(z)$ implied by the $G(z)$ curves shown here is difficult: first, because the time averaging of G is done in z rather than in b space, and second, because we do not know the temporal evolution of either f or κ .

Tackling the second issue is beyond our scope here, but averages of G in buoyancy space are shown in Figs. 9 and 10 for the base cases with vertical f and κ profiles, respectively. The right panels of the two figures are zoomed in versions of their respective left panels. The left panels show only the b range that on average occupies the depth range of 170–250 m, while the b range in the right panels goes from a 5th percentile b at 250 m to a 95th percentile b at 170-m depth. It is clear from the figures that the areas of strong vertical convergence

FIG. 6. As in Fig. 5, but using κ instead of f profiles.

and divergence in the time-averaged G curves that were very apparent in the depth-averaged cases are not present in the buoyancy-averaged cases. The picture that emerges from the buoyancy-averaged view is instead one of rather uniform upwelling in the BBL and downwelling in the SML. The variability is higher at higher b levels because these b levels occur at shallower depths, sometimes in and above the permanent halocline. Together the buoyancy and depth-averaged G curves give an interesting view of the water mass transformation in the deep parts of the EGB. It can be summarized as follows: Multiple depth ranges are hotspots for water mass transformation and the water situated there undergoes rapid transformation. However, when averaged in buoyancy space, there are no apparent transformation hotspots; rather, we find transformation rates that are similar for large buoyancy ranges. Thus, the action of the transformation hotspots is naturally spread fairly even

over the water masses that typically occupy the depth range of interest here, leading to similar transformation rates for most water masses.

Another interesting aspect of the buoyancy-averaged figures is that they show rather similar net transformation rates (i.e., G), even though G is considerably larger at many depths when averaged in depth space for the f curves than for the κ curves. Our net transformation rates can be compared to earlier estimates of basin residence time and yearly salt loss to give an indication of their validity as an approximation of real EGB buoyancy flux curves. Feistel et al. (2006) estimated a residence time of 21 years for deep water in the EGB using a two-box model. However, this estimate included the effect of inflows. Feistel et al. (2006) hence concluded that the observed salt loss of about 0.15 psu per year during the 1983–93 stagnation period suggests a residence time of about a century

