Ocean Subduction☆

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Glossary

Potential vorticity A dynamical tracer, conserved by fluid in adiabatic and inviscid flow, which is determined by the absolute spin of the fluid and the stratification.
Seasonal boundary layer A region over which the mixed layer and seasonal thermocline occur. The base of the seasonal boundary layer is defined by the maximum thickness of the winter mixed layer.
Subduction The transfer of fluid from the mixed layer into the stratified thermocline.
Subduction rate The volume flux per unit horizontal area passing from the mixed layer into the stratified thermocline.
Thermocline A region of enhanced vertical temperature gradient over the upper 1 km of the ocean.
Ventilation The transfer of fluid from the mixed layer into the ocean interior.

Units

1 Sv 10^6 m^3 s^{-1}

Introduction

Ocean subduction involves the transfer of fluid from the mixed layer into the stratified thermocline (Fig. 1). The upper ocean is ventilated principally through the subduction process, while the deep ocean is ventilated through open-ocean convection and

☆*Change History*: May 2018. Richard G Williams and Andrew Meijers updated the subsection "Competing paradigms: gyre-scale subduction versus eddy stirring," added new section "Subduction in the Southern Ocean," updated Conclusions, added Figs. 10, 15, 16, and 17, and updated Further Reading and References and also completed minor editing of the text.

This is an update of R.G. Williams, Ocean Subduction, *Encyclopedia of Ocean Sciences* (2nd Edn), edited by John H. Steele, Academic Press, 2001, pp. 156–166.

 https://doi.org/10.1016/B978-0-12-409548-9.11297-7

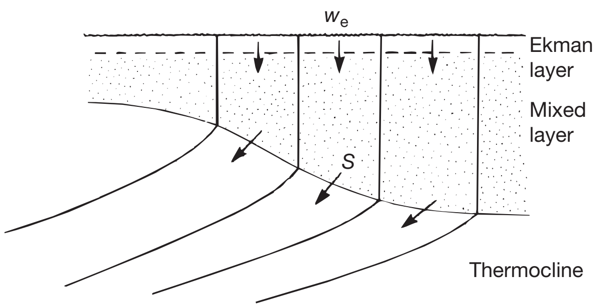

Fig. 1 A schematic diagram showing isopycnals in the thermocline outcropping into a vertically homogeneous mixed layer. The subduction rate, *S*, measures the rate at which fluid passes into the stratified thermocline through the vertical and horizontal transfer from the mixed layer. In comparison, the wind-induced, Ekman downwelling, $-w_{e,}$ measures the vertical flux pumped down from the surface Ekman layer. Reproduced with permission from Marshall, J.C., Nurser, A.J.G., and Williams, R.G. (1993). Inferring the subduction rate and period over the North Atlantic. *Journal of Physical Oceanography* **23**, 1315–1329.

cascading down coastal boundaries. The term "ocean subduction" makes the geologicial analogy of a subduction zone where a rigid plate of the Earth's lithosphere slides beneath a more buoyant plate and into the hotter part of the mantle.

Ventilation connects the atmosphere and ocean interior through the transfer of fluid from the surface mixed layer into the ocean interior. Water masses are formed in the surface mixed layer and acquire their characteristics through the exchange of heat, moisture and dissolved gasses with the atmosphere. When the water masses are transferred beneath the mixed layer, they are shielded from the atmosphere and only subsequently modify their properties by mixing in the ocean interior. Hence, the ventilation process helps to determine the relatively long memory of the ocean interior, compared with the surface mixed layer.

Subduction occurs throughout the global ocean: over recirculating wind-driven gyres within ocean basins (with horizontal scales of several thousand kilometers), across frontal zones and convective, overturning chimneys (on horizontal scales of several hundred to tens of kilometers). The reverse of the subduction process leads to the transfer of fluid from the main thermocline into the seasonal boundary layer, which affects the properties of the mixed layer and air–sea interaction; this negative subduction is sometimes referred to as obduction.

Subduction Process

The subduction process involves the seasonal cycle of the mixed layer (Fig. 2). The mixed layer is vertically homogenous through the action of convective overturning and turbulent mixing. At the end of winter, the mixed layer is at its maximum density and

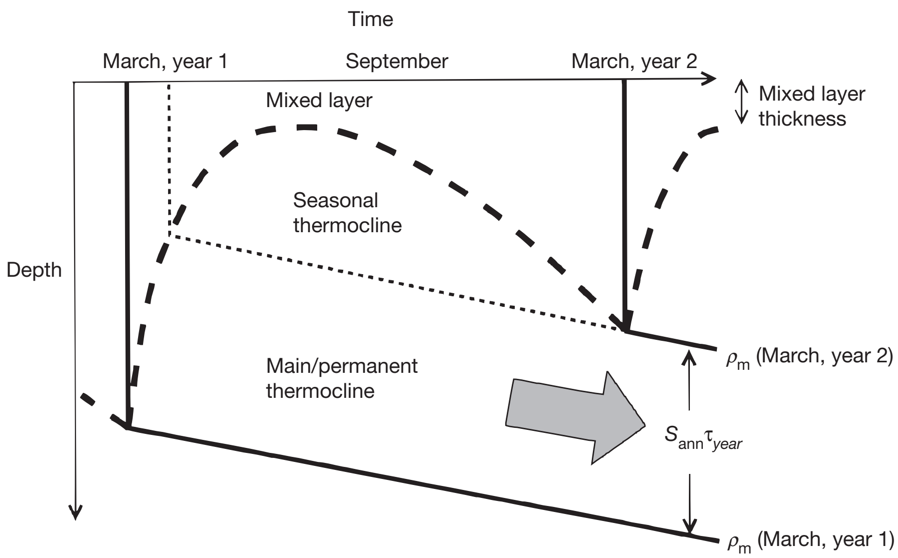

Fig. 2 A schematic diagram illustrating the seasonal cycle of the mixed layer following the movement of a water column. The mixed layer thins in spring and summer, and thickens again in autumn and winter. If there is an overall buoyancy input, the end of winter mixed layer becomes lighter and thinner from 1 year to the next (as depicted here). Consequently, fluid is subducted irreversibly from the mixed layer into the main or permanent thermocline. The mixed layer thickness is marked by the *thick dashed line*, and isopycnals ρ_m outcropping at the end of winter into the mixed layer by the *full lines*, and the isopycnal identifying the base of the seasonal thermocline by the *short-dashed line*. The annual subduction rate, S_{ann}, determines the vertical spacing between the isopycnals subducted from the mixed layer in March for consecutive years 1 and 2 (τ_{year} is 1 year). Further details see Williams and Follows (2011).

thickness, and overlies the main thermocline (where there is a strong vertical temperature gradient). During spring and summer, the surface warming forms a seasonal thermocline, which is capped by a thin, wind-stirred, mixed layer. During autumn and winter, the cooling of the surface ocean leads to a buoyancy loss and convective overturning. The mixed layer thickens and entrains fluid from the underlying thermocline until the cooling phase ceases at the end of winter.

Fluid is subducted from the mixed layer into the thermocline during the warming in spring and summer, whereas fluid is entrained into the mixed layer from the thermocline during the cooling in autumn and winter. Whether there is annual subduction of fluid into the thermocline depends on the buoyancy input into a water column. If there is a buoyancy input over an annual cycle (as depicted in Fig. 2), the mixed layer at the end of winter becomes lighter and thinner, than at the end of the previous winter. In this case, fluid is subducted into the main or permanent thermocline over the annual cycle.

Conversely, if there is a buoyancy loss over an annual cycle, the mixed layer at the end of winter becomes denser and thicker, than at the end of the previous winter. In this case, there is negative subduction and an annual transfer of fluid from the main thermocline into the seasonal boundary layer (defined by the maximum thickness of the winter mixed layer).

Seasonal Rectification

The subduction process leads to an asymmetrical coupling between the mixed layer and main thermocline. For example, the temperature and salinity relation in the main thermocline is observed to match that of the winter mixed layer, rather than the annual average of the mixed layer. This biased coupling is due to the seasonal cycle of the mixed layer. Isopycnals in the mixed layer (density outcrops) migrate poleward in the heating seasons and retreat equatorward in the cooling seasons. This seasonal displacement of density outcrops is much greater than the movement of fluid particles over an annual cycle (Stommel, 1979). Consequently, fluid subducted from the mixed layer in summer is re-entrained into the mixed layer during the subsequent cooling seasons (Fig. 3) (through the equatorial migration of density outcrops and winter thickening of the mixed layer). Fluid only succeeds in being permanently subducted into the main thermocline over a short window, 1 to 2 months long, at the end of winter when the density outcrops are at the equatorial end of their cycle (Williams et al., 1995). Hence, there is a seasonal rectification in the transfer of water-mass properties from the mixed layer into the ocean interior. Idealized tracer experiments suggest that the seasonal rectification still occurs even in the presence of a vibrant, time-varying circulation.

