



DYNAMIQUE DE LA CIRCULATION ESTUARIENNE DOUBLE ARCTIQUE

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À toutes les femmes de ma lignée qui n'ont
pas eu la chance d'explorer leur curiosité

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RÉSUMÉ

La circulation océanique arctique joue un rôle important dans le réchauffement climatique aux hautes latitudes par le phénomène de rétroaction glace-albédo. Celui-ci est amplifié par l'augmentation des flux de chaleur atmosphériques et océaniques dans le bassin Arctique, eux-mêmes influencés par la circulation océanique arctique. Il est donc essentiel de mieux comprendre la dynamique interne de la circulation de renversement afin d'évaluer son influence sur la région.

Selon la littérature, la circulation de renversement arctique serait formée de deux cellules. Une première, estuarienne, située dans la portion supérieure de la colonne d'eau et une deuxième, convective, se trouvant plus en profondeur. Cependant, les méthodes utilisées jusqu'à maintenant pour diagnostiquer la circulation de renversement met surtout en évidence la cellule convective et peu d'informations relatives à la cellule estuarienne peuvent en être extraites. Il est montré qu'avec l'utilisation de parcours reliant le détroit de Béring au détroit de Fram, les deux cellules peuvent être diagnostiquées simultanément.

Pour évaluer la réponse de cette circulation dans le régime transitoire du réchauffement climatique, le modèle "Community Earth System Model" (CESM) et le scénario de réchauffement "Representative Concentration Pathway 8.5" (RCP8.5) ont été utilisés. Il est montré que la cellule convective est forcée par une combinaison des flux de flottabilité à la surface du bassin arctique et de l'influence de la stratification nord atlantique. L'intensité de la cellule estuarienne quant à elle est fortement corrélée avec l'intensité de la cellule convective, mais ne présente pas de corrélation claire avec l'apport en eau douce dans le domaine. Avec le réchauffement climatique, des signes d'atlantification ont pu être diagnostiqués. La cellule convective devient plus intense, se concentre plus proche de la surface et persiste tout au long de l'année, ce qui restreint le domaine vertical de la cellule estuarienne.

À l'aide d'un modèle idéalisé de l'Arctique, les mécanismes de forçage de ces cellules ont été explorés. La formation et la fonte de glace de mer constituent le principal mécanisme de forçage de la cellule convective et influence grandement la cellule estuarienne. Le cycle saisonnier de la glace génère un cycle saisonnier de la circulation de renversement. En hiver, la formation de glace crée des eaux saumâtres en surface. Si celles-ci sont suffisamment denses, elles peuvent traverser l'halocline, aller jusqu'au fond du bassin et générer la cellule convective. En été, la fonte de glace et l'apport fluvial restaurent la stratification, ce qui renforce la cellule estuarienne. Si les flux de sel demeurent trop importants, la stratification ne sera pas restaurée et la circulation devient alors purement convective. Ainsi, selon l'intensité des flux de sel, trois régimes dynamiques existent : estuarien, double estuarien ou convectif. D'autres mécanismes de forçage, tels que la stratification dans les mers nordiques, le transport au détroit de Béring et l'intensité de l'anticyclone de Beaufort, renforcent la stratification, ce qui favorise la cellule estuarienne. Dans ce cas, pour maintenir l'intensité de la

cellule convective, le flux de sel à la surface doit augmenter de façon à permettre aux isopycnales de remonter jusqu'à la surface et maintenir le volume de création des eaux profondes. Finalement, une augmentation de la diffusivité renforce la cellule estuarienne puisqu'elle est forcée par le mélange. Finalement, grâce au modèle idéalisé les flux de chaleur océaniques et leur influence sur la température moyenne du bassin arctique ont pu être étudiés. Il est montré que le transport de chaleur au détroit de Fram est fortement corrélé à l'intensité de la cellule convective. Cette tendance se reflète dans la température moyenne au-dessus de l'halocline. Pour des circulations de type estuarien, la température moyenne de l'ensemble du bassin est plus chaude en raison de l'import d'eaux atlantiques chaude en profondeur. Dans le cas de la circulation estuarienne double, la cellule convective apporte des eaux refroidies en surface vers le fond du bassin diminue la quantité de chaleur moyenne. Dans le cas d'une circulation estuarienne double dominée par la convection, la convection des eaux de surface froides vers le fond est contrebalancé par le transport de chaleur à Fram qui devient éventuellement dominant et augmente la température moyenne.

Mots clés : Arctique, Circulation de renversement, Dynamique, Forçage, Flux de chaleur, Atlantification

ABSTRACT

The arctic ocean's circulation plays a significant role in warming at high latitudes due to its influence on atmospheric and oceanic heat import into the arctic basin which influences the albedo feedback mechanism. Therefore, to better understand the influence of the arctic ocean on the local climate it is important investigate the internal dynamics of the oceanic circulation.

According to the literature, the arctic's overturning is composed of two cells. An estuarine cell is located in the upper portion of the water column sitting above a convective cell. The traditional method of diagnosis of the overturning along meridians favors the convective cell and provides little information about the estuarine cell. We show that both cells can be simultaneously diagnosed by integrating the circulation along an axe going from Bering Strait to Fram Strait.

To assess the response of the overturning circulation under a transient regime such as global warming the Community Earth System Model (CESM) and the Representative Concentration Pathway 8.5 (RCP8.5) forcing scenario were used. It is shown that the convective cell is forced by a combination of surface buoyancy fluxes and influenced by North Atlantic stratification. The intensity of the estuarine cell is strongly correlated with the intensity of the convective cell. No clear correlation between the estuarine cell and freshwater input was found as it was established with the idealized model. Under climate warming, signs of Atlantification were detected. The convective cell strenghtens, shallows, and persists year-round, thereby constraining the estuarine cell closer to the surface.

With this diagnostic method in hand and using an idealized Arctic model, the forcing mechanisms of each of the two cells were explored. The cycle of sea ice formation and melt was found to be the main forcing mechanisms of the convective cell and to strongly influence the estuarine cell. Moreover, the seasonal cycle of sea ice is reflected into the overturning circulation. In winter, sea ice formation creates brine rejection and dense surface waters. If those are dense enough, they can break the halocline, sink to the bottom of the basin and create the convective cell. In summer, ice melt and river runoff provide freshwater at the surface and restratify the water column thereby strengthening the estuarine cell. For large salt fluxes, the stratification is no longer restored in summer and the circulation becomes purely convective. Therefore, depending on the intensity of the surface buoyancy fluxes, the circulation can be estuarine, double estuarine or convective. The Nordic Seas stratification, transport at Bering Strait, and the strength of the Beaufort High, all previously suggested forcing mechanisms of the circulation, reinforce stratification in the Arctic basin, favoring the estuarine cell. For the convective cell to maintain the same intensity, surface salt flux must increase to compensate for the stronger halocline. Finally, increasing the diffusivity strengthens the estuarine cell, which is known to be forced by mixing, and does not influence de convective cell. Finally,

using the idealized model, the oceanic heat fluxes and their influence on the mean temperature of the Arctic Basin were examined. The heat transport through Fram Strait is strongly correlated with the intensity of the convective cell. This relationship is reflected in the mean temperature above the halocline. However, the basin-wide mean temperature is warmer for purely estuarine circulations due to the import of Atlantic waters at depth. For double estuarine circulations, the convection of cooled surfaced waters to the bottom of the basin reduces the mean temperature until this effect is outweighed by heat transport through Fram Strait as the convective cell gains in intensity, and eventually becomes the dominant factor for heat content.

Keywords : Arctic, Overturning circulation, Dynamics, Forcing, Heat flux, Atlantification

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AMOC Atlantic Meridional Overturning Circulation

(Circulation de renversement de l'Atlantique)

CESM-LE Community Earth System Model - large ensemble

CMIP Coupled Model Intercomparison Project

GIN Seas Greenland, Iceland and Norwegian Seas

(Mers du Groenland, d'Islande et de Norvège)

IPCC Intergovernmental Panel on Climate Change

(Le Groupe d'experts intergouvernemental sur l'évolution du climat)

MITgcm Massachusetts Institute of Technology General Circulation Model

MOC Meridional Overturning Circulation (Circulation méridionale de renversement)

PREC_F Diagnostic de précipitation du CESM-LE

RCP8.5 Representative Concentration Pathway 8.5

SFWF Surface freshwater flux (Flux d'eau douce à la surface)

SHF Surface heat flux (Flux de chaleur à la surface)

IOFF_F Diagnostic de ruissellement de la glace continentale du CESM-LE

LISTE DES SYMBOLES

- A** Amplitude du flux de sel à la surface [m/année]
- $\Delta\langle\Theta\rangle_A$ Anomalie de la température moyenne du bassin arctique [$^{\circ}\text{C}$]
- $\Delta\langle\Theta\rangle_{A200m}$ Anomalie de la température moyenne du bassin arctique de 0 à 200 m [$^{\circ}\text{C}$]
- ΔQ_{Bering} Anomalie du flux de chaleur au détroit de Béring [$^{\circ}\text{C/s}$]
- ΔQ_{Fram} Anomalie du flux de chaleur au détroit de Fram [$^{\circ}\text{C/s}$]
- ΔQ_{Surf} Anomalie du flux de chaleur à la surface du bassin [$^{\circ}\text{C/s}$]
- L** Transport de la branche inférieure de la circulation de renversement arctique [Sv]
- M** Transport de la branche intermédiaire de la circulation de renversement arctique [Sv]
- Ψ Fonction de courant [Sv]
- Ψ_{regime} Indice du type de circulation arctique
- Q** Flux de chaleur moyen [$^{\circ}\text{C/s}$]
- Q_{Bering} Flux de chaleur au détroit de Béring [$^{\circ}\text{C/s}$]
- Q_{conv} Flux de convection limite [$^{\circ}\text{C/s}$]
- Q_{Fram} Flux de chaleur au détroit de Fram [$^{\circ}\text{C/s}$]
- Q_{Surf} Flux de chaleur à la surface du bassin [$^{\circ}\text{C/s}$]
- $\langle\Theta\rangle_A$ Température moyenne du bassin arctique [$^{\circ}\text{C}$]
- U** Transport de la branche supérieure de la circulation de renversement arctique [Sv]
- v** Vitesse méridionale [m/s]
- w** Vitesse verticale [m/s]

INTRODUCTION GÉNÉRALE

Une conséquence frappante du réchauffement climatique est que l'augmentation de la température de surface est environ deux fois plus importante en Arctique que sur le reste du globe. Cette tendance est constatée autant par les observations des dernières décennies (Steele et al., 2008) que par les projections des modèles climatiques globaux pour les prochaines décennies (Manabe and Stouffer, 1980; Serreze and Barry, 2011; Marshall et al., 2015). Pour expliquer ce phénomène que l'on nomme amplification Arctique, plusieurs mécanismes ont été suggérés tels que la diminution de la couverture de glace de mer (Hansen et al., 1984; Serreze et al., 2007; Screen and Simmonds, 2010), l'augmentation des transports de chaleur atmosphériques et océaniques vers l'Arctique (Graversen et al., 2008; Chylek et al., 2009; Yang et al., 2010), la modification de la couverture nuageuse (Hansen et al., 1984; Francis and Hunter, 2006), l'interaction entre la vapeur d'eau et la radiation solaire dans l'atmosphère (Holland and Bitz, 2003; Graversen and Wang, 2009; Abbot et al., 2009), le dépôt de suie sur la neige (Hansen and Nazarenko, 2004) et l'augmentation de la concentration de carbone sous forme d'aérosol dans l'atmosphère (Shindell and Faluvegi, 2009). Ces mécanismes sont répertoriés et expliqués plus en détail dans la revue de littérature de Serreze and Barry (2011).

Un des mécanismes ayant la plus grande importance sur le réchauffement de l'Arctique est la diminution de la couverture de glace de mer étant donné le mécanisme de rétroaction glace-albédo (Holland and Bitz, 2003; Serreze and Barry, 2011). La disparition graduelle de la banquise pluriannuelle implique qu'une fraction de plus en plus importante du rayonnement solaire est absorbée dans l'océan au lieu d'être réfléchi par la glace. La diminution de la glace pluriannuelle est influencée par plusieurs mécanismes, dont l'augmentation des flux de chaleur atmosphériques (Mahlstein and Knutti, 2011) et particulièrement par l'augmentation des flux de chaleur océaniques vers l'Arctique (Marshall et al., 2014; Sirevaag et al., 2011; Steele et al., 2010).