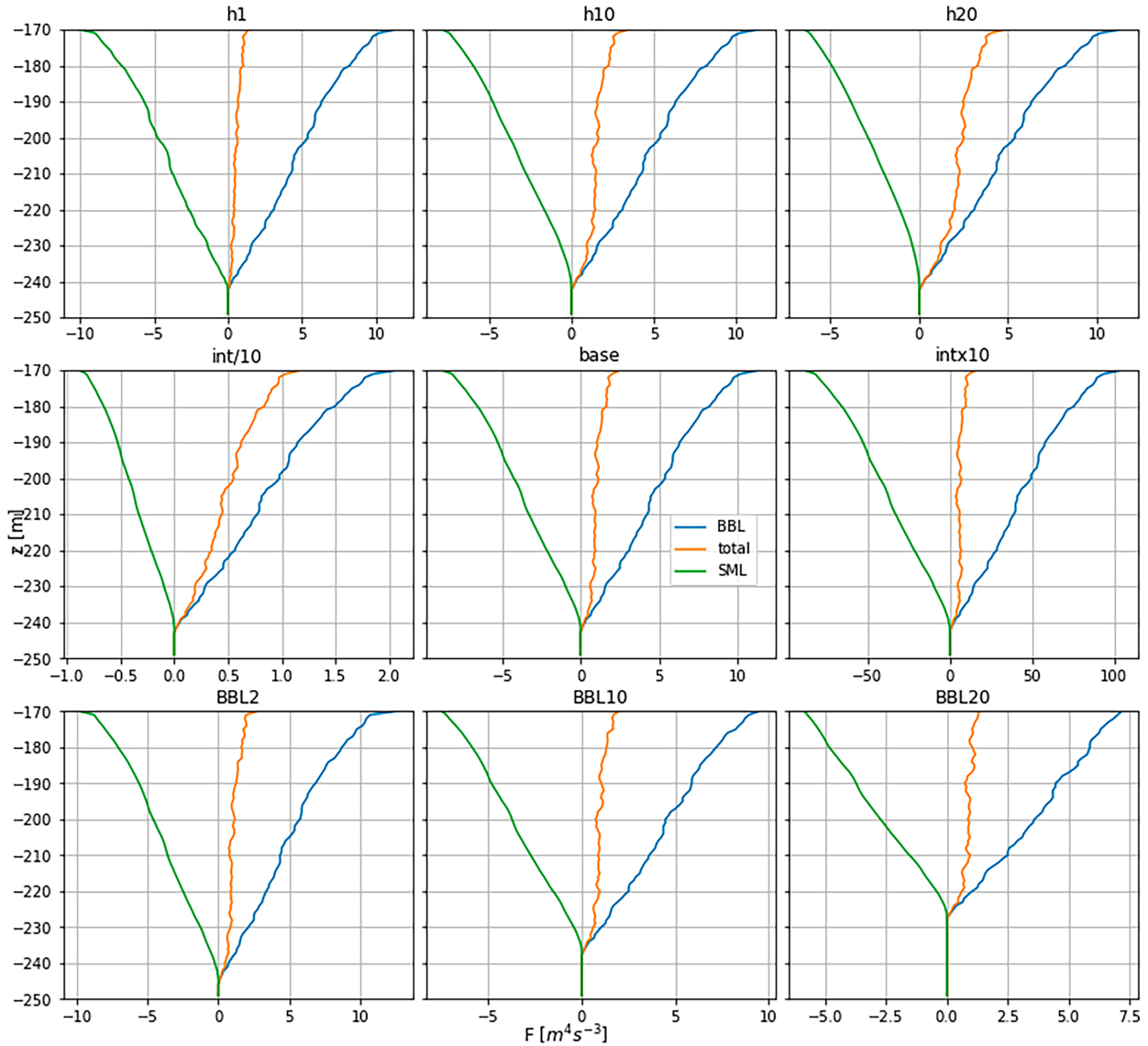


FIG. 7. Vertical profiles of F , F_{BBL} , and F_{SML} for the different f profiles defined by the parameters given in Table 1. The profiles are time averaged in depth space. The lines show the median values of the respective terms, while shading shows the 5th–95th percentile range.