Formation of Mode Waters

There are local maxima in subduction and ventilation processes leading to the formation of characteristic "mode" waters, which are apparent in a volumetric census of the temperature and salinity characteristics of the ocean. These mode waters consist of large volumes of weakly stratified fluid with nearly vertically homogeneous properties. For example, strong buoyancy loss to the

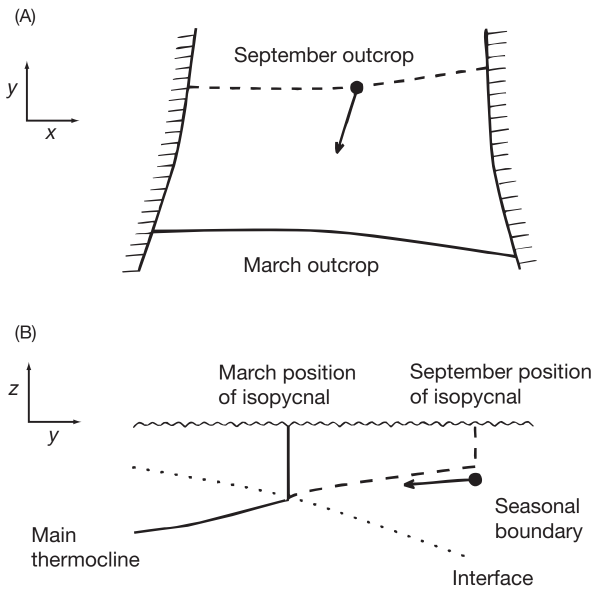

Fig. 3 A schematic diagram illustrating the seasonal rectification of subduction for (A) a plan view at the sea surface and (B) a meridional section. Over the cooling season in the Northern Hemisphere the outcrop of a density surface migrates equatorwards from September to March (denoted by *dashed* and *full lines*). Consider a fluid parcel (circle) subducted from the vertically homogeneous mixed layer and advected equatorwards along this density surface. If the fluid parcel is advected past the March outcrop during the year, then it is subducted into the main thermocline (the upper extent of which is shown by the *dotted line*), otherwise it is eventually entrained into the mixed layer. Reproduced with permission from Williams, R.G., Spall, M.A., and Marshall, J.C. (1995). Does Stommel's mixed-layer "Demon" work? *Journal of Physical Oceanography* **25**, 3089–3102.

atmosphere over the Gulf Stream leads to the formation of 18°C mode water within the mixed layer, which is subducted following a buoyancy input along the anticyclonic circulation south of the Gulf Stream (Woods and Barkmann, 1986). Signals of mode water formation over the North Atlantic are revealed in diagnostics of water-mass formation from surface buoyancy fluxes (shown later in Fig. 12).

Definitions of Subduction Rate

The instantaneous subduction rate, S, the volume flux from the mixed layer into the stratified thermocline (Cushman-Roisin, 1987; Marshall et al., 1993) is given by

$$S = -\frac{\partial h}{\partial t} - w_b - \boldsymbol{u}_b.\nabla h \equiv -w_b - \frac{D_b h}{Dt}, \tag{1}$$

where S is defined as positive when fluid is transferred into the thermocline (Fig. 4); h is the thickness of the mixed layer, w_b and $\boldsymbol{u}_b$ are the vertical velocity and horizontal velocity vector, respectively, at the base of the mixed layer, and $D_b/Dt \equiv \partial/\partial t + \boldsymbol{u}_b.\nabla$ is the Lagrangian rate of change following the horizontal flow at the base of the mixed layer. The units of S are given by a volume flux ($m^3\,s^{-1}$) per unit horizontal area (m^2), so are equivalent to the units for a velocity ($m\,s^{-1}$). The subduction rate is defined as the volume flux per unit horizontal area in Eq. (1) *relative* to the base of the instantaneous mixed layer. Hence, S includes a contribution from the change in thickness of the mixed layer, as well as from the vertical and horizontal movement of fluid particles. Equivalently, S in Eq. (1) can be evaluated from the vertical velocity and the Lagrangian change in mixed-layer thickness following the horizontal circulation.

The subduction into the main thermocline is physically more important, than that into the seasonal thermocline, since water-mass properties become shielded from the atmosphere for longer timescales within the main thermocline. The annual rate of subduction into the main thermocline, S_{ann}, is evaluated from the volume flux per unit horizontal area passing through a control surface, H, which overlies the main thermocline (Marshall et al., 1993):

$$S_{ann} = -w_H - \boldsymbol{u}_H.\nabla H \tag{2}$$

where w_H and $\boldsymbol{u}_H$ are the vertical velocity and horizontal velocity vector at the depth H. The control surface is defined by the base of the seasonal thermocline given by the maximum thickness of the winter mixed layer (Fig. 1).

Alternatively, S_{ann} may be evaluated directly from S in Eq. (1) using a Lagrangian integration following the movement of a water column over an annual cycle (Williams et al., 1995). For example, for the mixed-layer cycle shown in Fig. 2, S_{ann} controls the vertical spacing between isopycnals subducted at the end of consecutive winters. Hence, a high rate of subduction leads to the formation of a mode water with a weak stratification (and a large vertical spacing between subducted isopycnals).

Gyre-Scale Subduction

Subduction occurs over ocean basins predominately through the gyre-scale circulation. The surface wind-stress drives anticyclonic and cyclonic recirculations, referred to as subtropical and subpolar gyres respectively, within a basin. The wind forcing induces

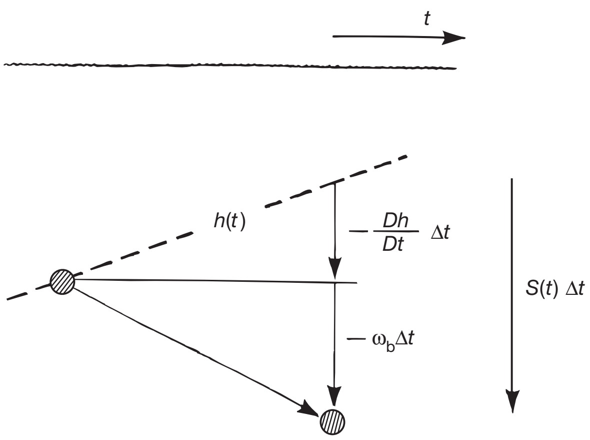

Fig. 4 A schematic diagram showing a particle being subducted from the time-varying base of the mixed layer into the thermocline. The vertical distance between the particle and the mixed layer, $S(t)\Delta t$, is the sum of the vertical displacement of the particle, $-w_b\Delta t$, and the shallowing of the mixed layer following the particle, $-(Dh/Dt)\Delta t$, where h is the thickness of the mixed layer, w_b is the vertical velocity and Δt is a time interval. Reproduced with permission from Marshall, J.C., Nurser, A.J.G., and Williams, R.G. (1993). Inferring the subduction rate and period over the North Atlantic. *Journal of Physical Oceanography* **23**, 1315–1329.

downwelling of surface fluid over the subtropical gyre and upwelling over the subpolar gyre. The gyre-scale subduction rate defined in Eq. (2) depends on both the vertical and horizontal circulations together with the thickness of the end of winter mixed layer.

Subduction Rate Over the North Atlantic

The North Atlantic is the most actively ventilated oceanic basin. The wind-induced downwelling reaches magnitudes of 25 to 50 m year^{-1} over the North Atlantic (Fig. 5A), which is characteristic of most basins. However, the annual subduction rate is significantly enhanced over the vertical transfer through the lateral transfer across the sloping base of the mixed layer (Marshall et al., 1993). The surface buoyancy loss to the atmosphere leads to the winter mixed layer thickening poleward, from typically 50 m in the subtropics (e.g., at 20°N) to 500 m or more in the subpolar gyre (e.g., at 50°N).

Over the subtropical gyre, the annual subduction rate reaches between 50 and 100 m year^{-1} (Fig. 5B). There is a band of high subduction rates south of the Gulf Stream, which are controlled by the lateral transfer. Elsewhere, the vertical and horizontal transfers are comparable over the subtropical gyre.

Over the subpolar gyre, fluid is generally transferred from the main thermocline into the seasonal boundary layer. Fluid is eventually returned from the seasonal boundary layer and into the interior through deep convection events or through subduction along the western boundary of the subpolar gyre. The negative subduction rates reach several 100 m year^{-1} over the North Atlantic (Fig. 5B), which is controlled by the lateral transfer rather than the vertical transfer. This negative subduction leads to water-mass properties of the main thermocline being transferred into the downstream winter mixed layer. This negative subduction is particularly important for the biogeochemistry in transferring inorganic nutrients from the thermocline into the winter mixed layer. These nutrients are entrained into the euphotic zone and enable high values of biological production to occur.