Effectivement, les caractéristiques de la colonne d'eau de l'océan Arctique permettent

la création de glace pluriannuelle. La Figure 1 illustre la topographie de l'Arctique ainsi que les profils en profondeur de la température et de la salinité. Dans cette région, puisque la température des eaux de surface se situe autour du point de congélation, la densité n'est pas régie par la température, mais plutôt par la salinité. Ainsi, la colonne d'eau est caractérisée par une couche d'eaux froides et douces à la surface provenant du Pacifique, des rivières et de la formation de la glace. Sous cette couche se trouvent les eaux provenant de l'Atlantique qui sont plus chaudes et plus salées. La couche d'eau froide à la surface permet d'isoler la glace formée dans cette région de l'eau chaude provenant de l'Atlantique et de maintenir celle-ci sur plusieurs années. À la Figure 1 b), on note un fort gradient de salinité entre 50 et 200 mètres que l'on nomme halocline et qui permet d'isoler les eaux plus chaudes et denses de la surface. La structure de l'halocline est donc importante pour décrire l'évolution de la banquise pluriannuelle (e.g., Aagaard et al. (1981) et Polyakov et al. (2018)). Si l'épaisseur de l'halocline diminue, il en résulte une plus forte exposition de la banquise à la chaleur des eaux atlantiques se trouvant sous l'halocline et la fonte de la banquise est favorisée. À l'opposé, si la production d'eaux peu denses en surface augmente, l'halocline devient plus profonde et augmente l'isolation de la banquise à la chaleur contenue dans les eaux profondes.

La majorité du flux de chaleur océanique entre en Arctique par la mer de Barents et par le détroit de Fram (Arthun et al., 2019; Docquier and Koenigk, 2021; Oldenburg et al., 2024). Contre-intuitivement, la circulation de renversement de l'Atlantique Nord diminue avec le réchauffement climatique, alors même que l'apport de chaleur vers l'Arctique augmente (Bitz et al., 2006; Oldenburg et al., 2018). Cela suggère que la dynamique interne de l'océan Arctique joue un rôle déterminant dans son bilan de chaleur. Pour mieux comprendre cette dynamique interne, nous pouvons nous intéresser à la circulation de renversement arctique qui décrit la circulation de la surface jusqu'au fond projetée le long d'un axe allant du Sud vers le Nord. Il est également nécessaire d'approfondir notre compréhension de l'influence de cette circulation sur le bilan de chaleur du bassin et ce tant pour des régimes à l'équilibre que pour des régimes transitoires tel que le réchauffement climatique. Afin de répondre à cette question, il convient tout d'abord d'examiner les caractéristiques de la circulation de renversement

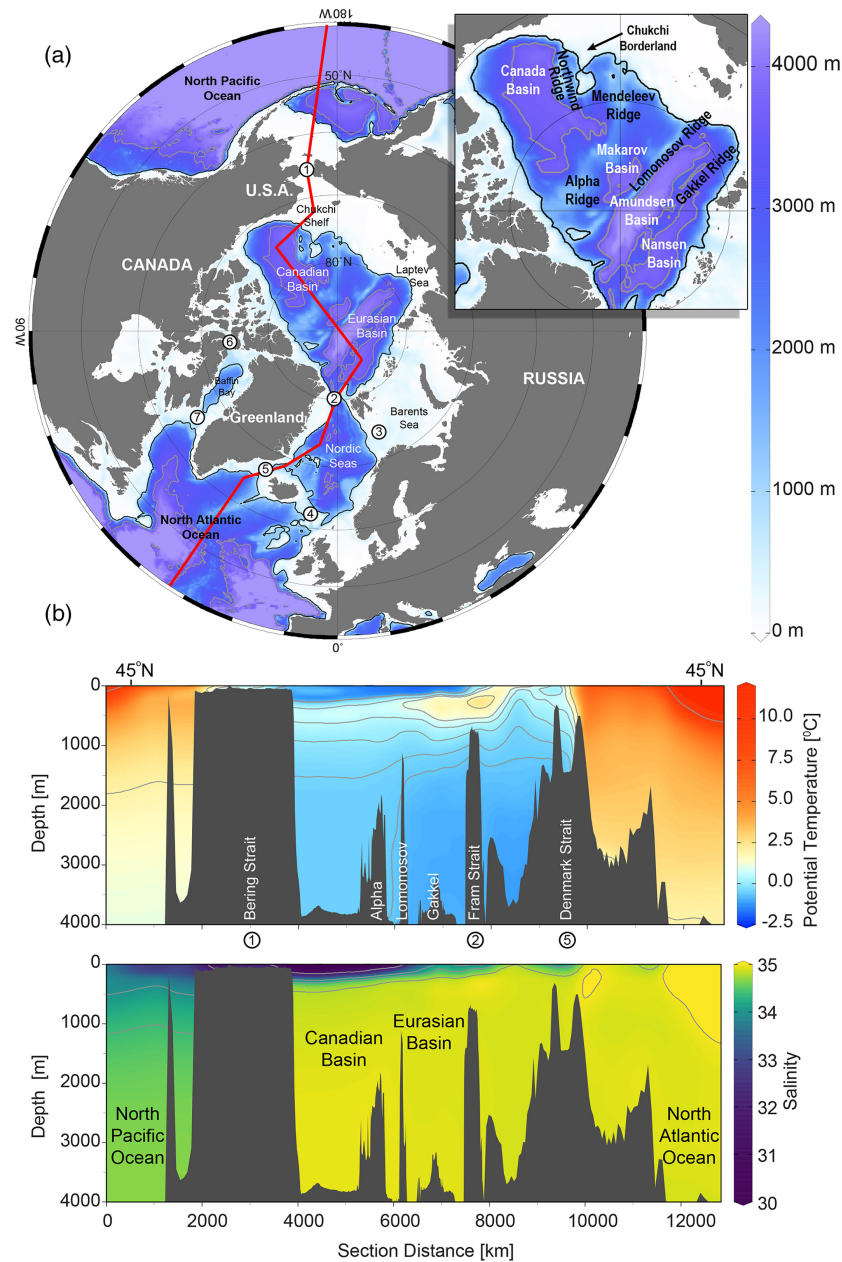


FIGURE 1 – (a) Carte des caractéristiques géographiques principales de l'Arctique ; l'encadré présente l'océan arctique en détails. Les contours bathymétriques à 1000m et 3500m sont montrés et les positions numérotées réfèrent au détroit de Béring (1) et au détroit de Fram (2), à l'ouverture de la mer de Barents (3) , à la crête topographique reliant le Groenland et l'Écosse (4), au détroit du Danemark (5), au détroit de Lancaster (6) et au détroit de Davis (7). La ligne rouge marque la section montrée en (b) (haut) température potentielle (°C) et (bas) section de salinité à partir de l'océan Pacifique (gauche), au travers de l'océan arctique jusqu'à l'océan Atlantique (droite). Les données utilisées proviennent de la Base de données Mondiale sur les Océans(World Ocean Database (WOD18)), toutes les données dans la période 2005-2007 (Boyer et al., 2018), compilées comme par l'Atlas Océanique Mondial (World Ocean Atlas (WOA18))(Garcia et al., 2019). Tirée de Timmermans et al. (2015).

arctique.

On utilise typiquement la *Meridional overturning circulation* (MOC) pour décrire le transport de masse masses d'eaux et de traceurs en fonction de la latitude et de la profondeur

$$\Psi(y, z) = \frac{1}{T} \int_0^T \int_E^O \int_{-H}^z -v(x, y, z', t) dz' dx dt . \quad (1)$$

où v est la vitesse méridionale, H est la profondeur en un point (x, y) , E et O sont les frontières Est et Ouest et T est une période temporelle. La MOC décrit la trajectoire des masses d'eau dans l'océan global. Elle est générée par des flux de quantités de mouvement et des flux de densité à la surface (température et sel) et explique la stratification globale (e.g. Marshall and Speer (2012)). La partie de la MOC qui est diagnostiquée en Atlantique (l'AMOC) est générée par la production d'eaux denses par la convection profonde ayant lieu en hiver aux hautes latitudes. Il existe une faible branche de l'AMOC qui circule dans les mers nordiques¹, au-delà de la crête topographique reliant le Groenland et l'Écosse. Cette branche est constituée en surface par une extension du *gulf stream* et en profondeur par un débordement d'eaux denses de l'Arctique par le détroit de Fram (Fahrbach et al., 2001). Cette circulation dans les mers nordiques gouverne les échanges entre l'Atlantique Nord et l'Arctique. L'intensité de cette circulation dépend non seulement des mécanismes dans l'Atlantique Nord, mais aussi des mécanismes dans l'Arctique lui-même. Par contre, si la MOC nous permet de diagnostiquer la circulation de renversement jusqu'aux mers nordiques, elle ne nous permet pas de le faire dans l'arctique, et ce pour deux raisons principales : (i) le parcours d'intégration tend vers zéro au pôle nord, ce qui implique que la MOC tend elle aussi à zéro à ce point, (ii) une boucle de renversement traversant l'Arctique des mers nordiques jusqu'au Détroit de Béring va subir une suppression due à la géométrie des coordonnées polaires puisque celle-ci va

1. Le terme «mers nordiques» est utilisé dans ce qui suit pour désigner les mers situées entre l'Islande et Svalbard, et incluent les mers de l'Islande, de Norvège et du Groenland.

entrer et sortir d'un même parcours d'intégration. De plus, en utilisant des parcours d'intégration circulaires, des régions dynamiquement très différentes, telles que le Gyre de Beaufort connu comme étant un réservoir d'eaux douces et les mers nordiques lieu de formation d'eaux denses, sont combinées ce qui empêche la dissociation de leurs effets.

La circulation dans l'océan arctique est de type double estuarienne (Stigebrandt, 1981; Carmack and Wassmann, 2006). La Figure 2 présente une version simplifiée de celle-ci qui est formée par deux cellules de transport, l'une dite estuarienne, superposée sur l'autre dite convective. Pour diagnostiquer les deux cellules de cette circulation, il est essentiel d'utiliser des parcours d'intégration adaptés à la géométrie de l'Arctique comme nous le verrons au Chapitre 1.

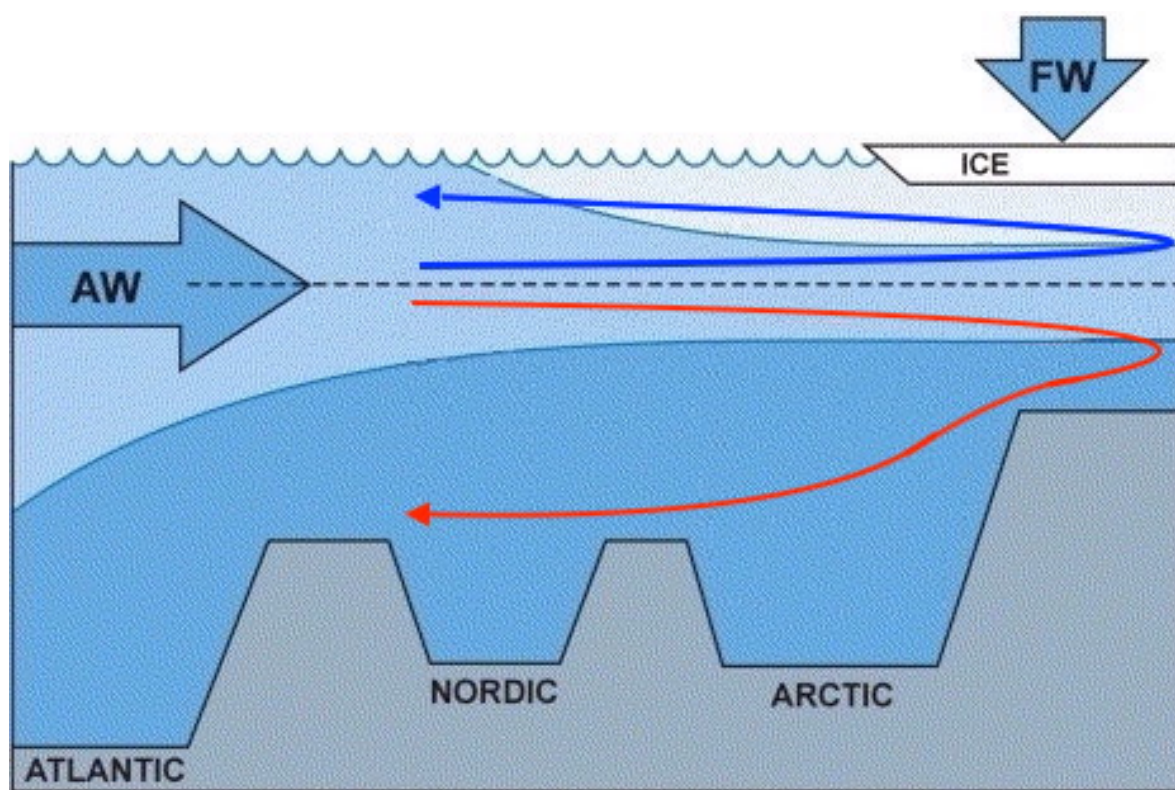


FIGURE 2 – L'océan arctique fonctionne comme un estuaire double. Les mentions AW et FW identifient respectivement le flux d'eaux atlantiques et le flux d'eaux douces. Adapté de Carmack and Wassmann (2006)

Les modèles théoriques décrivant la circulation de renversement dans l'Arctique et la structure de la colonne d'eau nous renseignent sur les principaux mécanismes physiques gouvernant chaque cellule de la circulation de renversement

Les premiers modèles théoriques développés pour décrire la circulation dans l'océan Arctique utilisent une circulation de type estuarienne (Aagaard et al., 1985). Ces modèles sont forcés par la flottabilité en considérant un équilibre entre les flux d'eaux salées provenant des mers nordiques et les flux d'eaux relativement douces quittant l'Arctique. La Figure 3 présente une configuration typique d'un modèle de ce genre incluant une seule boucle de renversement. Dans leur revue de littérature, Timmermans et al. (2015) notent que les modèles théoriques de ce type ont comme caractéristique commune d'utiliser le principe de conservation de la masse et des flux hydrauliques de sorte à ce que le flux constant entre les mers nordiques et l'Arctique soit entretenu par le mélange dans l'Arctique intérieur et un apport de flottabilité. Ainsi, si la cellule estuarienne se comporte comme un estuaire typique, elle devrait être forcée par le gradient horizontal de salinité entre le détroit de Béring et l'Atlantique, l'entraînement dû à l'apport de masse par le détroit de Béring et les rivières et le mélange (Walín, 1977; Burchard et al., 2019).