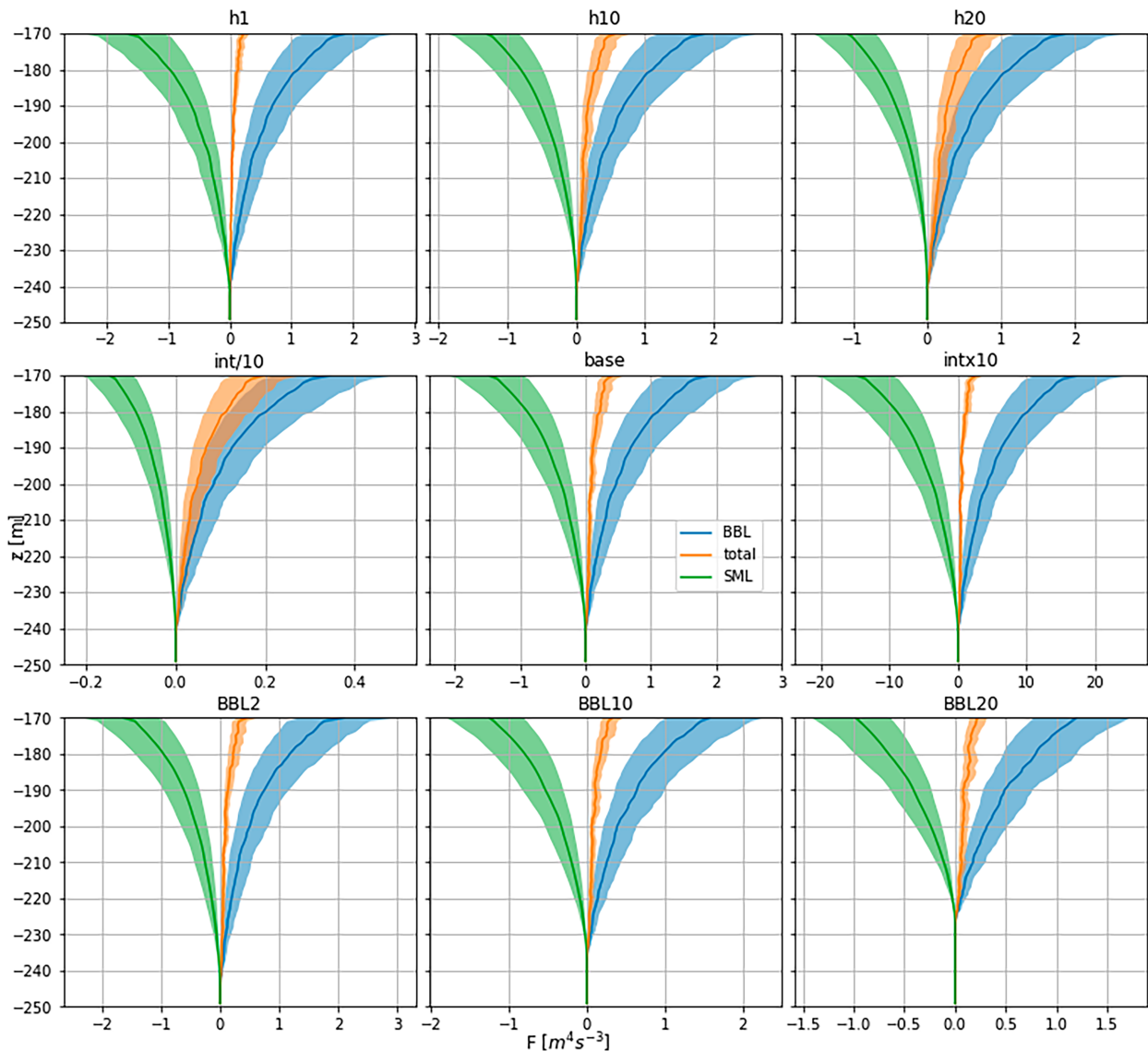
in the absence of inflows. From bulk arguments, $\tau = V/Q$, where τ is the residence time, Q is the freshwater supply, and V is the volume of the Baltic Sea; Meier and Kauker (2003) estimated a residence time of 35 years for the whole Baltic Sea. From our calculations, one may similarly estimate a residence time as, $\tau = V/G$, for the volume below 170 m. Yielding estimates of 3 and 6 years for the f and κ cases, respectively. Our estimates thus deviate substantially from earlier estimates, but the large difference is a result of differing methodologies and those residence times are thus not comparable to the estimate by Meier and Kauker (2003). Yearly buoyancy gain at a fixed depth level on the other hand can be estimated from

$$\frac{\partial b}{\partial t} \approx \frac{G(b')N^2(b')s_{\text{year}}}{A(b')}, \quad (20)$$

where $G(b')$, $N^2(b')$, and $A(b')$ are the medians of the respective quantities calculated from all stratification profiles at the buoyancy $b' = 0.0125 \text{ m s}^{-2}$, and $s_{\text{year}} = 31\,536\,000$ is the amount of seconds in a year. Plugging the numbers in gives an estimate of the yearly buoyancy gain at the given level. The buoyancy gain can then be related to a salt loss through

$$\frac{\partial S}{\partial t} \approx \frac{\partial b}{\partial t} \frac{\partial S}{\partial b} \approx \frac{\partial b}{\partial t} \frac{1}{g\beta}, \quad (21)$$

where β is the haline contraction coefficient, here estimated using TEOS-10 for $T = 6^\circ\text{C}$, $S = 12$ psu, and at a depth of 170 m. Moreover, a simple estimate of stratification-induced variability can be derived through substituting the median $G(b')$ in Eq. (20) by the 5th and 95th percentile of $G(b')$. The

FIG. 8. As in Fig. 7, but using κ instead of f profiles.

resulting salt loss is 0.07 ($-0.11, 0.42$) psu yr^{-1} for the f profile and 0.031 ($-0.004, 0.14$) psu yr^{-1} for the κ case. That is, our base f curve gives a yearly salt loss that is compatible with that observed during the 1983–1993 stagnation period, while that for the base κ curve is on the low end but not overly far from the observationally based estimate. It is also worth noting that the aforementioned stagnation period has a comparably fast salt loss as compared to other stagnation periods as is evidenced in Fig. 1, indicating that also the salt loss estimated for the base κ case might not be very far from reality.