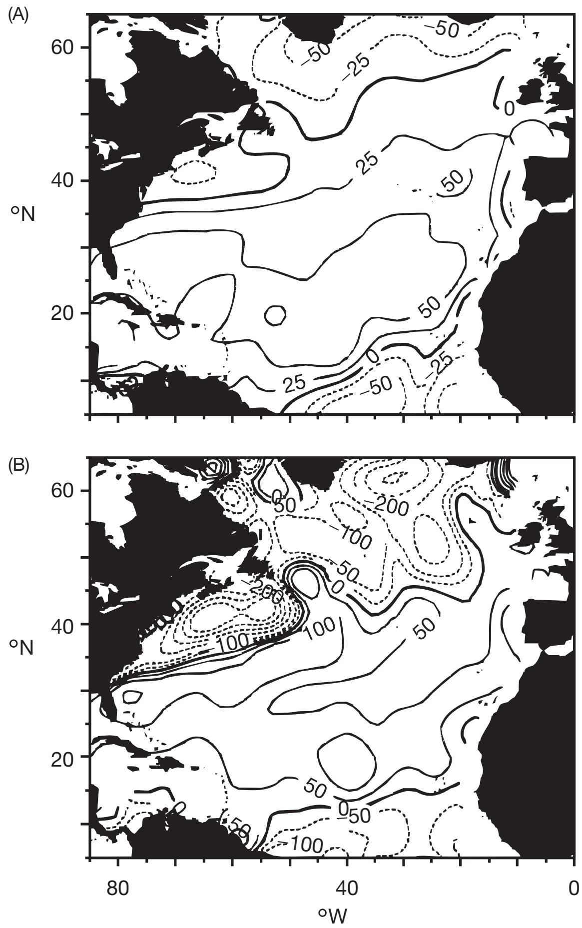

Fig. 5 Diagnostics for (A) wind-driven, Ekman downwelling velocity (contours every 25 m year^{-1}) and (B) annual subduction rate into the main thermocline of the North Atlantic (contours every 50 m year^{-1}). The annual subduction rate is evaluated from $S_H = -w_H - \boldsymbol{u}_H \cdot \nabla H$ with the interface H defined from the mixed-layer thickness at the end of winter, $\boldsymbol{u}_H$ evaluated from a density climatology using thermal-wind balance, and w_H derived from the wind-stress climatology and linear vorticity balance. The subduction rate represents the volume flux per unit horizontal area which ventilates the stratified thermocline. Reproduced with permission from Marshall, J.C., Nurser, A.J.G., and Williams, R.G. (1993). Inferring the subduction rate and period over the North Atlantic. *Journal of Physical Oceanography* **23**, 1315–1329.

Evidence From Transient Tracers

Characteristic signals of subduction and ventilation are provided by transient-tracer distributions, such as tritium and CFCs, over the ocean interior. Quantitative information on the ventilation process is provided when the transient tracers have a known source function and lifetime. For example, in the North Atlantic, the tritium–helium age distribution along potential density surfaces suggests a broad-scale effective ventilation of the subtropical gyre from the northeast and gradual aging following the anticyclonic circulation (Fig. 6) (Jenkins, 1987, 1988). The invasion of transient tracers into the main thermocline provides an integrated measure of ventilation including the contribution of the time-mean circulation and the rectified contribution of eddies. For example, the rate of ventilation inferred from the tracer age distribution is two to three times greater than the rate of wind-induced (Ekman) downwelling, which is possibly consistent with the gyre-scale, subduction rate diagnostics in Fig. 5. However, the rate of ventilation inferred from influx of tracers also includes a diffusive contribution, which differs for different tracers, making a more exact comparison with the subduction rate based on volume fluxes difficult to achieve.

Subduction and the Main Thermocline

Formation of the Main Thermocline

Throughout the ocean, there is a persistent thermocline separating the mixed layer and the weakly stratified, deep ocean. In ideal thermocline theory, the subduction process forms an upper thermocline over the subtropical gyre. The wind-driven circulation drives downwelling over a subtropical gyre, which leads to fluid being subducted from the mixed layer into the adiabatic interior of the ocean—see the seminal work of Luyten et al. (1983).

The thermocline is an advective feature formed through the circulation tilting the horizontal contrast in surface density over the subtropical gyre into the vertical. The subducted thermocline extends over the mixed-layer density range within the subtropical gyre. However, observations reveal that the thermocline also extends over denser isopycnals which do not outcrop within the subtropical gyre. Thermocline models incorporating diffusion suggest that this denser thermocline may be formed through the vertical convergence of downwelling of subtropical fluid and upwelling of denser fluid originating from outside the subtropical gyre, such as from the subpolar gyre or Southern Ocean (Samelson and Vallis, 1997).

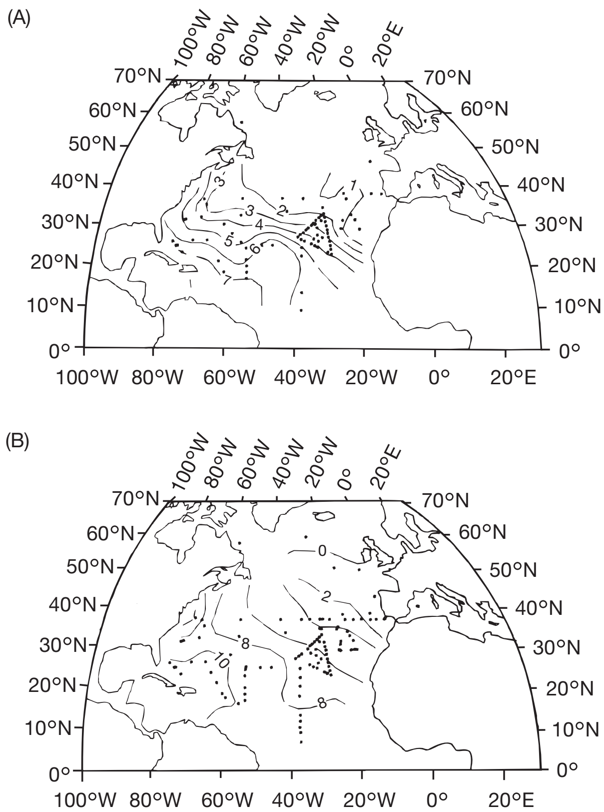

Fig. 6 Observations of the tritium–helium age (years) on potential density surfaces $\sigma_\theta = 26.5$ (upper panel) and $\sigma_\theta = 26.75$ (lower panel) in the North Atlantic. These surfaces are ventilated from the northeast and the tritium–helium age increases following the anticyclonic circulation of the subtropical gyre. Reproduced with permission from Jenkins, W.J. (1988). The use of anthropogenic tritium and helium-3 to study subtropical gyre ventilation and circulation. *Philosophical Transactions of the Royal Society of London A* **325**, 43–61.

Dynamical Tracer Distributions

The subduction process helps to determine the stratification within the main thermocline, and hence the distribution of a dynamic tracer, the potential vorticity. Fluid parcels conserve potential vorticity for adiabatic, inviscid flow. Potential vorticity depends on the absolute spin of the fluid and stratification. A large-scale estimate of potential vorticity is provided by

$$Q = -\frac{f}{\rho}\frac{\partial \sigma}{\partial z}, \tag{3}$$

where $f = 2\Omega \sin\phi$ is the Coriolis parameter or planetary vorticity, Ω is the Earth's angular velocity, ϕ is the latitude, σ is a potential density and ρ is a reference density. Climatological maps of large-scale potential vorticity over the main thermocline reveal different regimes consisting of open, closed or blocked Q contours along potential density surfaces (Fig. 7) (McDowell et al., 1982). The open Q contours thread from the stratified interior to the mixed-layer outcrop at the end of winter (Fig. 7A). Conversely, closed Q contours do not intersect the mixed layer and usually contain regions of nearly uniform Q (Fig. 7B). The blocked contours run zonally from coast to coast and are usually associated with weak meridional flow unless directly forced. The open Q contours dominate for lighter surfaces, whereas the closed or blocked Q contours dominate for denser surfaces.

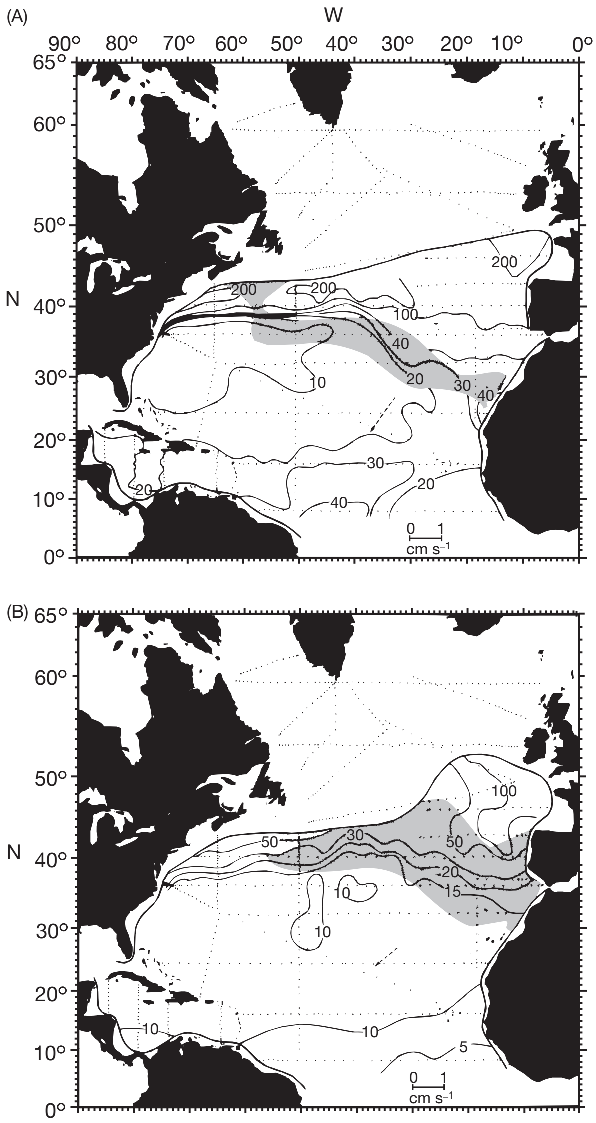

Fig. 7 Diagnosed distribution of a dynamic tracer, the large-scale potential vorticity Q (10^{-11} m^{-1} s^{-1}), over the North Atlantic: (A) between the potential density $\sigma_\theta = 26.3$ and 26.5 surfaces, the Q contours appear to be closed in northwest Atlantic, or to be open and thread back to the winter outcrop (*shaded area*) in the northeast Atlantic; and (B) between the $\sigma_\theta = 26.5$ and 27.0 surfaces, the Q field appears to be relatively homogeneous south of the winter outcrop over most of the subtropical gyre. Reproduced with permission from McDowell, S., Rhines, P.B., and Keffer, T. (1982). North Atlantic potential vorticity and its relation to the general circulation. *Journal of Physical Oceanography*, **12**, 1417–1436.