Certains modèles, tels que ceux de Eldevik and Nilsen (2013), Lambert et al. (2016) et Haine (2021), considèrent une circulation estuarienne double. Celle-ci est formée par le refroidissement et la formation d'eaux profondes dans les mers nordiques (soit une circulation estuarienne négative) et les entrées d'eaux douces qui créent un forçage positif de flottabilité (soit une circulation estuarienne positive). Dans le cas de Lambert et al. (2016), par exemple, le modèle est utilisé pour étudier la sensibilité d'une circulation estuarienne double aux flux d'eaux douces. Il est montré que le système est plus sensible à la distribution spatiale des eaux douces qu'à l'amplitude de leur flux entrant dans la région. Ceci est dû au fait qu'une circulation estuarienne double est plus stable qu'une circulation estuarienne simple et donc que pour sortir la circulation de son état d'équilibre, de forts flux d'eaux douces localisés sont nécessaires. Ces conclusions sont par la suite appliquées à l'Arctique pour émettre l'hypo-

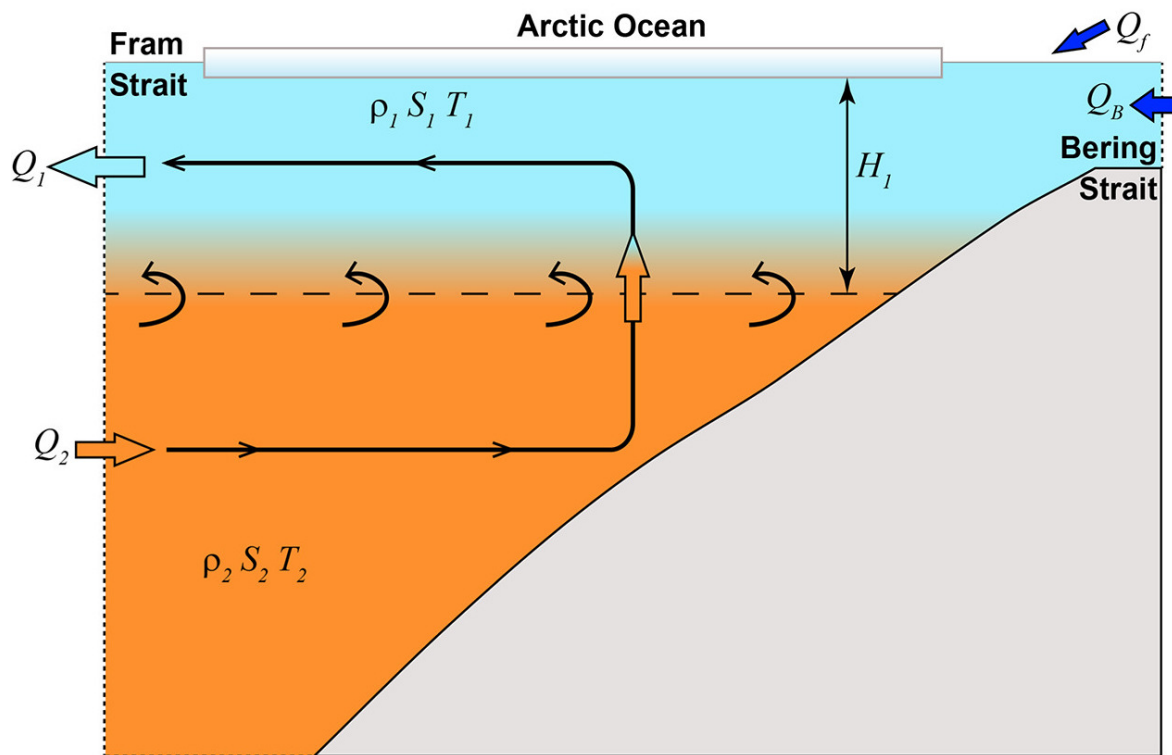


FIGURE 3 – Schématisation d'un système idéalisé de type estuarien à deux couches (voir la Figure 2 de Stigebrandt (1981)). La couche supérieure est constituée des eaux polaires qui vont de l'océan arctique aux mers nordiques se situant à gauche sur le diagramme. La couche inférieure quant à elle est régénérée par les eaux atlantiques allant des mers nordiques à l'océan arctique. Le mélange et l'entraînement des eaux atlantiques dans la couche supérieure dirigent le flux entrant d'eaux atlantiques. Tirée de Timmermans et al. (2015).

thèse que la circulation de renversement est relativement insensible à une augmentation des flux d'eaux douces. À l'inverse, le modèle de Haine (2021) indique qu'en cas d'augmentation des flux d'eaux douces la cellule estuarienne peut s'effondrer. Celle-ci peut également subir le même sort pour une forte augmentation de l'import de chaleur dans le bassin Arctique par les eaux atlantiques. On note toutefois, que ce modèle fixe le transport des eaux atlantiques ce qui n'est pas contraint dans l'océan réel.

Un autre modèle de type circulation estuarienne est celui de Spall (2013). Le modèle présente une version simplifiée de l'Arctique permettant de reproduire les caractéristiques

importantes de l'halocline et de la circulation des eaux atlantiques dans la région. Spall utilise tout d'abord un modèle numérique dans une configuration idéalisée pour créer un modèle de circulation estuarienne auquel est ajouté un rappel de température à la surface, des vents uniformes, une source d'eaux douces à la surface et sur le périmètre du bassin intérieur et un modèle de glace couplé à ce modèle. Dans la même étude, un modèle conceptuel permettant d'estimer plusieurs paramètres tels que la profondeur de l'halocline est suggéré. La concordance des résultats avec ceux du MITgcm supporte la validité du modèle théorique. Celui-ci permet par la suite d'identifier les mécanismes dominants pour la formation de l'halocline et la circulation des eaux atlantiques. L'auteur conclut que ceux-ci sont les flux turbulents aux frontières et la diffusion verticale dans le bassin intérieur.

En utilisant une version du MITgcm dans une configuration idéalisée, Lique et al. (2015) montrent que le rotationnel du vent a une influence significative sur la circulation des eaux Atlantique dans le bassin Arctique et donc qu'il doit être pris en considération pour décrire correctement la circulation de renversement et par le fait même le transport de chaleur en Arctique. Plus spécifiquement, Lique et al. (2015) étudient l'interaction des eaux de surface avec les eaux intermédiaires et le temps de réponse associé à divers forçages. Leur modèle permet l'entrée et la sortie d'eaux dans le bassin et inclut une topographie simplifiée, l'application d'un vent à la surface ayant un rotationnel non nul et un faible rappel des profils de température et salinité avec une constante de rappel de 10 ans. Les auteurs ont choisi de ne pas considérer l'influence de la glace sur le système et la résolution du modèle est trop faible pour étudier l'effet des flux turbulents. L'analyse de sensibilité effectuée dans cette étude suggère que l'intensité de la circulation de surface (gyre de Beaufort) est corrélée linéairement au rotationnel du vent tandis que l'intensité de la circulation des eaux intermédiaires (eaux atlantiques) est corrélée linéairement aux flux entrants par la frontière ouverte. Tandis que l'influence de la couche intermédiaire sur la couche de surface est faible, l'intensité de la circulation de surface peut grandement moduler celle de la couche intermédiaire, confirmant l'effet du rotationnel du vent sur les flux d'eaux atlantiques dans l'Arctique.

Un des modèles théoriques les plus achevé décrivant le couplage entre le rotationnel du vent et la gyre de Beaufort est celui de Doddridge et al. (2019). Ils mettent en relation l'effet du vent, du régulateur glace-océan et des flux turbulents pour décrire la profondeur de la couche halocline dans la gyre de Beaufort. Le régulateur glace-océan est un concept précédemment proposé (Meneghello et al., 2018a,b) qui se base sur le fait que le stress à l'interface glace-océan dépend de la vitesse relative entre la glace et l'océan. Le modèle théorique est validé en utilisant un modèle de gravité réduite considérant une couverture de glace totale. En variant la vitesse de la glace et la diffusivité des flux turbulents, les hypothèses du modèle théorique quant à la sensibilité du gyre de Beaufort à ces paramètres sont vérifiées. Il est montré que la vitesse de la glace influence l'intensité du gyre à l'état d'équilibre et le temps nécessaire pour atteindre cet état. Le modèle de gravité réduite concorde également avec la prédiction du modèle théorique stipulant qu'en l'absence de flux turbulents, la profondeur des isopycnales à l'équilibre augmente anormalement avec la vitesse de la glace selon une relation linéaire. Finalement, la validité du modèle à gravité réduite est confirmée en comparant les résultats de celui-ci à un modèle de circulation globale. Par contre, cette étude est limitée au bassin Arctique et n'inclut pas de frontière ouverte vers les mers nordiques et l'Atlantique Nord. Aucun flux de masse ou de chaleur n'est pris en considération dans ce système fermé.

Finalement, en utilisant des parcours d'intégration alternatifs, Bretones et al. (2022) s'intéressent particulièrement à la cellule convective de la circulation estuarienne double et à son lien avec l'AMOC. Les auteurs trouvent qu'un déplacement des zones de convection associé à l'entrée des eaux atlantiques vers le nord à la suite du retrait de la couverture de glace amplifie l'intensité de la cellule convective dans le bassin eurasién. Cela met en évidence le rôle de l'AMOC sur la circulation de renversement arctique.

Ces études contribuent à la compréhension de la circulation de renversement en Arctique, mais ne permettent pas de diagnostiquer simultanément les deux cellules de la circulation estuarienne double puisque les approches couramment utilisées pour diagnostiquer

le renversement arctique induisent une perte d'information. Dans ce contexte, le premier objectif de cette thèse est de caractériser la cellule estuarienne et la cellule convective de la circulation estuarienne double arctique. Le deuxième objectif est d'identifier les mécanismes de forçage dominants associés à chacune de ces cellules. Le troisième objectif est d'analyser l'influence de la circulation de renversement arctique sur le flux de chaleur vers le bassin arctique. Ces trois objectifs s'inscrivent dans l'objectif global de cette thèse, soit d'améliorer notre compréhension théorique de la circulation estuarienne double arctique.

Pour ce faire, nous avons exploré la dynamique de la circulation estuarienne double en deux temps. Le chapitre 1 présente la nouvelle approche que nous avons développée afin de diagnostiquer simultanément les deux cellules de la circulation estuarienne double. Par la suite, à l'aide du modèle *Community Earth System Model - large ensemble* (CESM-LE) (Kay et al., 2015) nous avons caractérisé l'évolution de cette circulation pour un scénario de forçage historique et le scénario de réchauffement climatique *Representative Concentration Pathway 8.5* (RCP8.5). Subséquemment, nous établissons des corrélations entre les mécanismes de forçage suggérés dans la littérature et l'intensité des cellules estuarienne et convective. Le modèle CESM-LE est un modèle du *Coupled Model Intercomparison Project* (CMIP) pour lequel seules des sorties de modèles sont accessibles et la grille du modèle océanique limite les analyses que nous pouvons effectuer avec celui-ci. Ainsi, dans le deuxième chapitre de cette thèse, pour approfondir notre compréhension de l'influence des mécanismes de forçage, nous avons créé un modèle simplifié de l'Arctique. Nous utilisons une configuration du MITgcm, un modèle de circulation globale résolvant les équations de Boussinesq pour un fluide incompressible à l'aide de la méthode des volumes finis (Marshall et al., 1997), sur laquelle il est facile d'évaluer le bilan de chaleur du domain entier. Avec ce modèle, nous explorons l'influence des flux de flottabilité, du transport de masse et de sel au détroit de Béring, du vent, de la diffusivité et de la stratification dans les mers nordiques sur la circulation de renversement arctique. Cela nous permet d'établir une relation entre les mécanismes de forçages et les cellules de la circulation de renversement. Par la suite, nous évaluons l'influence de ces cellules sur les flux de chaleur et la température à l'équilibre du bassin arctique.