A somewhat trickier comparison to observations can be done by applying similar ideas to those used above, to the sinking of the tracer patch injected by Holtermann et al. (2012). Those authors injected tracer at the isopycnal surface where $\rho = 9.9 \text{ kg m}^{-3}$. Two weeks later, they found peak tracer concentrations around the $\rho = 9.92 \text{ kg m}^{-3}$ isopycnal

surface. Assuming the patch underwent 2 weeks of sinking at a uniform velocity implies a $\partial b/\partial t = -1.6 \times 10^{-10} \text{ m s}^{-3}$. Using Eq. (20) with G_{SML} instead of G and without s_{year} gives estimates of -8.5×10^{-11} ($-1.5 \times 10^{-10}, -3.2 \times 10^{-11}$) m s^{-3} for the base case with the vertical f profile and -1.3×10^{-11} ($-2.6 \times 10^{-11}, -3.9 \times 10^{-12}$) m s^{-3} for the base case with the vertical κ profile. In both cases, we find implied sinking velocities that are somewhat smaller than those observed. However, the uncertainties in these computations are very large, so it is hard to see the result as either a refutation or a validation of either the buoyancy flux curves or the hypothesis that the observed sinking was influenced by the SML dynamics. On the one hand, the tracer patch could have done all its sinking in a much shorter time than 2 weeks between the observations, making the disagreement larger. On the other hand, our estimated $\partial b/\partial t$ is a bulk estimate for the whole $\rho = 9.9 \text{ kg m}^{-3}$ isopycnal surface, while we