Competing Paradigms: Gyre-Scale Subduction Versus Eddy Stirring

Different hypotheses involving gyre-scale subduction or eddy stirring have been invoked to explain these contrasting tracer distributions seen in the data.

Open tracer contours for Q (such as in Fig. 7A) are usually associated with gyre-scale subduction. Fluid is subducted from the end of winter mixed layer and, when the flow is adiabatic and inviscid, streamlines become coincident with Q contours in the thermocline. A thermocline model example is shown in Fig. 8 where Q along potential density surfaces in the thermocline is determined by subduction from the overlying mixed layer (Williams, 1991). The isopleths of Q reveal the anticirculation of the subtropical gyre. The Q contrast in the thermocline can become small given realistic end of winter mixed-layer variations. In an ideal thermocline model, each potential density surface can be separated into a ventilated zone, which divides unventilated regions of a western pool and an eastern shadow zone (Luyten et al., 1983). The unventilated zones are not directly connected by the time-mean streamlines to the overlying mixed layer, although they may still be indirectly ventilated through mixing acting along the coastal boundaries. The ventilated zone extends over most of the subtropical gyre for light potential density surfaces, but contracts eastward for denser surfaces.

Nearly uniform tracer distributions (such as in Fig. 7B) are instead usually associated with stirring by mesoscale eddies. The eddy stirring can homogenize conserved tracers within closed streamlines (Rhines and Young, 1982). An eddy-resolving model example is shown in Fig. 9 for a pair of wind-driven gyres. Advection by the gyre circulation leads to tracer contours and streamlines becoming nearly coincident and closed over much of the interior domain. Eddy stirring transfers tracer down gradient within these

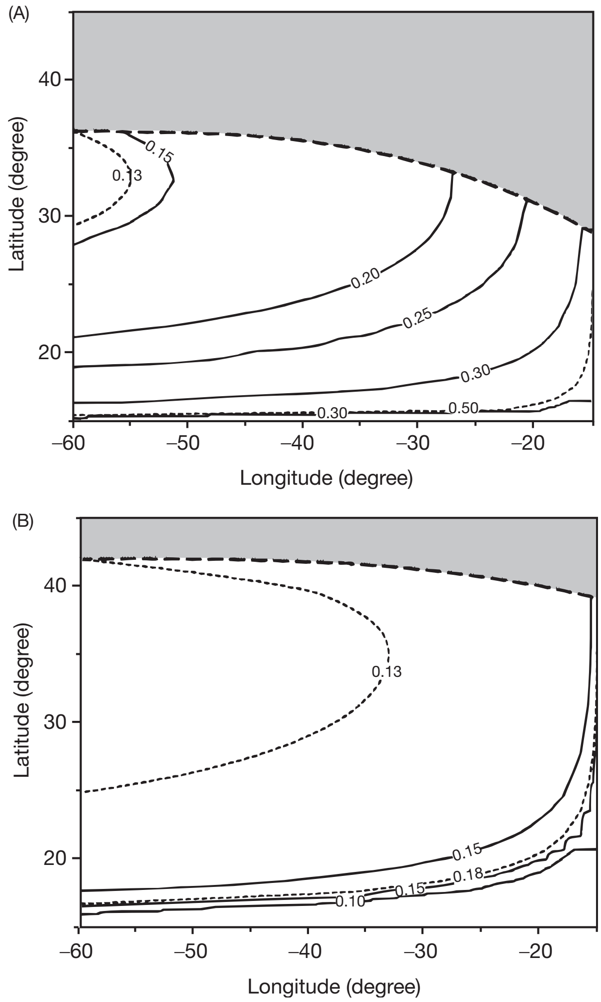

Fig. 8 An ideal thermocline model solution for the dynamic tracer, the large-scale potential vorticity Q (10^{-9} m^{-1} s^{-1}), along (A) potential density $\sigma_\theta = 26.4$ surface and (B) $\sigma_\theta = 26.75$ surface. Fluid is subducted across the density outcrop in the mixed layer (*thick dashed line*) into the stratified thermocline within a ventilated zone (between the *thin dashed lines*). Streamlines are coincident with the potential vorticity contours in the adiabatic interior. Given appropriate mixed-layer variations, the subducted potential vorticity can become nearly uniform, as obtained in (B). These model solutions are for an imposed anticyclonic wind forcing and a mixed layer, which becomes thicker to the north and denser to the north and east. Reproduced with permission from Williams, R.G. (1991). The role of the mixed layer in setting the potential vorticity of the main thermocline. *Journal of Physical Oceanography* **21**, 1803–1814.

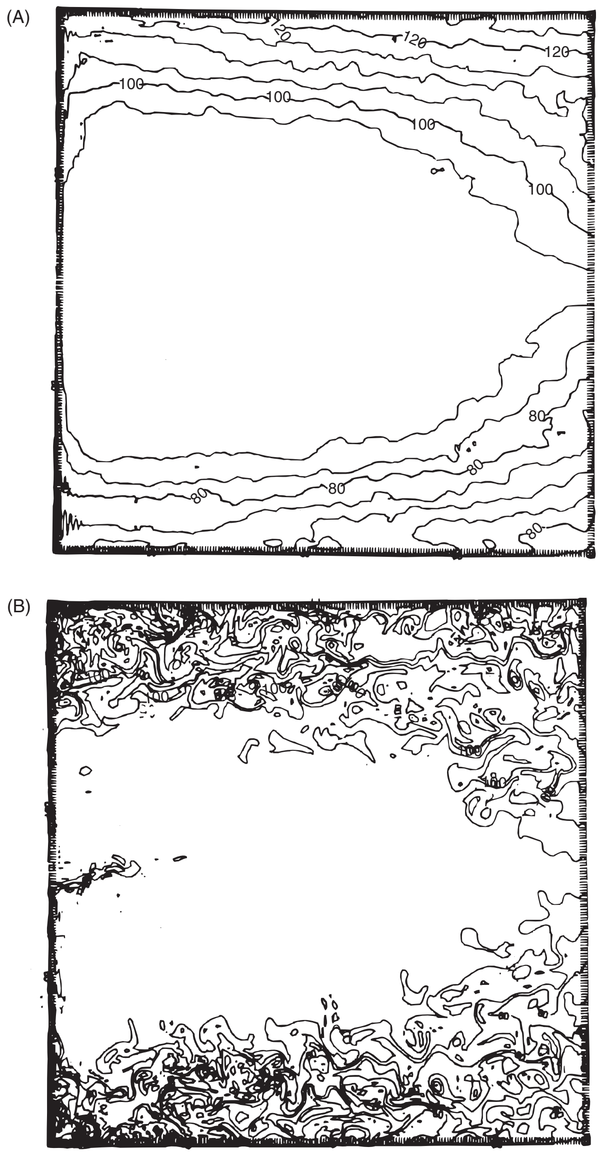

Fig. 9 An eddy-resolving, three layer, model solution for the dynamic tracer, potential vorticity, for a double, wind-driven gyre over a rectangular basin: plan views of (A) climatological mean and (B) instantaneous potential vorticity for the intermediate layer. The wind forcing drives a pair of wind gyres antisymmetric about the middle latitude, which are unstable generating a vigorous mesoscale eddy circulation. The eddy stirring leads to the time-averaged potential vorticity [in (A)] becoming nearly uniform within the circulating gyres, while outside the gyres there is a polewards increase arising from the meridional gradient in planetary vorticity. The instantaneous potential vorticity [in (B)] shows nearly perfect homogenization within the pair of gyres. At the edges of the gyre, the eddy activity is apparent through the winding up of potential vorticity contours. An experiment by W. Holland. Reproduced from Rhines, P.B. and Young, W.R. (1982). Homogenization of potential vorticity in planetary gyres. *Journal of Fluid Mechanics* **122**, 347–367, with permission from Cambridge University Press.

closed contours, which forms extensive regions of nearly uniform tracer. This interpretation is supported through observations of nearly uniform distributions of tritium, as well as potential vorticity, along weakly ventilated potential density surfaces within the wind-driven gyres of the North Pacific and North Atlantic.