ARTICLE 1

MODIFICATIONS DE LA CIRCULATION ESTUARIENNE DOUBLE ARCTIQUE INDUITE PAR LE RÉCHAUFFEMENT CLIMATIQUE

Cet article, intitulé “*Warming-Induced Changes in the Arctic Ocean’s Double Estuarine Circulation*”, est en cours de préparation pour être soumis au *Journal of Physical Oceanography*. Il est le résultat d’un travail conjoint avec les Professeurs Louis-Philippe Nadeau, David Straub et Bruno Tremblay. L’idée originale de ce projet de recherche est le résultat de discussions ayant eues au sein d’un groupe de lecture incluant les auteurs de ce papier. En tant que première auteure, ma contribution fut essentielle à l’élaboration de la revue de littérature, au développement de la méthodologie innovante, à l’analyse et à l’interprétation des résultats ainsi qu’à la rédaction de l’article. Le Professeur Louis-Philippe Nadeau, deuxième auteur, a été présent à toutes les étapes du projet. Il m’a guidé dans le développement de la méthode, l’analyse des résultats, l’interprétation de ceux-ci et finalement a contribué à la révision de l’article. David Straub, troisième auteur, a apporté des commentaires à l’interprétation des résultats et a participé à la rédaction de l’article. Finalement, Bruno Tremblay a fourni l’idée originale du projet et a participé à l’interprétation des résultats. Ces résultats ont été présentés à la conférence *Ocean Sciences* à l’hiver 2024 et au *First General Assembly of the Norwegian Geophysical Society* à l’été 2025.

1.1 Résumé en français

L’amplification arctique et l’augmentation de la température de l’océan Arctique engendrent un intérêt grandissant quant à l’influence de cet océan sur le climat aux hautes latitudes. Une grande partie de cette influence est liée à l’effet des conditions arctiques sur la circulation méridionale de renversement de l’Atlantique (AMOC). Pour examiner ces liens, il est préférable d’étendre l’AMOC dans le bassin arctique. Définir cette extension de manière triviale en utilisant des parcours d’intégration suivant des parallèles de latitude n’est pas idéal

puisque la circulation arctique ne suit pas une symétrie zonale. Ici, deux extensions aux parcours d'intégration sont proposées. La première est une extension le long de "l'axe court" s'étalant de la côte russe vers la côte ouest du Groenland. La deuxième suit un "axe long" allant du détroit de Béring au détroit de Fram. Ces deux approches décrivent de manière plus informative la circulation. Le modèle CESM forcé à l'aide du scénario historique et RCP8.5 (1920-2100) est utilisé pour diagnostiquer et comparer la cellule estuarienne et la cellule convective plus profonde le long de ces parcours. Pour l'axe court, une cellule convective simple est observée alors que le long de l'axe long, une circulation estuarienne double similaire à celle préalablement présentée dans la littérature devient évidente. Augmenter la stratification dans l'Atlantique Nord en conséquence du réchauffement climatique amplifie la cellule convective de la circulation et engendre une colonne d'eau plus chaude au-dessus de 1500m. Cette observation est cohérente avec l'atlantification de l'Arctique actuelle telle que rapportée dans la littérature. Cette amplification, dans le temps, de la branche inférieure de la circulation semble être causée par une augmentation des flux négatifs de flottabilité à la surface et ne semble pas être contrôlée par des mécanismes de forçage délocalisés. D'un autre côté, la branche estuarienne de la circulation ne gagne pas en intensité avec la progression du réchauffement climatique et ne semble pas être corrélée avec les flux de flottabilité locaux à la surface. Cela pourrait être un indicateur du temps de réponse de la cellule estuarienne à l'apport en eau douce.

1.2 Abstract

In the context of Arctic Amplification, there has been increasing interest in how changing conditions in the Arctic are linked to the Atlantic Meridional Overturning Circulation (AMOC). To examine these links, it is desirable to extend the AMOC into the Arctic basin. Simply defining this extension based on zonal integrals is not ideal in that the overturning circulation of the Arctic does not follow a zonal symmetry. Here, two alternative extension pathways are proposed: a ‘short axis’ extending from the coast of Russia towards the western side of Greenland and a ‘long axis’ extending from Bering Strait towards Fram Strait. The CESM model, run under historical and RCP8.5 forcing (1920–2100), is used to diagnose and compare the upper estuarine and deeper convective overturning circulations along these two pathways. Along the short axis, a single overturning cell appears whereas along the long axis a double estuarine circulation similar to that previously introduced in the literature becomes evident. Increasing stratification in the North Atlantic related to global warming enhances the overturning branch of the circulation and results in a warmer water column above 1500 m, consistent with an ongoing Atlantification of the Arctic previously reported in the literature. This amplification over time of the lower branch appears to be due to increased negative buoyancy fluxes at the surface and does not seem to be correlated with remote forcing. The estuarine branch of the circulation, on the other hand, does not increase in intensity as global warming progresses and does not appear to correlate with local freshwater fluxes at the surface.

1.3 Introduction

The Arctic is disproportionately affected by global warming (Serreze and Barry, 2011; Steele et al., 2008; Yadav et al., 2020) and related to this Arctic amplification is a rapid and extensive decline of sea ice (Holland et al., 2006; Serreze et al., 2007) which further intensifies Arctic warming through the sea ice–albedo feedback (Screen and Simmonds, 2010;

Holland and Bitz, 2003; Serreze and Barry, 2011). Increasing evidence also suggests that Arctic amplification is accelerated by changes in air-sea heat fluxes and Arctic circulation, and this has prompted a growing interest in the overturning circulation in the Nordic Seas and Arctic Ocean (Timmermans and Marshall, 2020; Haine, 2021; Wang et al., 2024). For example, there is a growing consensus that Arctic sea ice loss is shifting North Atlantic deep-water formation northward into the Arctic Basin (Lique et al., 2018; Bretones et al., 2022; Bitz et al., 2006). This serves to strengthen a deep overturning cell that imports Atlantic origin waters to the Arctic in its upper branch and exports denser waters in its lower branch. Evidence also show that lying above this convectively driven deep Arctic overturning cell is a shallower estuarine cell (Brown et al., 2025). How best to characterize the spatial structure of this double estuarine circulation and the interplay between the two cells of which it is comprised, however, remains poorly understood.

Previous studies of Arctic overturning circulation have focused on the import of Atlantic-origin water and an overall Atlantification of the Arctic (Asbjørnsen et al., 2020; Muilwijk et al., 2018; Wang et al., 2024). This process transports heat into the Arctic and influences dense water formation over the Eurasian shelves (e.g., Bitz et al., 2006; Spall, 2013; Lique et al., 2015; Oldenburg et al., 2018; Bretones et al., 2022), which has profound implications for the region's stratification, circulation, and for climate feedbacks. Atlantification can be thought of in terms of the Atlantic Meridional Overturning Circulation (AMOC) evolving so as to extend into the Arctic (Ingvaldsen et al., 2021; Zhang and Thomas, 2021).

The AMOC is traditionally defined in terms of zonal integrals; however, for a variety of reasons, this definition is ill-suited to the Arctic Ocean geometry. First, convergence of meridians implies that zonally-integrated transport vanishes at the pole, so that an Arctic extension of the AMOC based on zonally integrated transport would terminate in the middle of the Arctic basin. Additionally, a throughflow extending between Bering and Fram Straits would largely cancel out in a zonally-integrated framework. Similarly, latitudinal integration also lumps together other dynamically distinct regions such as Beaufort Gyre, known for its

freshwater storage, with the Nordic Seas, known as a key site of dense water formation.

Recognizing these limitations, Bitz et al. (2006) and Bretones et al. (2022) used alternative integration pathways to trace the northward extension of the AMOC into the Arctic. Both studies found a single, convectively-driven cell extending between Fram Strait and the Eurasian shelves, which are known to be a site of deep water formation (Lique et al., 2018; Polyakov et al., 2020). Zhang and Thomas (2021) went further, arguing that the AMOC does not merely reach the Nordic Seas but extends continuously into the Arctic basin as a unified Atlantic-Arctic overturning cell. Whether it should properly be identified as part of the AMOC, however, remains an open question: deep water formation in the Arctic is driven by shelf convection and brine rejection rather than by the open-ocean deep convection that characterizes the North Atlantic (Rhein et al., 2011; Buckley and Marshall, 2016), raising the question of whether these are extensions of the same circulation system or dynamically distinct phenomena.

A growing consensus suggests that there is also a shallower estuarine cell in the near-surface waters so that, globally, the Arctic overturning circulation can be thought of as a two-cell system (e.g., Stigebrandt, 1981; Carmack and Wassmann, 2006; Rudels, 2012; Rudels and Carmack, 2022a; Tsubouchi et al., 2024; Brown et al., 2025; Dörr et al., 2026). Theoretical support for this comes from modelling studies (e.g., Aagaard et al., 1985; Eldevik and Nilsen, 2013; Lambert et al., 2016; Haine, 2021). In a classical estuarine circulation, diapycnal mixing progressively increases the salinity of the surface waters as they flow out to sea. To balance both the mass and salt budgets then requires that there be an inflow of ocean water at depth. That is, an overturning cell is implied. Transport in the upper branch of this overturning cell is both stronger and more saline than the upstream river transport which provides a freshwater input at the surface (Cardoso, 2022). In the Arctic context, it is a combination of river runoff and the relatively low salinity northward transport through Bering Strait that provide the net throughflow. This is thought to lead to an estuarine-like cell that sits above a convectively-driven cell similar to the cell diagnosed by Bitz et al.

(2006) and Bretones et al. (2022) and fuelled by dense water formation related to surface heat loss, brine rejection, and sinking over the Eurasian shelves (Lique et al., 2018; Carmack and Wassmann, 2006). Throughout this work, we refer to these two cells of the double estuarine circulation as the estuarine cell and the convective cell, respectively.

The main goal of this study is to assess how the structure of overturning cells in the Arctic is influenced by the choice of integration pathway, and to identify a compelling pathway from which to diagnose the double estuarine structure of Arctic overturning. We find that the apparent contradiction between model-based diagnoses (which show a single cell) and theoretical descriptions of the Arctic (which suggest two cells) is directly related to the choice of integration path. We compare two main pathways: a short axis extending from Greenland to the East Siberian Sea, and a long axis extending from Bering Strait to Fram Strait. Using output from CESM simulations, we show the convective cell projects onto both (although it projects more strongly on the short axis). The upper, estuarine cell, however, is evident when using the long axis but nearly absent for the short axis. This distinction reflects the driving mechanisms of the two cells, for example with the convective cell being fed by dense water formation along the Eurasian shelves, with the short axis originating in the center of this region. Conversely, the estuarine cell is driven by basin-wide upwelling that connects a deep inflow of Atlantic-origin water to a surface outflow of freshwater extending between Bering and Fram Straights. Because the sense of this outflow is essentially perpendicular to the short axis, the long axis is better suited for diagnosing this circulation.

We are also interested in how these two cells evolve under the RCP8.5 forcing scenario. We find the convective cell to strengthen and shallow with time, with this being linked to increases in both mid-depth stratification and negative winter surface buoyancy fluxes. As the convective cell shallows, the estuarine cell becomes increasingly confined to the surface, although its transport remains largely unchanged. The paper is organized as follows. Section 2 describes the integration pathways and methodology. Section 3 presents the results, and conclusions are given in Section 4.

1.4 Methodology

We use a CMIP5 model, which provides standardized global warming scenarios and a robust representation of the global circulation. Specifically, we selected the CESM-LE model (Kay et al., 2015). Its nominal 1° resolution does not permit eddy dynamics, which are instead parameterized. Related to this parametrization, bolus and submesoscale velocities are provided in the model output. We use these together with the monthly-mean Eulerian velocity for our analysis. Like most CMIP models, not all the detailed features of the Arctic are accurately reproduced (Heuzé et al., 2023). Large scale circulation features are, however, reasonably well produced and we expect the main features of our analysis to be robust to choice of model.

We use the historical scenario covering 1920-2006 and the RCP8.5 global warming scenario covering 2006-2100. The RCP8.5 scenario is considered the ‘baseline’ scenario without the introduction of policies to reduce global warming (Riahi et al., 2011). It also has the highest baseline emission in the Intergovernmental Panel on Climate Change (IPCC) fifth assessment report and in the CMIP5 experiments as other scenarios predict a stabilization or reduction of greenhouse gas emission. The RCP8.5 scenario is therefore a conservative view of greenhouse gas emission following current trends (Riahi et al., 2011). It provides a framework to investigate the potential for a regime change and the possible collapse of one of the branches of the double estuarine circulation. Preliminary analyses were conducted on multiple ensemble members of the model. Since only a small variability was found between the members and this study is interested in the general features of the circulation, a representative member was selected for the full analysis.