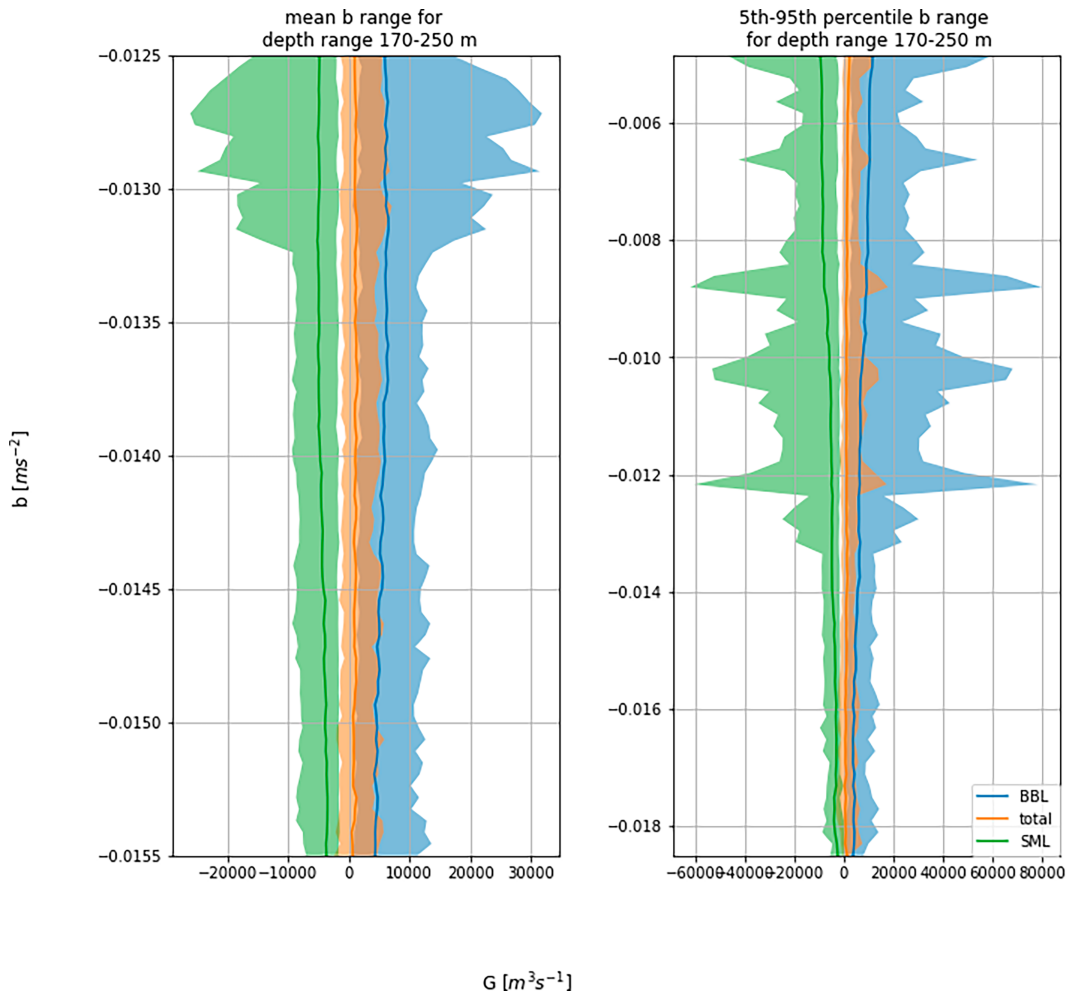


FIG. 9. Vertical profiles of G , G_{BBL} , and G_{SML} for the base f profile defined by the parameters given in Table 1. The profiles are time averaged over the buoyancy space. (left) The various G terms over the buoyancy range that on average occupies the depth range of 170–250 m. (right) The buoyancy range from a 5th percentile b at a depth of 250 m to a 95th percentile b at a depth of 170 m. The lines show the median values of the respective G terms, while shading shows the 5th–95th percentile range.

know that the sinking is only substantial in areas sufficiently close to the boundary, suggesting that the implied sinking could also be underestimated. Moreover, we know mixing to be highly variable in both time and space in ways that are not captured in our simple profiles. Thus, it seems fair to consider the SML dynamics as a plausible mechanism explaining the observed sinking, or at least part of it, but conclusive evidence is clearly lacking.

Apart from the vertical fluxes that form the basis of our analyses, strong lateral transport processes are also needed for the observations of nearly flat isopycnal surfaces and intense boundary mixing to be compatible. In the global ocean, it is thought that baroclinic eddies play a major role in lateral homogenization along isopycnal surfaces (Griffies et al. 1998; Hieronymus and Nycander 2013a). The Baltic Sea also has baroclinic eddies, often called Beddies, but their potential role in lateral mixing in the EGB is not well understood (Reissmann et al. 2009). In fact, generally speaking, it is not

entirely well understood how the lateral mixing in the deep parts of the EGB is achieved. However, several things are known. Holtermann et al. (2014) found that lateral homogenization of tracer patches released in a modeling study was relatively rapid, with a time scale for homogenization of about 100 days, which also agrees well with estimates in Holtermann et al. (2012). These authors also managed to connect the lateral spreading to topographic waves and a basin scale rim current, suggesting the role of baroclinic eddies to be less prominent in the Baltic Sea than in the global ocean for lateral spreading.

For each stratification and every vertical f and κ profile, we calculate the amplification factor $\mathcal{A}$. These profiles are quite noisy, as the denominator G in Eq. (12) takes on small values at some depths. Table 2 shows the vertical median of $\mathcal{A}$ in the 170–250-m range based on the median G profiles from Figs. 5 and 6. The value of $\mathcal{A}$ varies with the flux profile, but the more plausible flux curves give values between 2 and 3. It thus seems likely that the net diapycnal flow G is a factor of