Gyre-scale subduction and eddy stirring are not though mutually exclusive processes, and can occur simultaneously and modify each other. Ventilation helps control the input of tracer and the tracer contrast along the winter outcrop of the potential density surface. At the same time, eddies act to stir any tracer contrasts leading to a smearing out of subducted mode water signals in the ocean interior. On the fine scale, the ventilation process may be viewed in a chaotic manner (MacGilchrist et al., 2017), eddy stirring acting to draw out thin tracer filaments, so that interior properties connect to the surface ocean via a complicated set of pathways (Fig. 10).

In accord with this view, observational diagnostics from transient tracers and float trajectories suggest that the implied Peclet number (a nondimensional measure of advection/diffusion) is typically <10. This value is much smaller than the high value

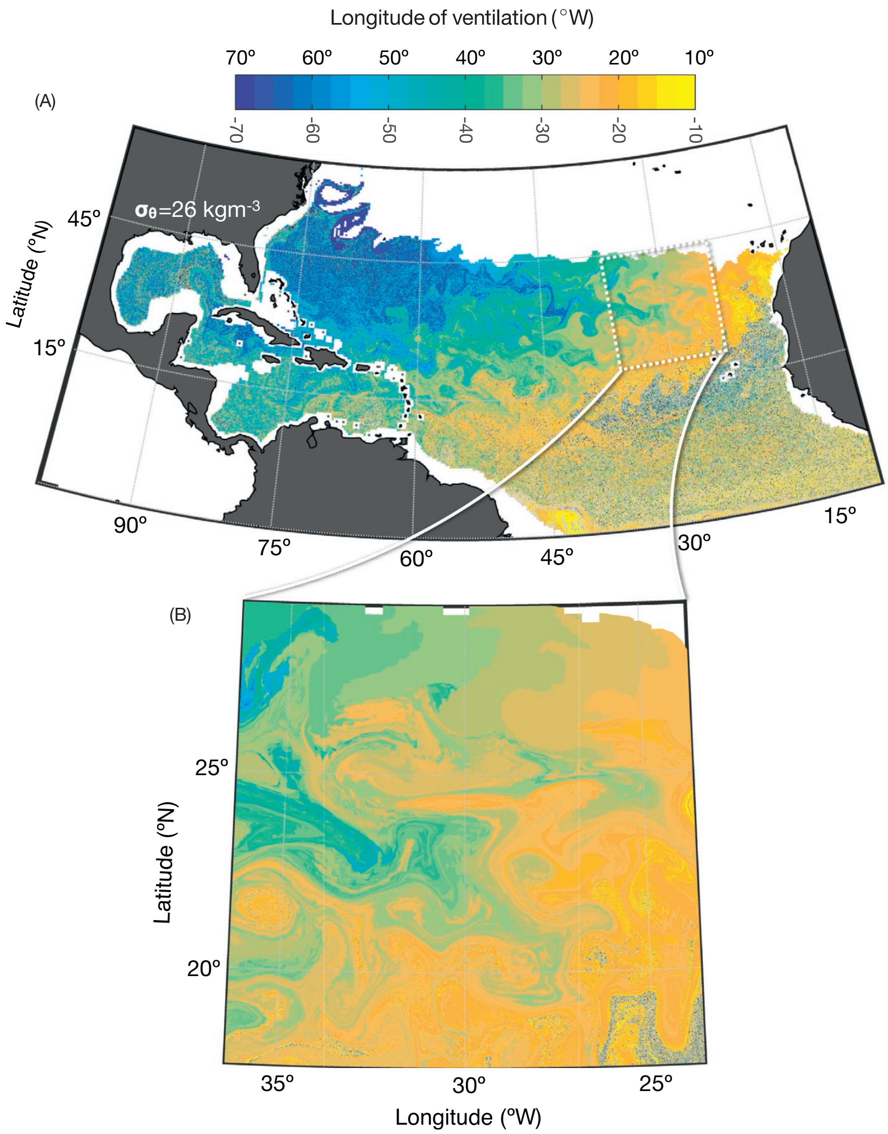

Fig. 10 A model assessment of particle trajectories labeled by their longitude when they are subducted at the end of March along potential density surface 26.0 kg m^{-3}. The analysis uses 5-day mean output from an eddy-permitting ORCA025 model at 1/4° horizontal resolution at the equator. In (A), the subducted particles show a signal consistent with the expected gyre-scale circulation across the basin, while in (B) a zoomed version over 10°x10°, there are stirring signals, highlighting the chaotic nature of the ventilation process when there is a representation of ocean eddies. Reproduced from MacGilchrist, G.A., Marshall, D.P., Johnson, H.L., Lique, C., and Thomas, M. (2017). Characterizing the chaotic nature of ocean ventilation. *Journal of Geophysical Research: Oceans*, **122**, 7577–7594, https://doi.org/10.1002/2017JC012875.

expected from nondiffusive, ideal thermocline models. Hence, the stirring of tracer contrasts by fine-scale circulations appears to be significant along potential density surfaces throughout a basin and is particularly important in regions of weak background flow.

The Relationship Between Subduction, Potential Vorticity and Buoyancy Forcing

Local, kinematic connection

The relationship between subduction rate, the mixed layer evolution and underlying potential vorticity is illustrated here. For a particle subducted from the mixed layer into the underlying thermocline (with a continuous match in potential density as depicted in Fig. 2), the subduction rate (1) and potential vorticity (3) are kinematically related:

$$Q = -\frac{f}{\rho S}\frac{D_b\sigma_m}{Dt}, \quad (4)$$

where $D_b\sigma_m/Dt$ is the Lagrangian change in mixed-layer potential density, σ_m, following the velocity $\boldsymbol{u}_b$ at the base of the mixed layer, which is usually taken to be a geostrophic streamline. Hence, Q in the thermocline is diagnostically related to the evolution of the mixed-layer density following the horizontal flow and the subduction rate—see Fig. 8 for an example of the resulting Q solution for an ideal thermocline model of a subtropical gyre (Williams, 1991).

Alternatively, Eq. (4) can be re-arranged to highlight how subduction only occurs when the mixed-layer density becomes lighter following a geostrophic streamline:

$$S = -\frac{f}{\rho Q}\frac{D_b\sigma_m}{Dt}, \tag{5}$$

where $S > 0$ occurs when $D_b\sigma_m/Dt < 0$. This relation reflects how subduction relies on fluid being capped by a lighter mixed layer and hence transferred into the thermocline (Fig. 2). For gyre-scale subduction, the required buoyancy input can be provided by an annual average of a surface buoyancy flux or a convergent Ekman flux of buoyancy; the latter contribution dominates in climatological diagnostics over the subtropical gyre of the North Atlantic (as implied from buoyancy diagnostics associated with Fig. 5).

However, these kinematic relations (4) and (5) cease to be useful in a time-averaged limit in the presence of an active eddy circulation due to the difficulty in defining an appropriate streamline. The Lagrangian change in mixed layer density, $D_b\sigma_m/Dt$, might even vanish following the time-mean geostrophic streamline and provide a misleading result from Eq. (5). Instead, the role of the time-varying circulation in subduction and an integrated view of the buoyancy forcing need to be considered.

Integral connection

An integral view of the ventilation process may be obtained by considering the conversion of water masses within a basin or restricted domain (Fig. 11) (Walin, 1982). Buoyancy fluxes at the sea surface and diffusive fluxes within the ocean leads to water masses being transformed from one density class to another. The convergence of this diapycnal volume flux, or water-mass transformation, defines the rate of water-mass formation within a particular density class. In the limit of no diffusive fluxes within the ocean, the rate of water-mass formation rate is equivalent to the area-integrated subduction rate over the particular density class into the main thermocline (Fig. 11) (Marshall et al, 1999); this connection is exploited for the later example of a convective chimney in Fig. 14.

The contribution of surface buoyancy fluxes to the rate of water-mass formation has been estimated from climatological air–sea fluxes. While there are significant errors in the air–sea fluxes, the diagnostics reveal how the buoyancy forcing and subduction process leads to the preferential formation of distinct modes waters in the North Atlantic (Fig. 12), rather than creating similar rates of formation for all densities (Speer and Tziperman, 1992).

Role of Eddies and Fronts

Mesoscale eddies can directly assist in the subduction process through modifying the subduction rate and diffusing the downstream properties of subducted fluid. Eddies lead to an effective stirring of water-mass properties along isopycnals (Fig. 10). This process can lead to subducted mode waters becoming rapidly smeared out, such that the downstream water-mass properties reflect an average of the upstream characteristics determined in the mixed layer at the end of winter.