To define the AMOC, one typically integrates along a zonal path. Here, we instead integrate along different paths, which are perpendicular to two natural axes of the Arctic and are approximately perpendicular to one another. The first or “short axis” extends from the

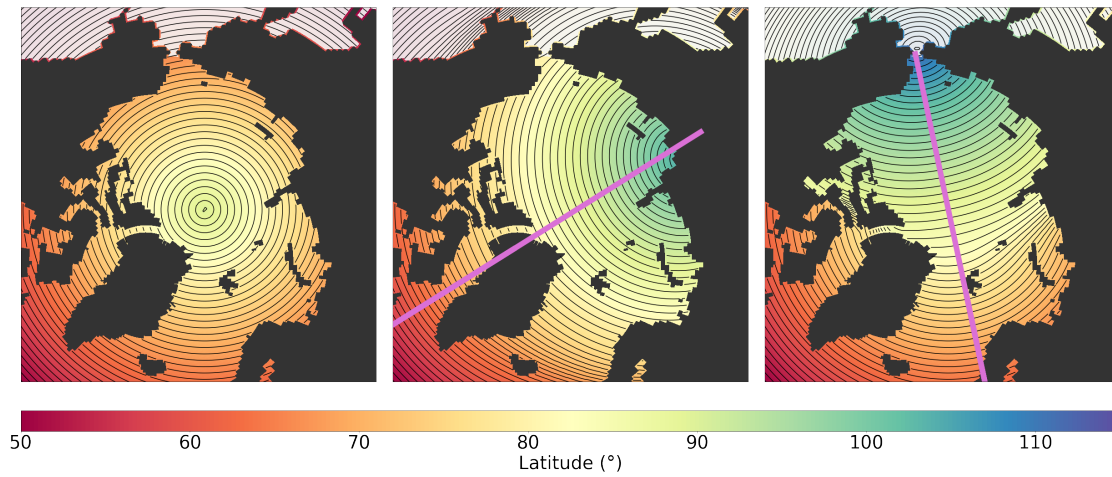


Figure 4 – Integration paths used in this study: (left) latitudinal paths, (middle) short-axis paths, and (right) long-axis paths. Black lines indicate the individual paths, and the pink line shows the axis onto which they are projected.

East Siberian Sea to Northwestern Greenland (Fig. 4 middle) and the second, “long axis” goes from Bering Strait to Fram Strait (Fig. 4, right). Other intermediate choices were also explored, but as these gave results that were intermediate to those for long and short axes, they will not be presented here. To do this, we use the method of Roberts et al. (2006) to displace the north pole to a location of our choosing. Using the displaced coordinates, we obtain new “zonal” paths, which are approximately perpendicular to the short and long axes, visualized by the pink lines in Fig. 4. The exact positions of these lines are somewhat arbitrary, as any displaced longitude could have been used for this purpose.

To facilitate comparison across projection choices and with previous studies, coordinate systems for both the long and short axes are blended smoothly back to the standard latitude-longitude coordinate system in the North Atlantic. This blending is implemented over a transition region. For the short axis, this blending is relatively straightforward since the original and displaced coordinate systems closely coincide over the transition region (around 70°N in the standard grid, Fig. 14). For the long axis, the displaced “latitudes” are nearly perpendicular to meridians over much of the transition region. To isolate the Baffin Bay

circulation from the broader Arctic circulation, and to minimize distortion, we use the shallow regions of the archipelago to impose a sharp transition between the original and displaced grids (Fig. 13). (See the appendix for more details about the method).

To characterize the circulation, we compute an overturning streamfunction by integrating the time-averaged vertical velocity in the modified latitude–longitude coordinates (y and x , respectively). Letting Y_N denote the Arctic end of a given path or axis, the streamfunction is defined as

$$\Psi(y, z, t) = \frac{1}{T} \int_0^T \int_y^{Y_N} \int_{X_E}^{X_W} w(x, y', z, t), dx, dy', dt, \quad (1.1)$$

where w is the monthly averaged vertical velocity, X_E and X_W mark the eastern and western limits of each modified latitude circle (Fig. 4), and T is an averaging period specified separately for each analysis where applicable. The resulting vertical and along-axis transports are given by $\Psi_y = W$ and $-\Psi_z = V$.

Note that one might instead have chosen to compute Ψ using vertical integrals of horizontal velocity in depth; however, because CESM uses an Arakawa-B grid, an interpolation is required and we found results to be sensitive to how this was carried out. For this reason, we prefer to define Ψ by integrating vertical velocity, w , as above.

The overturning is also commonly computed in density rather than depth coordinates, since the two can differ substantially in regions with significant zonal isopycnal deviations, such as the Southern Ocean (e.g., Hirst and McDougall, 1998). Indeed, a growing body of literature on Arctic and subpolar dynamics has favored diagnosing the streamfunction in isopycnal space (Pickart and Spall, 2007; Oldenburg et al., 2022; Tsubouchi et al., 2021; Brown et al., 2025), an approach that is particularly valuable for studying heat transport at the Arctic gateways and for relating the circulation to surface water mass transformation. However, integrating vertical velocity in density-space bins likewise requires interpolation, and our results proved similarly sensitive to that choice. We therefore adopt depth coordinates, which is justified given that our focus is the Arctic basin, where isopycnals are approximately horizontal

and the distinction between depth and isopycnal coordinates is expected to be minimal.

Horizontal velocities are nonetheless required at the Bering Strait, where they are incorporated into the calculation in different ways depending on the chosen integration axis. For the long axis, the Bering throughflow served to constrain Ψ at Y_N , as this latitude marks the origin of the integral. Along the short axis, it was instead incorporated into eq. 1.1 as an imposed horizontal flux at the latitudinal coordinate corresponding to the strait.

A technical issue arises in that the CESM model parametrizes overflow in the Greenland, Iceland and Norwegian (GIN) Seas region by injecting and removing water through the bottom topography, thus creating open boundary conditions on either side of this region of parameterized flow (Smith et al., 2010). To avoid the complexities of explicitly accounting for these parameterized overflow fluxes, we calculate Ψ using eq. 1.1 integrating from south to north up to GIN sea and then from north to south on the other side. A small discontinuity in the streamfunction occurs at the GIN Seas junction—where the North Atlantic and Arctic panels meet in Fig. 5—but visual inspection suggests it is negligible.

1.5 Results

1.5.1 Influence of the integration paths geometry

We begin our investigation by contrasting how the circulation projects onto the long and short axes, and how these projections differ from those obtained using standard spherical coordinates. The left panels of Fig. 5 show the geometry of the integration pathways, with the pink line indicating the horizontal distance used to plot the overturning streamfunction in the right panels. To allow a consistent comparison across coordinate systems, the 53rd parallel in the standard polar grid (green dashed line in Fig. 5) is used as a reference point. South of this, all three coordinate systems coincide. Distances are then measured northward along the pink “axis.” As an additional visual reference, the location where each pathway crosses

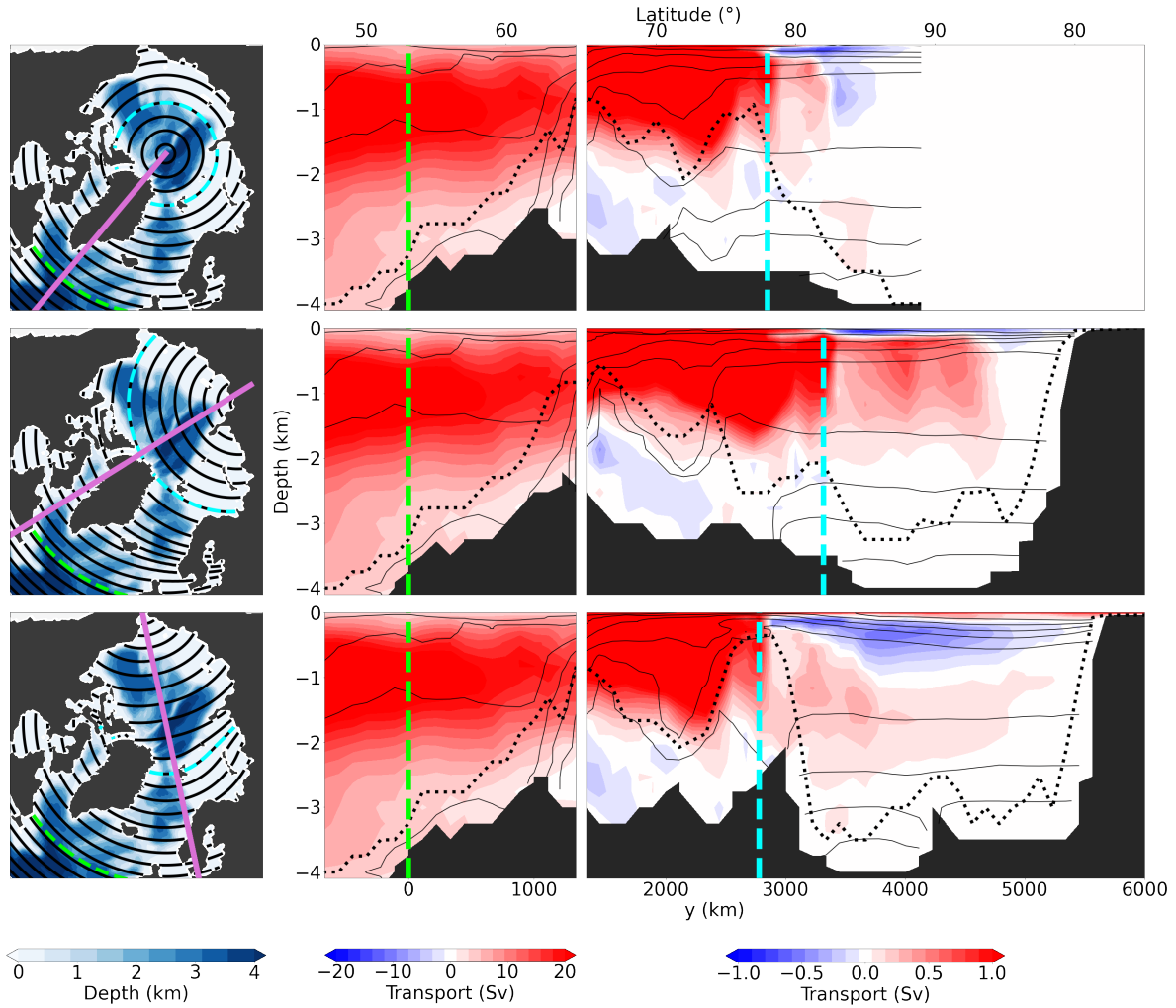


Figure 5 – Averaged overturning circulation computed along various integration paths (shown at left) using CESM-LE data for the historical scenario (1920–1925). The black solid lines are σ_2 isopycnals ranging from 36 to 37.46 kg/m^3 on a log scale. The black dotted line and black shaded region indicate the 75th and 99th percentiles of the topography along each integration path, respectively. The green dashed line marks the southernmost path beyond which all configurations become identical. The cyan dashed line marks the path crossing Fram Strait. The pink line is used to calculate the distance from the origin in the right panels.

Fram Strait (which separates the Arctic Basin from the Nordic Seas) is indicated by the cyan dashed line.

The standard polar coordinates (top row) reveal limited information in the Arctic, as the circulation is confined to a small region, obscuring its broader extent. Moreover, since the latitudinal range of the Beaufort Gyre and of the Nordic Seas overlap, this projection combines the two. Then, since circulation in the Nordic Seas is stronger, its influence on Ψ masks that of the Beaufort Gyre. The short axis (middle row) aligns roughly with pathways used by Bitz et al. (2006) and Bretones et al. (2022), and the circulation we diagnosed for this path reveals a strong convective cell, consistent with their results. Some hint of an estuarine cell is also evident at y positions close to the Siberian coast (e.g., $y > 3500$ km), although this cell is weak and confined to a thin layer of the surface. Overall, this pattern reflects the well-established view that dense water formed through brine rejection originates primarily on the Eurasian shelves, where it drives strong convection.

As projected onto the short axis, the convective cell dominates and the estuarine cell is visible only at very shallow depths. Projected onto the long axis, contributions from the Canadian Basin, Eurasian Basins and Nordic Seas are not lumped together, but rather occur at different values of y along the axis. In the bottom row of Fig. 5, we see a more robust double estuarine structure, similar to that described in the literature (e.g., Tsubouchi et al., 2024; Brown et al., 2025; Dörr et al., 2026). In particular, the estuarine cell occupies a much larger range of depths, in places extending as deep as 1000m. Note that the streamfunction reverses signs (becomes red in the figure) in a thin layer above the estuarine cell. This is not a counter-rotating cell, but rather corresponds to the net throughflow associated with Bering Strait transport. In what follows, we focus on the long and short paths with an emphasis on the former, which is better suited to describe the Arctic's double estuarine circulation.