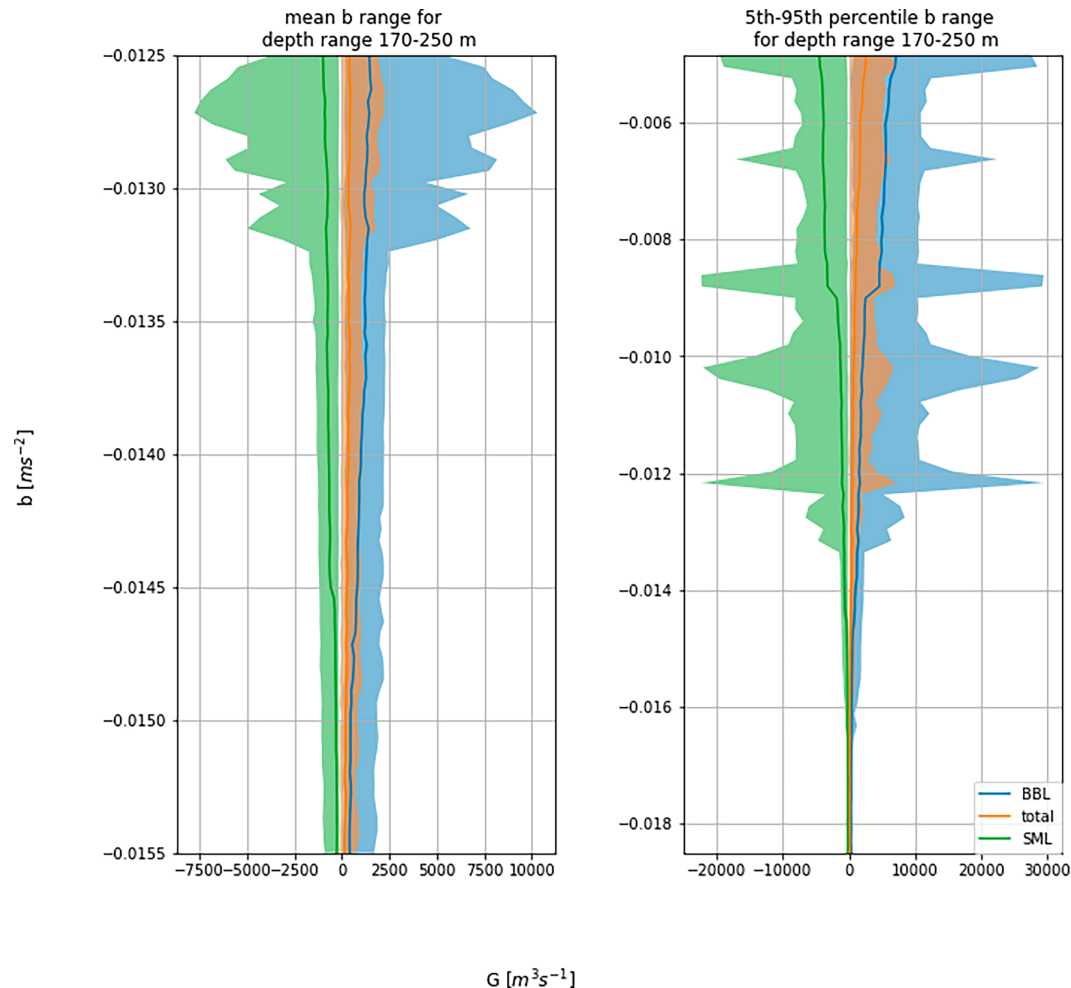


FIG. 10. As in Fig. 9, but using κ instead of f profiles.

2–3 smaller than its SML and BBL components. Moreover, it is evident from Eq. (13) that the velocity in the BBL should be much larger than that in the SML. An amplification factor between 2 and 3 was also found for the global ocean in model experiments by Ferrari et al. (2016), while a somewhat higher amplification factor reaching up to 5 was found for the global ocean in calculations similar to those done here by McDougall and Ferrari (2017). The amplification factor found here for the

EGB is thus in broad agreement with those found for the global ocean.

5. Conclusions

The main result of this short paper is that the deep diapycnal flow in the EGB is strongly controlled by the basin’s hypsographic curve. A result that is reminiscent of de Lavergne et al. (2017) who found the deep overturning in the global ocean to similarly be controlled by its respective basin’s hypsographic curves. The results derived here are reached using only simple considerations about the buoyancy and volume budgets. Nothing is assumed about the momentum budget nor is a steady state assumed. Given the generality of these assumptions and the fact that the results stand for a large range of different stratifications, vertical buoyancy flux curves, and diffusivity curves, we believe them to be very robust.