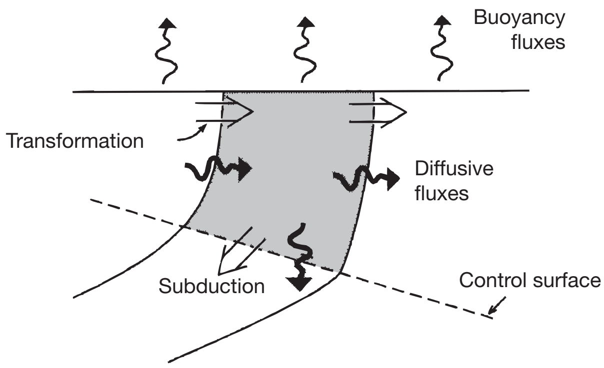

Fig. 11 A schematic diagram of the upper ocean showing the sea surface, two outcropping isopycnals and a fixed control surface across which subduction is monitored. Buoyancy fluxes at the sea surface and diffusive fluxes in the interior transform water masses from one density class to another. If there are no diffusive fluxes in the interior, the convergence of the transformation flux controls the subduction flux across the control surface (which is chosen to be the interface between the base of the end of winter mixed layer and the main thermocline). Reproduced from Marshall, J., Jamous, D., and Nilsson, J. (1999). Reconciling thermodynamic and dynamic methods of computation of water-mass transformation rates. *Deep-Sea Research I* **46**, 545–572. © 1999, with permission from Elsevier Science.

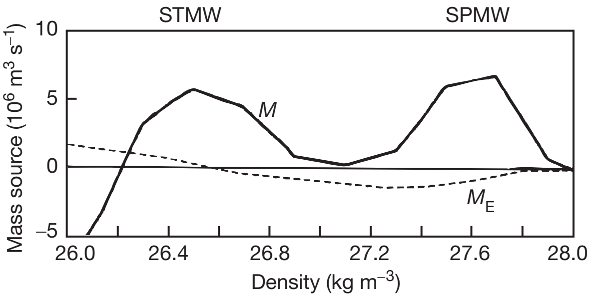

Fig. 12 Diagnosed water-mass formation (10^6 m^3 s^{-1}) versus density (kg m^{-3}) evaluated from climatological surface buoyancy fluxes (*full line*) and from Ekman pumping (*dashed line*) over the North Atlantic. The left-hand peak is centered near the density of subtropical mode water (STMW) and the right-hand peak spans subpolar mode water (SPMW) and Labrador Sea Water densities. Reproduced with permission from Speer, K. and Tziperman, E. (1992). Rates of water mass formation in the North Atlantic. *Journal of Physical Oceanography* **22**, 93–104.

Eddy-Induced Transport and Subduction

Eddies provide a volume flux and transport of tracers, which can modify the subduction rate. The eddy-driven transport, $\overline{\mathbf{u}'\Delta z'}$, is given by the temporal correlation in velocity and thickness of an isopycnal layer, Δz (Gent et al., 1995); the overbar represents a time average over several eddy lifetimes and the prime represents a deviation from the time average. An example of an eddy-driven transport is shown in Fig. 13A where the oscillating velocity leads to an eddy-driven transport through the volume flux in the thick "blobs" of fluid being greater than the return flux in the thin "blobs" of fluid (Lee et al., 1997). An eddy-induced transport velocity or "bolus" velocity within an isopycnic layer is defined by $\mathbf{u}^* = \overline{\mathbf{u}'\Delta z'}/\overline{\Delta z}$. The eddy "bolus" velocity is particularly large in regions of baroclinic instability, where the transport is generated through the flattening of isopycnals (Fig. 13B).

Evaluating the instantaneous subduction rate (1) in the presence of eddies is difficult, since time-averaged correlations in velocity with mixed-layer thickness and velocity with mixed-layer outcrop area are required. However, the subduction rate into the main thermocline and across the control surface, H, can be evaluated, in principle, by including the eddy-induced "bolus" velocity in Eq. (2) (Marshall, 1997) to give,

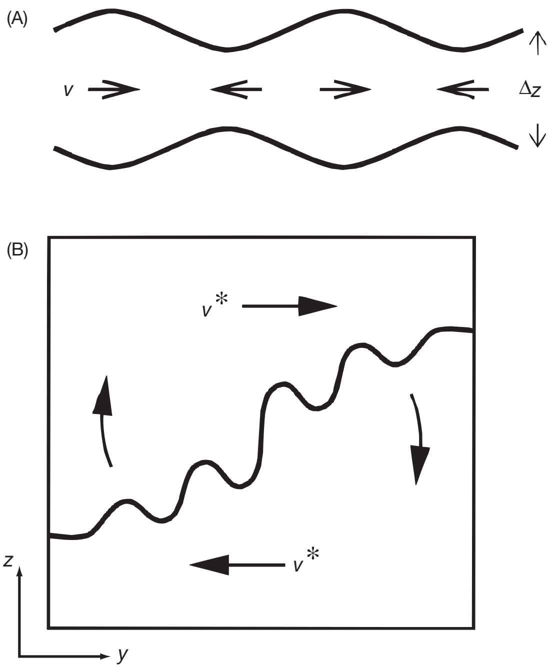

Fig. 13 A schematic diagram illustrating the concept of the eddy rectified transport and "bolus" velocity: (A) The time-averaged meridional volume flux in an undulating layer, $\overline{v\Delta z} = \bar{v}\overline{\Delta z} + \overline{v'\Delta z'}$, depends on advection by the time-mean meridional velocity $\bar{v}$ and the time-averaged correlations in the temporal deviations in velocity v' and thickness of an isopycnal layer $\Delta z'$. The time-averaged volume flux is directed to the right as drawn here, since $v > 0$ correlates with large layer thickness, while $v < 0$ correlates with small layer thickness. (B) The slumping of isopycnals in baroclinic instability induces a secondary circulation in which the bolus velocity, $v^* = \overline{v'\Delta z'}/\overline{\Delta z}$, is polewards in the upper layer and equatorwards in the lower layer. Eddy-driven subduction is likely to occur in baroclinic zones where there are strongly inclined isopycnals. Reproduced with permission from Lee, M.-M., Marshall, D.P., and Williams, R.G. (1997). On the eddy transfer of tracers: advective or diffusive? *Journal of Marine Research* **55**, 483–505.

$$S_{ann} = -(\bar{w}_H + w^*) - (\bar{\boldsymbol{u}}_H + \boldsymbol{u}^*).\nabla H, \tag{6}$$

where the vertical eddy-transport contribution, w^*, is obtained from continuity. Hence, the annual subduction rate includes contributions from the time-mean and time-varying circulation. The eddy contribution to subduction is expected to be large wherever the bolus velocity is significant, such as where currents have a large vertical shear and strongly sloping isopycnals.

Over a subtropical gyre, the subduction contribution from the time-mean circulation probably dominates over the eddy "bolus" contribution, since the wind forcing drives an anticyclonic circulation across contours of mixed-layer thickness. However, the eddy contribution to subduction becomes important throughout the Southern Ocean, as well as across separated boundary currents and inter-gyre boundaries.

Subduction Associated With Fronts and Convective Chimneys

Subduction and ventilation occurs over fine scales associated with narrow fronts and convective chimneys, as well as over the larger-scale circulation. Secondary circulations develop across a front following the instability of the front and acceleration of the along stream flow. Frontal modeling studies demonstrate how this secondary circulation leads to frontal-scale subduction, which injects low stratification into the thermocline. Observational studies have identified upwelling and downwelling zones on either side of a narrow front (10 km wide) with vertical velocities reaching 40 m day^{-1}, which is much greater than the background, wind-driven, downwelling velocity. Indirect support for frontal-scale subduction is also provided by separate observations of bands of short-lived chlorophyll penetrating downward for several 100 m into the stratified thermocline on the horizontal scale of 10 km; this frontal-scale process is also identified as being important in the atmosphere in transferring ozone-rich, stratospheric air into the troposphere.

An example of fine-scale subduction associated with a convective chimney is shown in Fig. 14 (Marshall, 1997). Buoyancy loss to the atmosphere drives the conversion of light to dense water within the mixed layer and the concomitant thickening of the mixed layer. Baroclinic instability of the chimney leads to an eddy-driven transport, which fluxes light fluid into the chimney at the surface and, in turn, subducts denser fluid into the thermocline. Hence, the eddy-driven transport enables recently ventilated dense waters to disperse away from a convective chimney. The eddy-driven influx of light fluid partly offsets the buoyancy loss to the atmosphere and can inhibit further mixed-layer deepening over the center of the chimney.