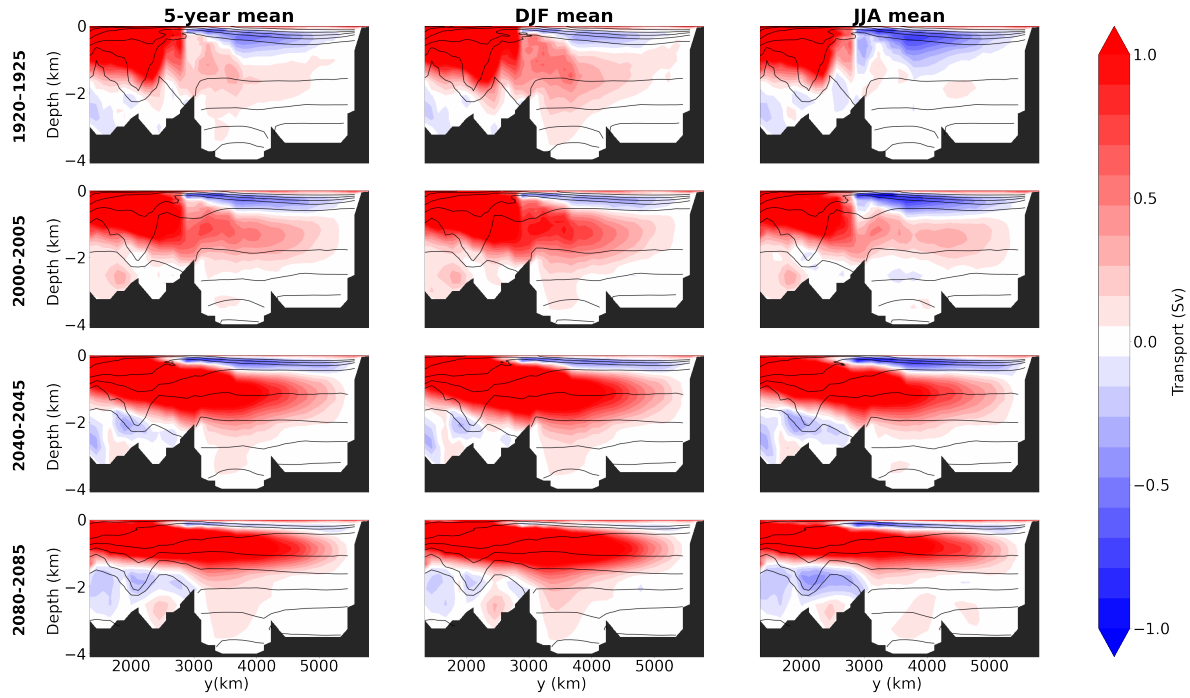


Figure 6 – Overturning circulation calculated along the long axis using CESM-LE data for the historical scenario (1920-1925 and 2000-2005) and the RCP8.5 scenario (2040-2045 and 2080-2085). The circulations are averaged over (right) a five-year period, (middle) the months of December–February, and (left) the months of June–August. The black solid lines are the isopycnals of σ_2 ranging from 36 to 37.46 on a log scale. The black region represents the 99th percentile.

1.5.2 Seasonal cycle and temporal evolution

The seasonal cycle of the convective cell is very similar as diagnosed along the long and short axes and, since the estuarine cell is all but absent as diagnosed along the short axis, here we focus on long axis results. Fig. 6 shows a pronounced seasonal cycle in the double estuarine circulation in the early 20th century before significant sea ice decline. For example, between 1920 and 1925, summertime Ψ shows the estuarine cell to be dominant and the convective cell to be all but absent. Conversely, in wintertime the convective cell is strong and the estuarine cell is reduced. While this seasonality persists over time, the annual mean reveals a marked strengthening and shoaling of the convective cell during the 21st-century, which progressively compresses the estuarine branch toward the surface (e.g., $y > 3500$ km).

This intensification of the convective cell aligns with findings by Bitz et al. (2006) and Bretones et al. (2022), who attribute the shift to the northward retreat of the winter sea ice edge into the Arctic Basin. Haine (2021) hypothesized that the associated increase in warm Atlantic Water inflow could trigger a “heat crisis,” wherein the estuarine branch collapses as the cold halocline layer vanishes. As the Eurasian Basin transitions from a salt-dominated β -ocean to a temperature-dominated α -ocean (Carmack, 2007), the halocline’s insulating capacity erodes. This transition enhances upward heat flux (Polyakov et al., 2017), creating a positive feedback loop that accelerates ice melt and further weakens the β -barrier. Crucially, this regime shift suggests that atmospheric cooling may supersede salt release (brine rejection) as the primary driver of surface water mass transformations (Lambert et al., 2018).

We also note that both the convective and the estuarine cells shift closer to the surface with time. Isopycnals (black lines) also deepen with time (Fig. 6), with this being related to an increase in mid-depth stratification. This increased stratification is likely related to the shallowing of both overturning cells and is consistent with the trend of Atlantification previously reported in literature (Wang et al., 2024). This concept is further explored in the next section where we analyze the role of remote and local forcing mechanisms in these

changes.

1.5.3 Link to the North Atlantic and the Nordic Seas

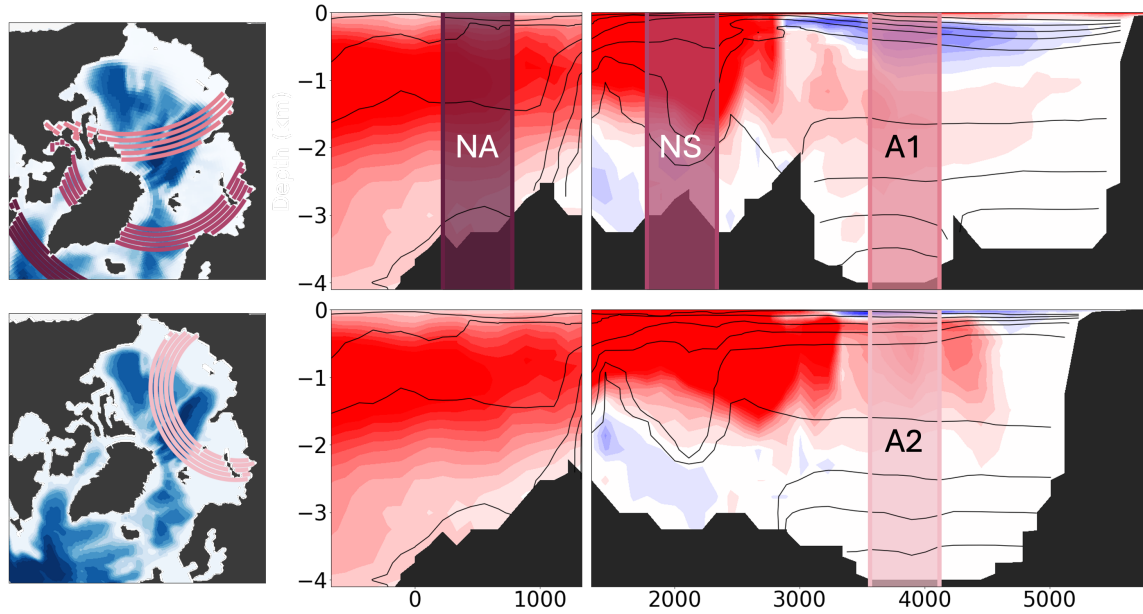


Figure 7 – Regions of the North Atlantic (NA), Nordic Seas (NS), Arctic along the long axis projection (A1) and the short axis projection (A2) over which the circulation, stratification, temperature and salinity are averaged in Fig. 8 and 9. The paths associated with each region are shown in the corresponding color in the left panels.

In the previous section, we noted that the Arctic’s convective cell is strengthening with time. Here, we explore how this change is related to changes in the North Atlantic and the Nordic Seas. For this, we average the overturning circulation over zones (ranges of y) corresponding to North Atlantic, Nordic Seas along our long axis and to Arctic conditions for both the long and short axes (see regions NA, NS, A1 and A2 in Fig. 7). We find a common feature across all four transects in Fig. 8: the convective cell shoals with time. Its strength, by contrast, is seen to weaken in the North Atlantic and to strengthen in the Arctic. This suggests that as the AMOC weakens, it is also extending northward to become a combined

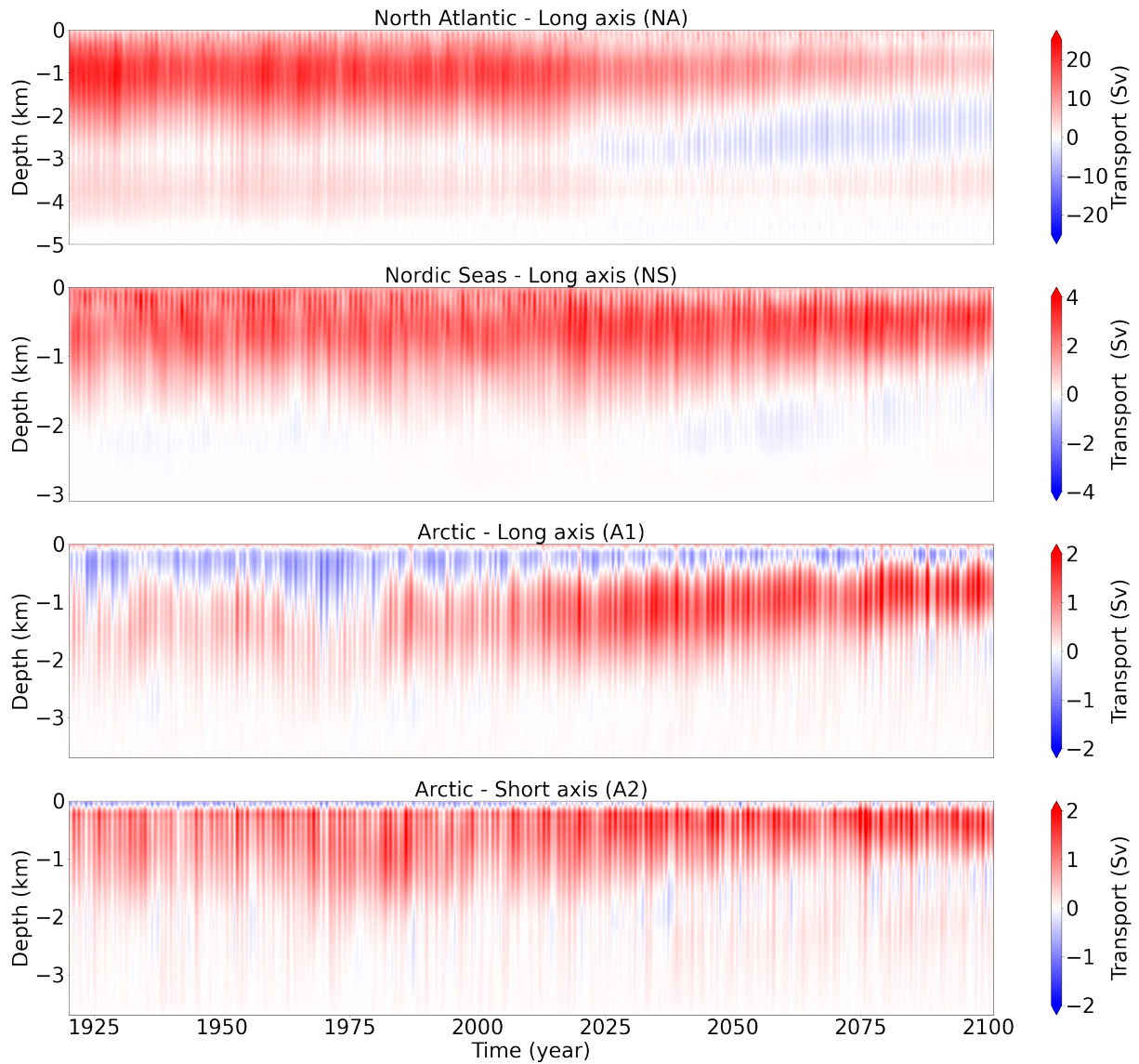


Figure 8 – Time evolution of the overturning circulation for (first row) the North Atlantic, (second row) the Nordic Seas, (third row) the Arctic along the long axis, and (fourth row) the Arctic along the short axis, for the historical (1920–2006) and RCP8.5 (2006–2100) scenarios. Regions are as defined in Fig. 7. A Gaussian moving average with a six-month standard deviation has been applied to each Hovmöller diagram.

Atlantic-Arctic cell. The Nordic Seas region, as diagnosed here, serves as an inflection point between weakening of the cell on the Atlantic side and strengthening on the Arctic side.

That the convective cell is weakening in the North Atlantic is consistent with previous work (Cheng et al., 2013; Ma et al., 2024; Liu et al., 2019). Similarly, strengthening of the convective cell in the Arctic is also consistent with results by Bitz et al. (2006) and Bretones et al. (2022) for the same time period. The literature relating to how conditions are evolving in the Nordic Seas is more mixed. Our results are consistent with Oldenburg et al. (2024) but in seeming contradiction to Arthun et al. (2023), who suggested that the influx of water from the Nordic Seas to the Arctic should increase under global warming. This apparent discrepancy may be model-related or may simply be a reflection of the large spatial variability within the Nordic Seas. For example, the net convective cell strength in our Nordic Seas band is the sum of opposite-signed contributions with, for example, circulation west of Greenland serving to weaken the cell and circulation to the east of Greenland serving to strengthen it (not shown).