In fact, as discussed earlier, the structure of the diapycnal flow implied here, with sinking in the SML and rising in the BBL may already have been, at least, partly validated by observations and very recently also in a numerical ocean

TABLE 2. Median $\mathcal{A}$ values for the different κ and f profiles.

	$\kappa(z)$	$f(z)$
Base	2.3	2.8
int/10	1.0	1.7
intx10	2.6	3.2
h1	3.2	4.3
h10	1.8	2.3
h20	1.3	2.0
BBL2	2.4	2.8
BBL10	2.3	2.7
BBL20	2.2	2.7

model (Henell et al. 2023). Holtermann et al. (2012) and Holtermann and Umlauf (2012) reported initial sinking and later rising during a tracer-release experiment in the EGB. The initial release was far from the boundary, and the rising started when the tracer patch came close to the boundary. The observed sinking could conceivably have been the work of chance or of other sinking mechanisms (Holtermann et al. 2012), but the similarity between the observed circulation and that derived here and in an ocean model by Henell et al. (2023) is nonetheless intriguing. Further work is certainly needed to better constrain the diapycnal velocity, the buoyancy flux curve, and their spatial and temporal variability in the EGB, for a proper validation of the theory to be possible. That is, not only G but also its forcing F is highly uncertain. Thus, similarities found here between observations and theory are essentially qualitative in nature, and more independent measures of F and G would be needed for a true quantitative test to be possible. Nevertheless, it should also be said that these problems are much easier to tackle in a data-rich coastal sea than in the comparably data-poor global ocean. The EGB might therefore in the future prove a valuable laboratory model where data needed for quantitative tests are much easier to gather than in the vastly bigger and deeper global ocean.

Apart from having a vertical profile strongly dependent on the hypsographic curve, it is also noteworthy that G , in our calculations, depends rather weakly on the stratification, in fact, very weakly for the vertical diffusivity profiles. This may have implications for the sensitivity of the diapycnal flow to climate change, as such changes imply stratification changes. However, one should be careful not to overinterpret, given that our model is not complete. First, it lacks the important inflow dynamics, meaning that it is in practice a model for the diapycnal flow during stagnation periods. Second, our f and κ profiles are very simplistic and may need to be complemented with additional transport processes. An interesting avenue of future research would be to quantify water mass transformation in the EGB in long integrations with numerical ocean models, where time-varying process-by-process estimates of water mass transformation can be computed as in Hieronymus et al. (2014). Another interesting finding regarding the strong control of the hypsographic curve on the transformation was the presence of transformation hotspots in certain depth ranges. Also interesting was that we found those hotspots to affect the water masses typically situated in the 170–250-m-depth range very similarly (i.e., G was similar for a large range of b), as was evidenced by the absence of such hotspots when G was averaged in buoyancy space.

Last, while the vertical structure of G is insensitive to changes in the magnitude of f and κ (i.e., to changes in B_0 and K_0) in our experiments. The strength of the overturning, the magnitude of G , is not. This is evidenced by the int/10 and intx10 cases that suggest that an order of magnitude increase in B_0 or K_0 implies a similar increase in G . In the context of climate change, our results do therefore not suggest that the magnitude of G could not change significantly in a warmer world. However, it does seem unlikely that the vertical structure of G during stagnation periods would be much affected by climate change or other perturbations likely to occur on time scales of decades to centuries, as it seems to depend primarily on the hypsographic curve.

Our results could potentially also have implications for nutrient cycling in the basin. The deep parts of the Baltic Sea are, as mentioned in the introduction, often oxygen depleted owing to poor ventilation and eutrophication. Our results suggest that upwelling and thus nutrient transport from the deeper parts toward the surface occur predominantly in a boundary layer along the sloping seafloor. The depth range used for the computations in this study is quite far from the euphotic zone, but nevertheless it would be interesting to investigate whether the circulation pattern with upwelling around the basin rim and downwelling in the SML could affect the spatial patterns of primary production in the basin.

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Data availability statement. The data presented here are taken from the open data archive SHARK web (<https://sharkweb.smhi.se/hamta-data/>).

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