Subduction in the Southern Ocean

The Southern Ocean and the previously discussed North Atlantic are the dominant locations in the global ocean for the subduction of water masses from the mixed layer into the thermocline. Unlike the North Atlantic, the Southern Ocean is also the region of most upwelling in the global ocean. The prevailing westerly winds and surface buoyancy forcing act to drive the Antarctic Circumpolar

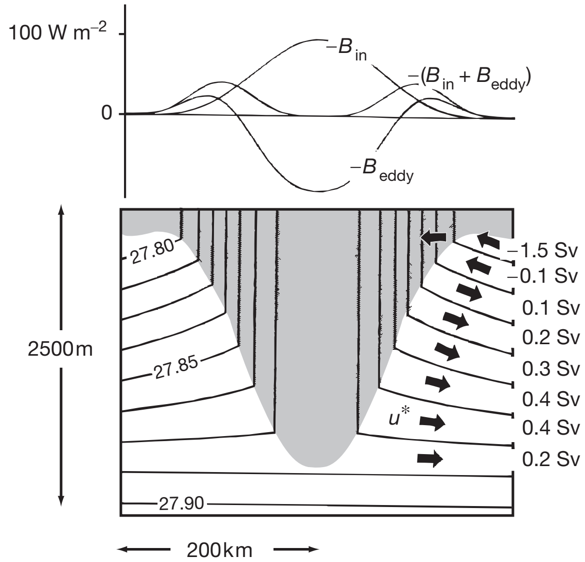

Fig. 14 A model solution for the eddy-driven subduction rate and volume fluxes into and out of a convective chimney (shaded) with isopycnals depicted by *solid lines*. Baroclinic instability of the chimney leads to an eddy-driven circulation. There is entrainment of warm, buoyant fluid over the upper kilometer with subduction of cold, dense fluid at greater depths—the net subduction of deep water is 1.6 Sv (1 Sv = 10^6 m^3 s^{-1}). The subduction rate is determined by the buoyancy forcing, $B_{in} + B_{eddy}$, where B_{in} is the surface buoyancy loss to the atmosphere and B_{eddy} is the eddy flux of buoyancy within the mixed layer. Reproduced with permission from Marshall, D.P. (1997). Subduction of water masses in an eddying ocean. *Journal of Marine Research*, **55**, 201–222.

Current (ACC), a strong current encircling Antarctica. This current is associated with steep north-south sloping isopcynals that intersect the mixed layer and drive regional patterns of positive and negative subduction. There is an important exchange of heat and gasses in the region arising from old waters being upwelled and exposed to the atmosphere south of the ACC and then being subducted to the north. The heat and freshwater exchanges are also associated with a vigorous meridional overturning circulation (Marshall and Speer, 2012), where upwelled waters are converted in different regions either to denser bottom waters (Antarctic Bottom Water) or to lighter intermediate and mode waters (Antarctic Intermediate water and Sub-Antarctic Mode water).

The annual-mean transport across the base of the mixed layer in the Southern Ocean may be decomposed into several main processes (Fig. 15A–D) (Sallée et al., 2010). Wind-driven vertical transports of up to ± 50 m year^{-1} are coherently upward poleward of the ACC (blue in Fig. 15A) and downward to its north (red in Fig. 15A), and largely set the net subduction away from the ACC. In the vicinity of the ACC, strong local gradients in mixed-layer depth and the meandering of the flow streamlines combine to create vertical exchanges of over ± 200 m year^{-1} with significant regional variability. The mixed-layer deepens dramatically south of the subtropical gyres and then more gradually shoals poleward of the ACC. The single greatest contributor to local subduction rates is the lateral subduction $-\bar{\boldsymbol{u}}_H.\nabla H$ (Fig. 15B), where powerful surface currents encounter changes in mixed-layer depth, particularly on the northern flank of the ACC where winter mixed-layer depths have a varying spatial structure and can exceed 600 m. This lateral transfer of fluid passes from the mixed layer into the thermocline and then back again leads to reversing patterns of positive and negative subduction rate along the ACC. Net subduction into the thermocline occurs preferentially over the eastern side of the subtropical basins, where there is a lightening of the mixed layer and branches of mean northward flow separating from the northern ACC. There is a general poleward eddy bolus transport, $\boldsymbol{u}^*$, implied from the strong gradients in vertical spacing between

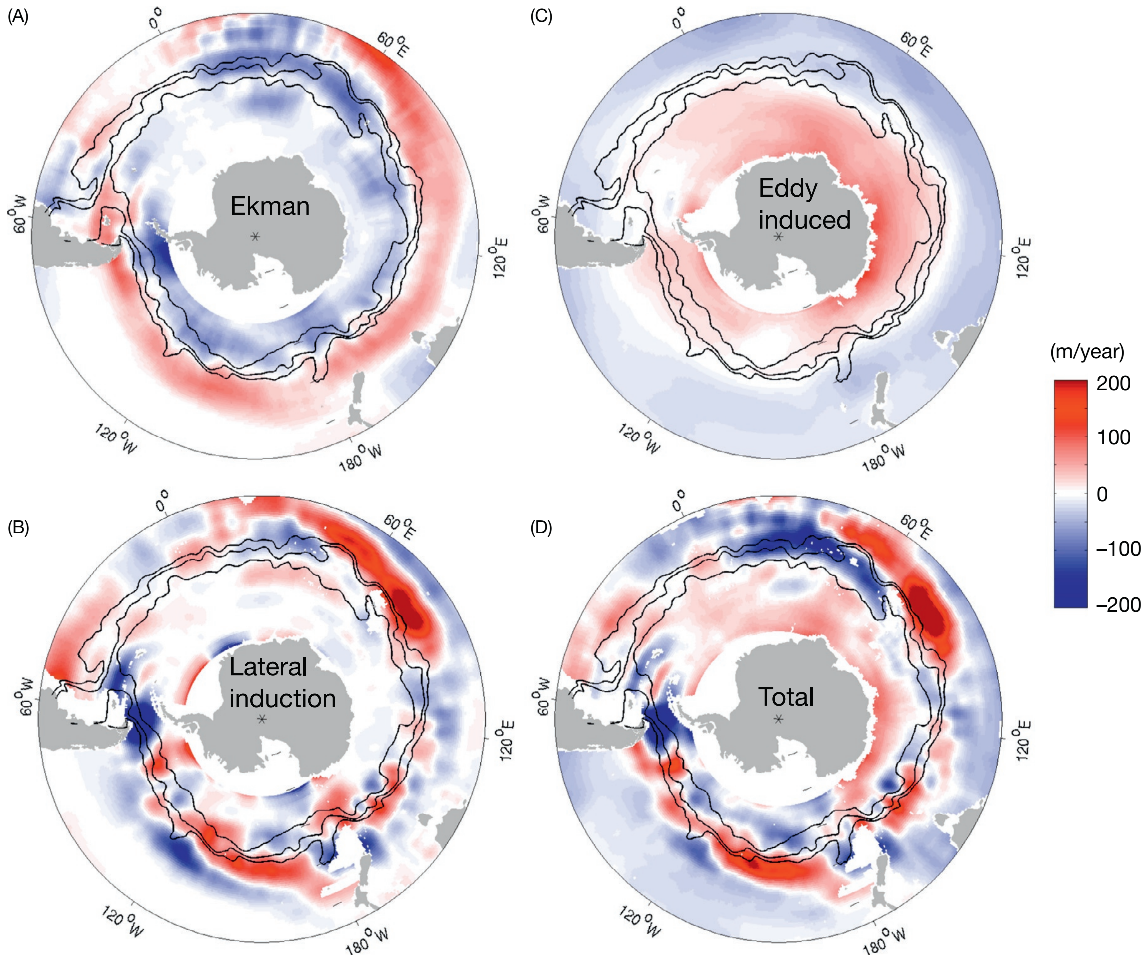

Fig. 15 Annual-mean subduction rate (m year^{-1}) in the Southern Ocean based on observationally-based, kinematic diagnostics. Following Eq. (6), the subduction rate is split into its (A) wind-driven (Ekman) component, (B) geostrophic lateral induction, (C) eddy-driven lateral induction and (D) the sum of all components. Positive values (*red*) denote subduction from the mixed layer into the thermocline. Net upwelling occurs poleward of the ACC, while net subduction occurs on its northern side, particularly where the ACC is deflected poleward, such as upstream of Drake Passage, south east of New Zealand and in the south east Indian Ocean. Reproduced from Sallée, J.B., Speer, K., Rintoul, S., and Wijffels, S. (2010). Southern Ocean thermocline ventilation. *Journal of Physical Oceanography*, **40**(3), 509–529.

density surfaces, which drives eddy subduction from $-\boldsymbol{u}^{*}.\ \nabla H$ with positive values on the poleward edge of the ACC and negative values to the north (Fig. 15C). The general zonal symmetry of the eddy subduction is broken north of the ACC by the presence of the Agulhas retroflection and Brazil currents, which modify the local mixed-layer slope and drive downwelling.

Subduction on the northern edge of the ACC in late winter typically forms Sub-Antarctic Mode Water (SAMW), characterized by low stratification and a minimum in the large-scale potential vorticity, Q. This mode water is transported into the interior of the subtropical gyres following open contours of Q. The spreading of this mode water is illustrated by passive tracers released into the thermocline at SAMW formation sites on the northern side of the ACC and spreading northward (Fig. 16) (Jones et al., 2016).