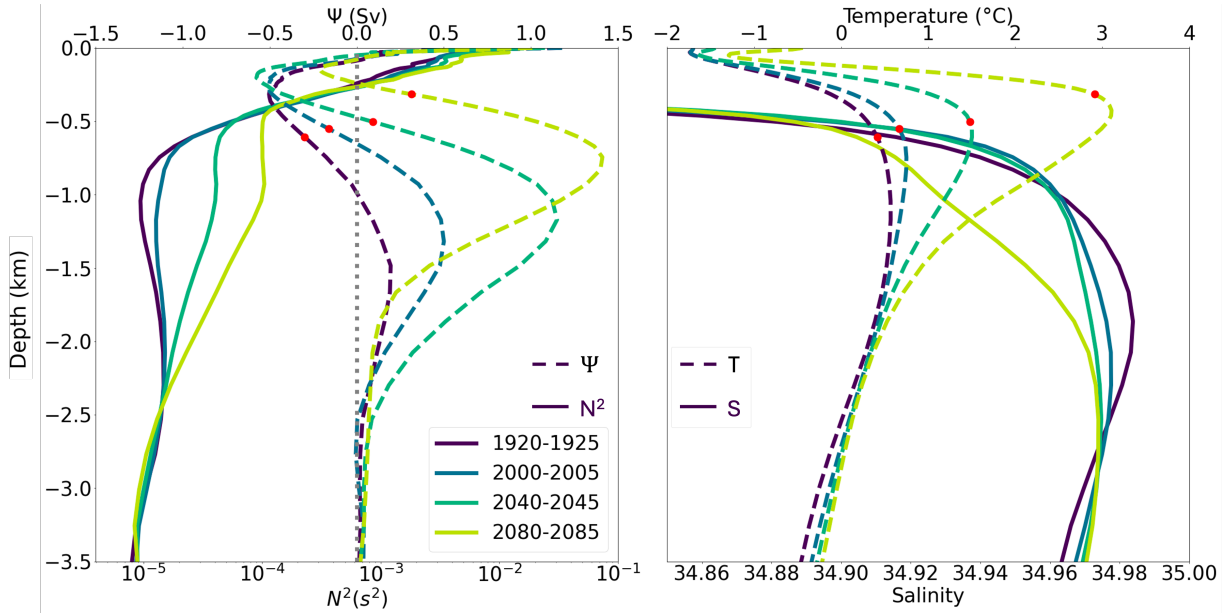


Figure 9 – Vertical profiles of (left) N^2 and Ψ and (right) temperature salinity averaged over the Arctic region A1 in Fig. 7. Colors correspond to four different time-averaging periods. The red dot marks the approximate depth of maximum Atlantic Water import into the Arctic.

The shallowing of the convective cell occurs in tandem with an evolution in the Arctic basin's stratification. This is visualized in Fig. 9, which shows the stratification averaged over

the Arctic region A1 (Fig. 7) to increase markedly in a layer centered near 1000m depth, with N^2 increasing by an order of magnitude there between the early 1920s and the 2080s. Above this, Atlantic Water is transported into the Arctic via the upper branch of the convective cell and the lower branch of the estuarine cell. We identify the core of this Atlantic Water inflow by the depth of maximum transport into the basin (where $d\Psi/dz$ reaches its maximum negative value), indicated by red dots in Fig. 9. Notably, these depths also coincide with the depth of maximal temperature (red dot in Fig. 9b). The progression of Atlantification is further evidenced by the evolution of temperature, salinity, and density averaged along the long axis paths (Fig. 10). Over time, the Atlantic Water inflow layer exhibits a marked increase in temperature and a decrease in density, which effectively enhances the mid-depth stratification. Salinity is seen to decrease over this same time period, although this change is concentrated in the near-surface portion of the water column. These results are consistent with finding by Polyakov et al. (2023); Kenigson and Timmermans (2021); Timmermans and Marshall (2020).

The link between Nordic Seas warming and the weakening and shoaling of the AMOC can be understood through the meridional density contrast between the northern deep-water formation regions and the broader Atlantic basin. Jansen et al. (2018) showed that both the transient and equilibrium AMOC responses are well captured by a thermal wind balance tied to this north–south density contrast (Gnanadesikan, 1999). Surface warming initially reduces the density of northern-sourced deep waters faster than that of the broader Atlantic basin, thereby weakening and shoaling the overturning on decadal timescales. In the Nordic Seas, the observed warming is consistent with this picture, where local restratification driven by weakening deep convection (Fig. 10) reduces the density contrast and thus the overturning strength. Beyond the Nordic Seas, however, the framework becomes less straightforward, as restratification in the Arctic coincides not with a weakening but with a strengthening of the convective cell, pointing to a distinct dynamical regime. Whether a thermal wind framework can be extended into the Arctic basin, and thereby used to connect local density changes to the broader overturning, remains an open question that could be addressed using the framework

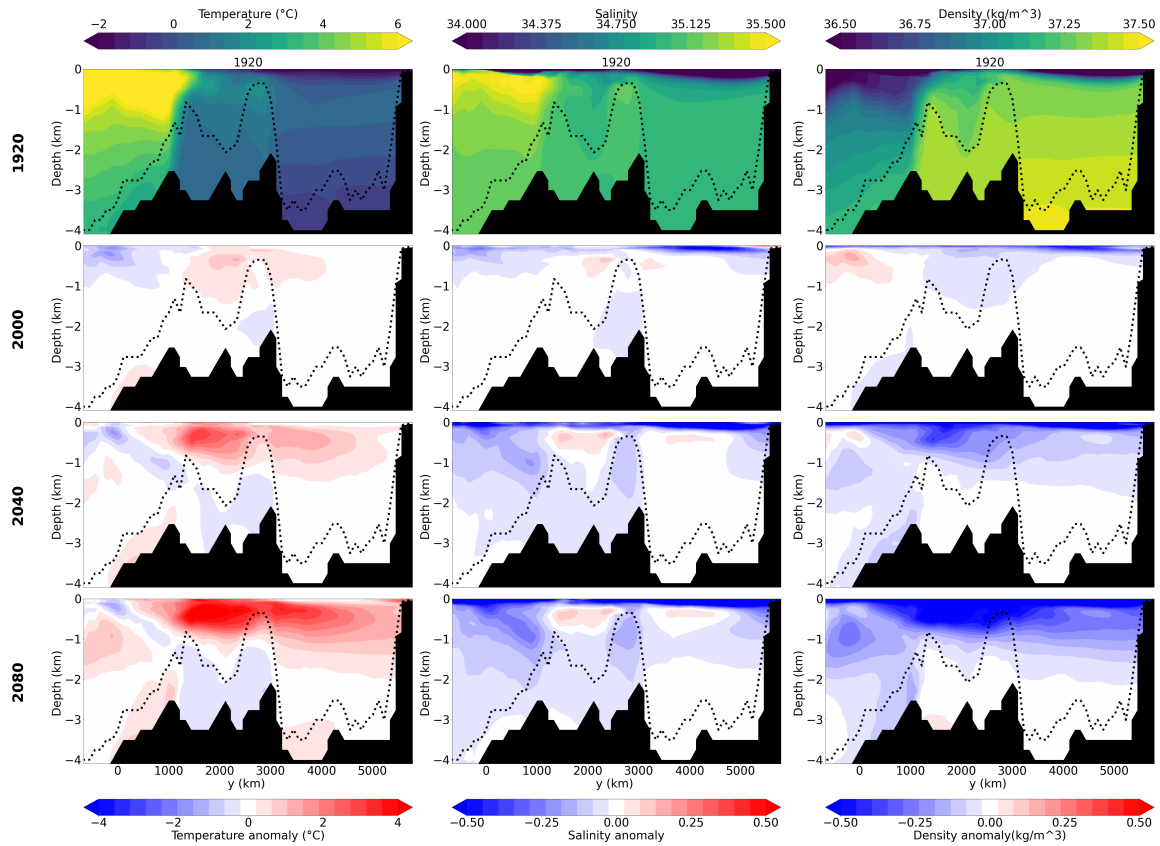


Figure 10 – Time evolution of (left) temperature, (middle) salinity, and (right) density for the historical and RCP8.5 scenarios. The top panel shows the 1920 climatology; subsequent panels show anomalies relative to 1920 for the years 2000, 2040, and 2080. The dotted black line indicates the 75th percentile of topography along each integration path, and the black region marks the deepest point along each path.

of Tessier et al. (2026).

That Arctic overturning strengthening occurs alongside restratification suggests a positive feedback loop. Enhanced mid-depth stratification limits the penetration depth of newly formed dense water, thereby shoaling the convective cell. Concurrently, warming of the upper ocean drives summer sea ice retreat. Assuming winter sea ice extent remains broadly unchanged (Fig. 15), this yields enhanced brine rejection during autumn freeze-up, intensifying convection (Bitz et al., 2006; Polyakov et al., 2020; Bretones et al., 2022). A stronger convective cell, in turn, draws more Atlantic Water into the Arctic, amplifying Atlantification, further warming the upper ocean, and enhancing mid-depth stratification — closing the feedback loop.

Using the ideal age as a proxy, we confirm increased convection in the Arctic with global warming. The bottom panels of Fig. 11 show ideal age anomalies relative to 1920, with the color of the bottom-right corner dot indicating the number of years elapsed since 1920. Values smaller than this number can therefore be interpreted as reflecting active ventilation or the advection of younger waters. Over time, water in the upper 2000m of the Arctic basin is renewed at a faster rate, becoming progressively younger, indicating both enhanced ventilation and import of younger waters from the Nordic Seas. In contrast, below 2000m, ventilation is minimal and the ideal age increases at the same rate as the calendar year. In the Nordic Seas, ideal age below 500m increases faster than the calendar year, consistent with an export of older Arctic bottom waters to this region, as expected given the intensification of the Arctic convective cell. Further south, in the North Atlantic, waters are no longer renewed at the bottom of the water column and also age at the same rate as the calendar year. Taken together, this suggests a stronger convective cell that is concentrated higher up in the water column.

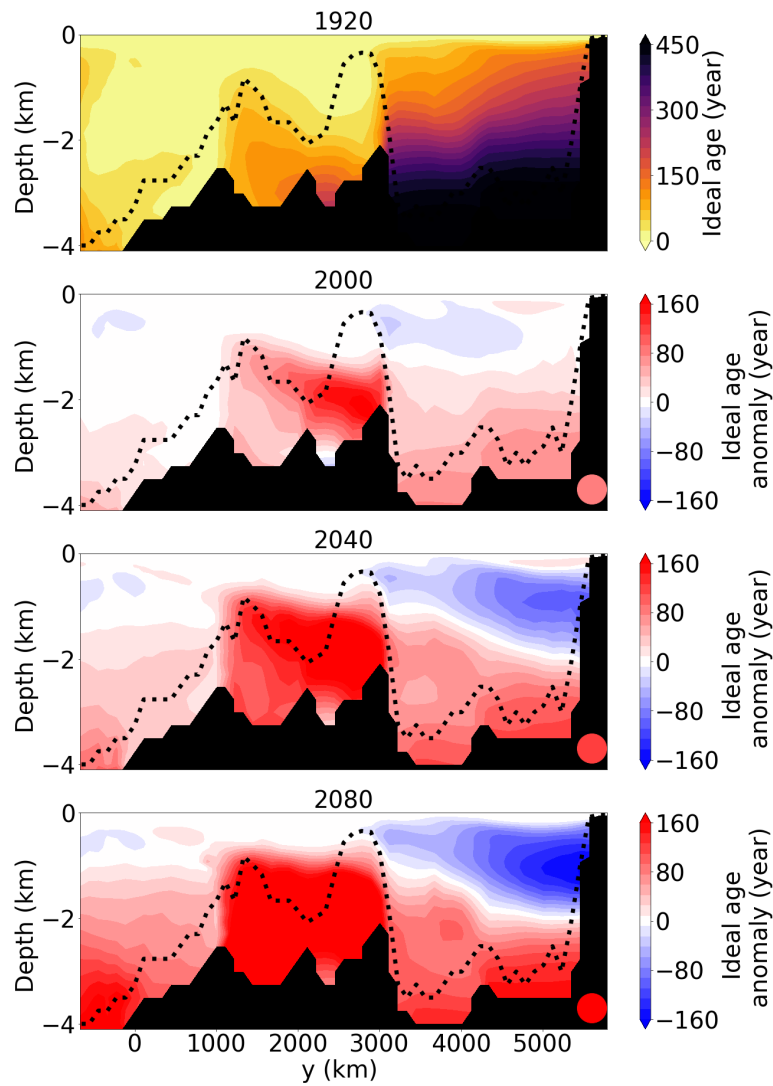


Figure 11 – Time evolution of ideal age averaged along the long axis for the historical and RCP8.5 scenarios. The top panel shows the 1920 climatology; subsequent panels show ideal age anomalies relative to 1920 for the years 2000, 2040, and 2080. The color of the dot in the bottom-right corner indicates the number of years elapsed since 1920, which approximately matches the ideal age at the bottom of the Arctic basin. The dotted black line indicates the 75th percentile of topography along each integration path, and the black region marks the deepest point along each path.