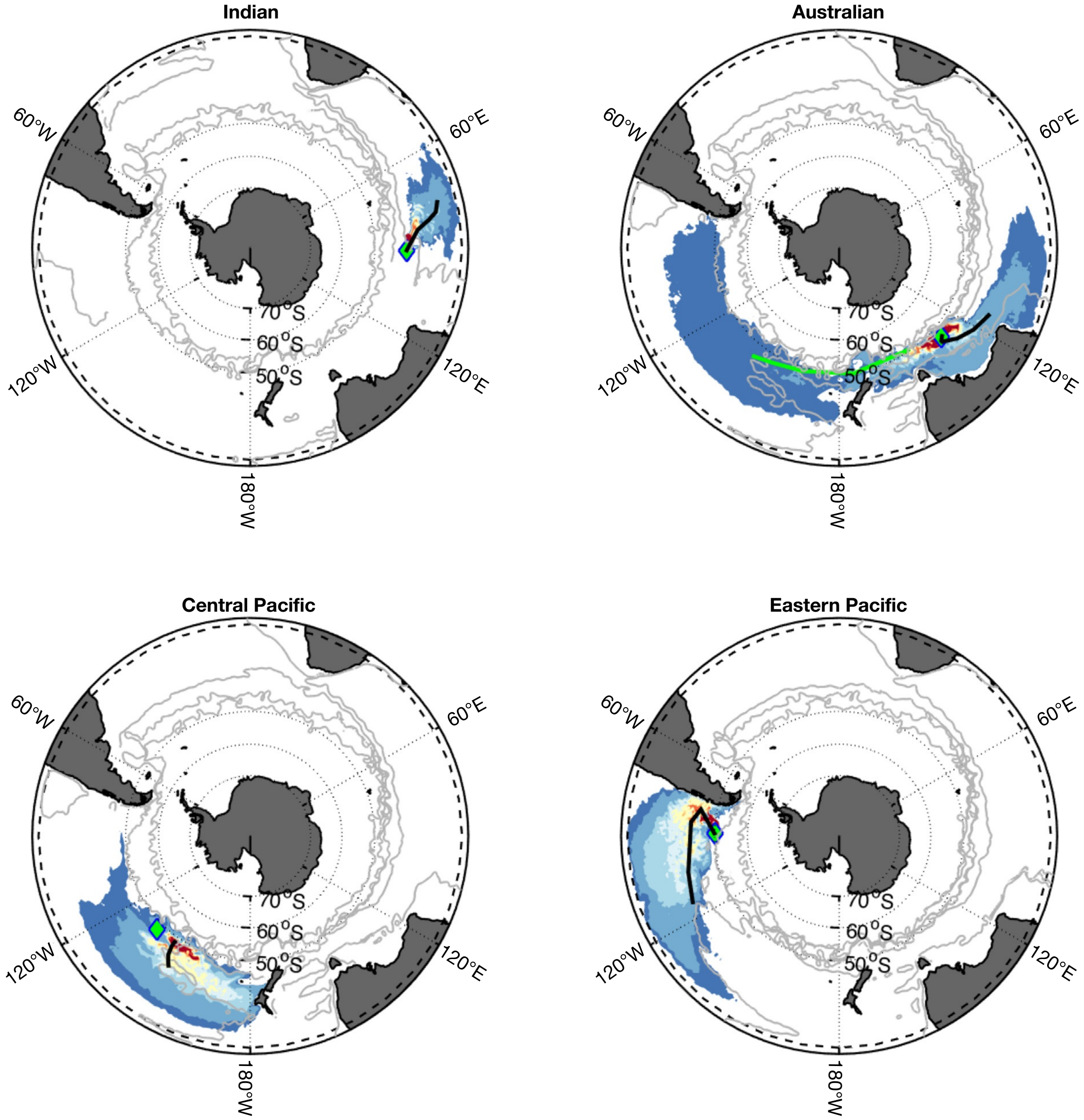

Fig. 16 An eddy-resolving model tracer release experiment, showing the 50-year histogram of the vertical integral of a passive tracer released at the base of the mixed layer at several Sub-Antarctic Mode Water formation sites (diamonds) in the Southern Ocean. The tracer clearly travels from these sites into the thermocline and spreads over the subtropical gyres. *Fine lines* indicate the approximate local geostrophic transport streamlines and the tracer follows these into the gyre, spreading laterally via eddy stirring. *Thick lines* indicate the position of the center of mass of the tracer. In each basin, the core of the tracer tends to travel equatorward, particularly at the eastern edge of the subtropical gyres before spreading predominantly westward through the action of Rossby waves and mesoscale stirring. Significant eastward transport is largely achieved along the northern flank of the ACC (e.g., *green line* in the Australian release) which then spreads more gradually toward the equator. Note that concentrations are scaled by maximum concentration in each release region. Reproduced from Jones, D.C., Meijers, A.J., Shuckburgh, E., Sallée, J.B., Haynes, P., McAufield, E.K., and Mazloff, M.R. (2016). How does Subantarctic Mode Water ventilate the Southern Hemisphere subtropics?. *Journal of Geophysical Research: Oceans*, **121**(9), 6558–6582.

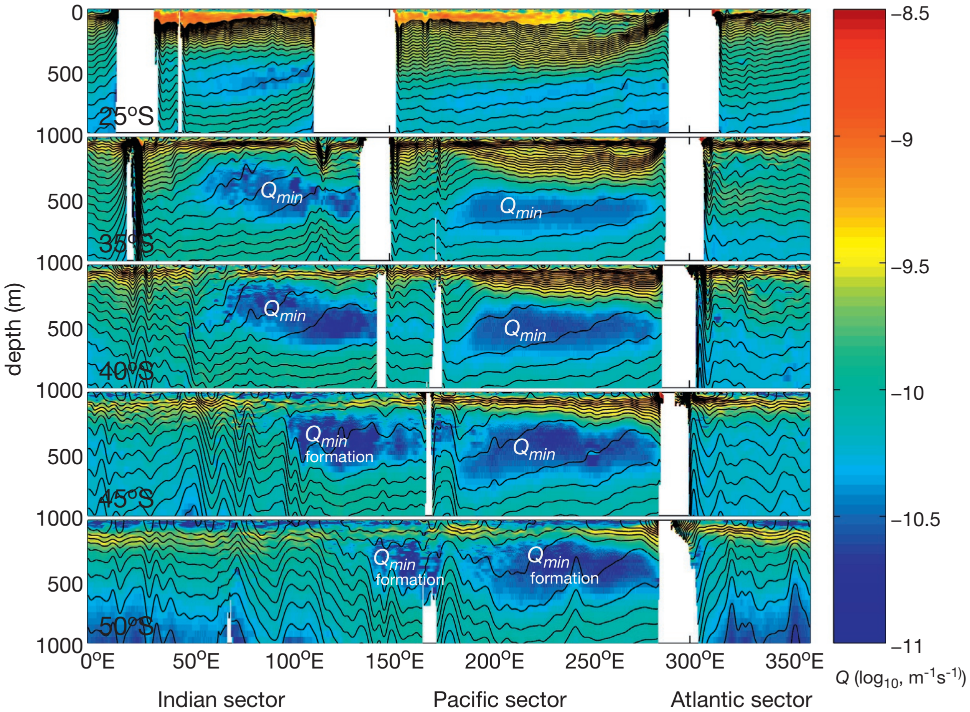

Fig. 17 Annual-mean zonal sections of large-scale potential vorticity Q along latitudes 25°S to 50°S of the Southern Ocean, taken from the CARS09 climatology ($\log_{10}$ scale, $m^{-1}\ s^{-1}$). Contours indicate lines of constant neutral density. Sub-Antarctic Mode Water (SAMW) formation is identified as regions of low vertical stratification with local minima in Q (marked by label of Q_{min}). The low Q appear most distinctly in the southeast Pacific and Indian oceans (near 50°S and 45°S respectively) and these signals spread westward and deepen northward along isopycnals following the gyre circulation. Following this northward spreading of the SAMW, there is a gradual increase in its Q and a reduction in its minimum due to the action of eddy stirring and diapycnal mixing. Replicated from Herraiz-Borreguero, L. and Rintoul, S.R. (2011). Subantarctic mode water: distribution and circulation. *Ocean Dynamics*, **61**(1), 103–126.

The tracers are advected to the north and spread laterally following the broadly anticyclonic gyre circulation. Eddy stirring leads to the downstream broadening of the tracer plume within the northern basins. The SAMW formation sites north of the ACC appear as a distinct minima in Q (around 45–50°S) in the south-east Pacific and south-east Indian oceans (Fig. 17) (Herraiz-Borreguero and Rintoul, 2011); these minima in Q progressively weaken as the mode water spread equatorward and anticyclonically within the subtropical gyres (Fig. 17) due to dilution from the combined effects of stirring and diapycnal mixing with surrounding water masses.

Conclusions

The subduction process controls the rate at which the upper thermocline is ventilated, as well as determining the water-mass structure and stratification of the upper ocean. Subduction leads to an asymmetrical coupling between the mixed layer and ocean interior: fluid is transferred into the permanent thermocline in late winter and early spring, rather than throughout the year. This transfer of fluid into the permanent thermocline helps to determine the relatively long memory of the ocean interior, compared with that of the surface mixed layer. Conversely, the reverse of the subduction process is also important. The transfer of thermocline fluid into the seasonal boundary layer affects the downstream water-mass properties of the mixed layer, and hence alters the air–sea interaction and biogeochemistry. For example, biological production is enhanced wherever nutrients are vertically and laterally fluxed from the thermocline into the mixed layer and the euphotic zone.

Subduction can occur over a range of scales extending over fronts, mesoscale eddies and gyres, as is clearly demonstrated for the North Atlantic and the Southern Ocean. Subduction determines how surface waters in contact with the atmosphere are communicated to the ocean interior, as well as how older waters are returned to the mixed layer. Consequently, the subduction process helps control the ocean uptake of additional heat and carbon supplied to the atmosphere from climate change.

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Further Reading

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