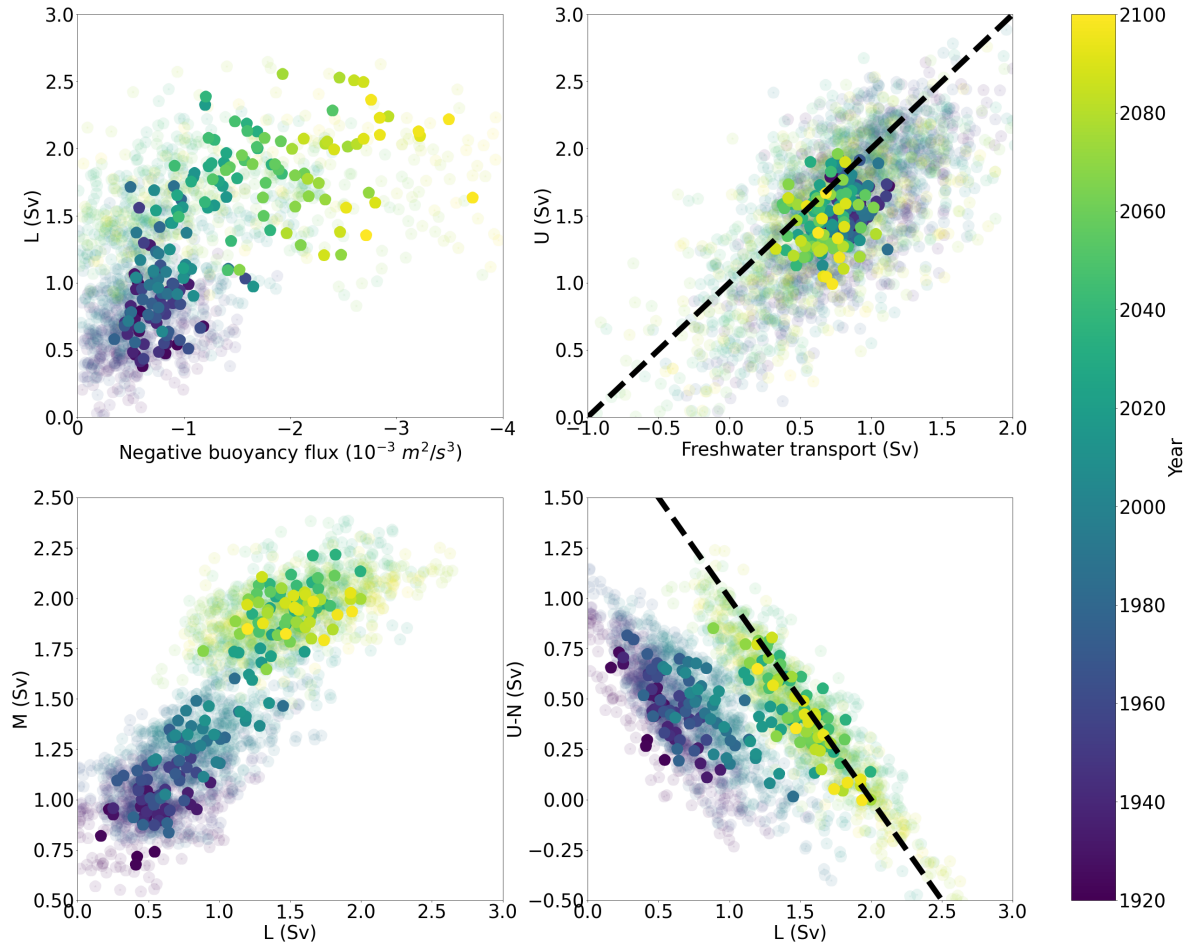


Figure 12 – Correlations between overturning branches and potential forcing mechanisms: (upper left) lower branch of the convective cell, L , against negative surface buoyancy flux; (upper right) upper branch of the estuarine cell, U , against annual surface freshwater flux (including Bering Strait inflow, precipitation, and ice melt runoff); (lower left) mid-depth transport, M , against L ; and (lower right) maximum estuarine cell intensity, $U-N$, against L . Dark dots indicate annual averages, except in the upper-left panel where December surface buoyancy fluxes are used. Light dots show monthly averages.

1.5.4 Surface fluxes

Just as the convective cell is related to deep water formation along the continental margin, the estuarine cell is related to freshwater input and diapycnal mixing (Rudels, 2010; Nummelin et al., 2016; Brown et al., 2025). With this in mind, we investigate correlations between these forcings and transport of these cells. It is convenient to characterize the two cells by three transport: U for the upper branch of the estuarine cell, M for the mid-depth flow, and L for the lower branch of the convective cell. Note that U is defined so as to include the net throughflow, which is strongly correlated with Bering transport and which we denote as N. The upper left panel of Fig. 12 shows correlations between L and negative buoyancy fluxes in December. (Details of how the buoyancy fluxes were calculated can be found in the appendix). On short time scales, no clear correlation is seen, however, over longer time scales, a positive correlation emerges. This is consistent with our view that the overall strengthening of the convective cell is locally driven by Arctic convection (Rudels, 2010; Brown et al., 2025), but that there is considerable interannual variability for which this correlation is less robust.

The upper right panel correlates U with freshwater transport (which, recall, is itself strongly correlated with N as described in the appendix). The light dots correspond to monthly averages and the dark dots to annual averages. Results show a strong correlation for the monthly values but no obvious correlation for the annual averages. This may be related to a time delay of estuarine cell's response to changes in the freshwater flux (Macdonald, 2000); however, attempts to quantify this were inconclusive because of large monthly variations in Bering transport. Also shown (lower left panel) is that M correlates well with L. Note that, again, this correlation appears differently, depending on the time scale under consideration. For example, a best fit slope to all points on the scatter plot is steeper than, say, a best fit slope through the yellow dots (corresponding to the last few decades of the simulation). Finally, the fourth panel shows U-N (which can be thought of as the baroclinic part of the estuarine cell) to be strongly anticorrelated with L, with a slope of -1 . The overall picture, then, is

that, on longer time scales, increased M implies increased L , implying long time changes in M are associated with the convective cell, consistent with the feedback loop described above. Superposed on this is variability in the partitioning of M into the upper and lower cells, with an increase in one implying a decrease in the other.

1.6 Conclusions

Using the CESM model with the historical and RCP8.5 global warming scenarios, we examined the Arctic's overturning circulation, and how it varies according to the choice of an axis along which it is defined. In particular, we considered a short axis extending from the East Siberian Sea to Northwestern Greenland and a long axis extending between Bering and Fram Straits. The long axis proved more useful for representing the double estuarine structure of the Arctic's overturning circulation, a feature that had previously been diagnosed only in basin-averaged estimates, without detailed characterization of its spatial structure. The short axis, by contrast, is better suited for analyzing the convective cell, this being because much of the deep water formation fuelling that cell occurs along the Eurasian shelves.

A strong seasonal cycle was evident in both cells. The historical simulation showed the estuarine cell to dominate in summer and be confined to the surface layer in winter. Conversely, the convective cell dominates in winter and retreats to the Nordic Seas in summertime. With global warming, the estuarine cell maintains its intensity but becomes confined to a thin layer at the surface. The convective cell, by contrast, becomes present year-round, shallows, and strengthens. As it strengthens in the Arctic, it remains stable in the Nordic Seas and weakens in the North Atlantic. This suggests that, as the AMOC weakens, it extends further north, leading to a further Atlantification of the Arctic.

Related to this Atlantification is an increase in mid-depth stratification related to a warming of upper ocean waters. This promotes summer ice melt which, in turn, leads to a larger ice free area in early winter, thus increasing air-sea heat transfer and deep formation

via brine rejection. Continuing in the loop, this increases convection, strengthens the lower cell and, with it, import of Atlantic waters.

We also attempted to find correlations between the strength of the two cells and their related local forcing mechanisms. Results were mixed, but did show that on long time scales, the outgoing transport of the convective cell (i.e., L) is well correlated with negative surface buoyancy fluxes in wintertime. The outgoing branch of the estuarine cell (defined to include the net throughflow) showed a weaker relationship with freshwater transport, with correlations emerging only on short time scales. Overall, increases in M were related to increases in L , at least on long time scales. That is, more import of Atlantic waters was associated with an increase in the convective cell. We further found an anticorrelation between annual maxima in the intensity of the convective cell and the baroclinic part of the estuarine cell. This suggests that, for a constant M , there can be variability in how the incoming Atlantic waters are partitioned into one part feeding the convective cell and another feeding the estuarine cell.

Several caveats apply to the present study. The analysis is based on a single member of a single model, and robustness should be assessed across ensemble members and model configurations. That said, the broad qualitative agreement between our diagnosed overturning and basin-scale observational and model-based estimates (e.g. Tsubouchi et al., 2024; Brown et al., 2025; Dörr et al., 2026) suggests that the main features reported here are not an artifact of this choice.

We also attempted to correlate convective cell strength with negative buoyancy fluxes, but this comparison relies on monthly averaged data, whereas the surface water mass transformations driving the convective cell likely require much higher temporal resolution to capture the episodic and spatially localized nature of brine rejection on the shelves. Linking the local cell strength to buoyancy fluxes will require a water mass transformation framework in density space along the “long axis”.

Beyond advancing our understanding of Arctic basin dynamics, a major motivation for

diagnosing the Arctic overturning is its application to the Arctic heat budget — a question not addressed in the present study. Oldenburg et al. (2024) found that Arctic warming occurs primarily through ocean heat transport at the Barents Sea Opening, driven by warmer inflowing water rather than changes in volume transport across the gateways. Our results are partially at odds with this picture as looking deeper into the Arctic we found a strengthening of the convective cell that coincides with the warming of the mid-depth Atlantic-origin water layer, suggesting that volume transport should play a role. This is, however, consistent with other diagnoses of the Arctic overturning (Bitz et al., 2006; Bretones et al., 2022). Linking convective cell strength to horizontal heat fluxes at the gateways remains an important direction for future work.

Acknowledgments

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Data statement

The CESM data used are freely available online and the code use to create the new integration paths is available on Github at https://github.com/RosalieShink/Arctic_overturning_diag.

1.7 Appendix: Methodology complement and sea ice cover

1.7.1 Integration Pathways

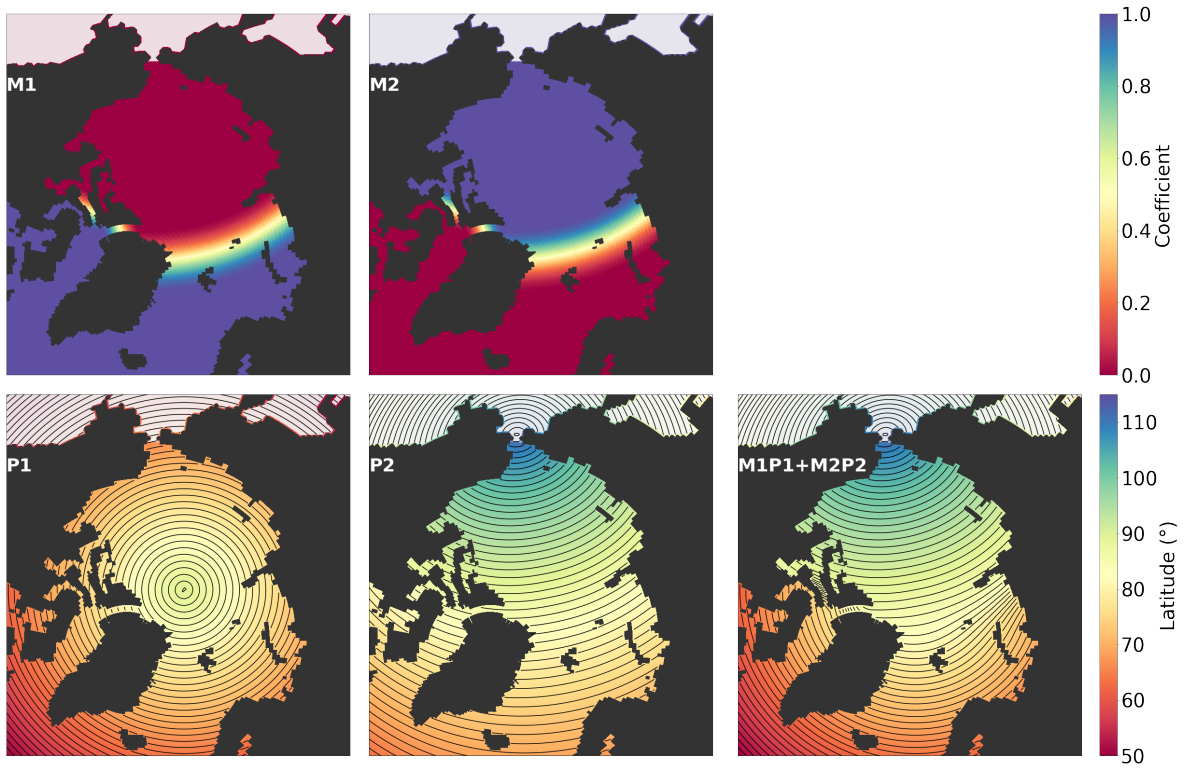


Figure 13 – Method used to create the long axis. The original latitude, P_1 , are combined with the displaced latitudes, P_2 , using a spatially dependent coefficient, M_1 and M_2 , to create a transition region. The resulting paths are $M_1P_1 + M_2P_2$.

The method used to construct the long-axis paths is illustrated in Fig. 13. Starting from the original latitude grid P_1 , a new grid P_2 is created by displacing the North Pole beyond Bering Strait using the method of Roberts et al. (2006), which employs a cosine transformation. To preserve the original paths in the North Atlantic and further south, P_1 and P_2 are blended in a transition region defined by a weighting coefficient ranging from 0 to 1, with $M_1 = 1 - M_2$ determining the relative contribution of each grid. The transition region spans latitudes 50° – 60° in P_2 , and a constant offset of 22° is added to P_2 to ensure a continuous transition. The final grid is given by $M_1P_1 + M_2P_2$. The same approach was applied to construct the short-axis projection (c.f. Fig. 14), with a transition region spanning 62° – 65° and a constant offset of 15° in P_2 .

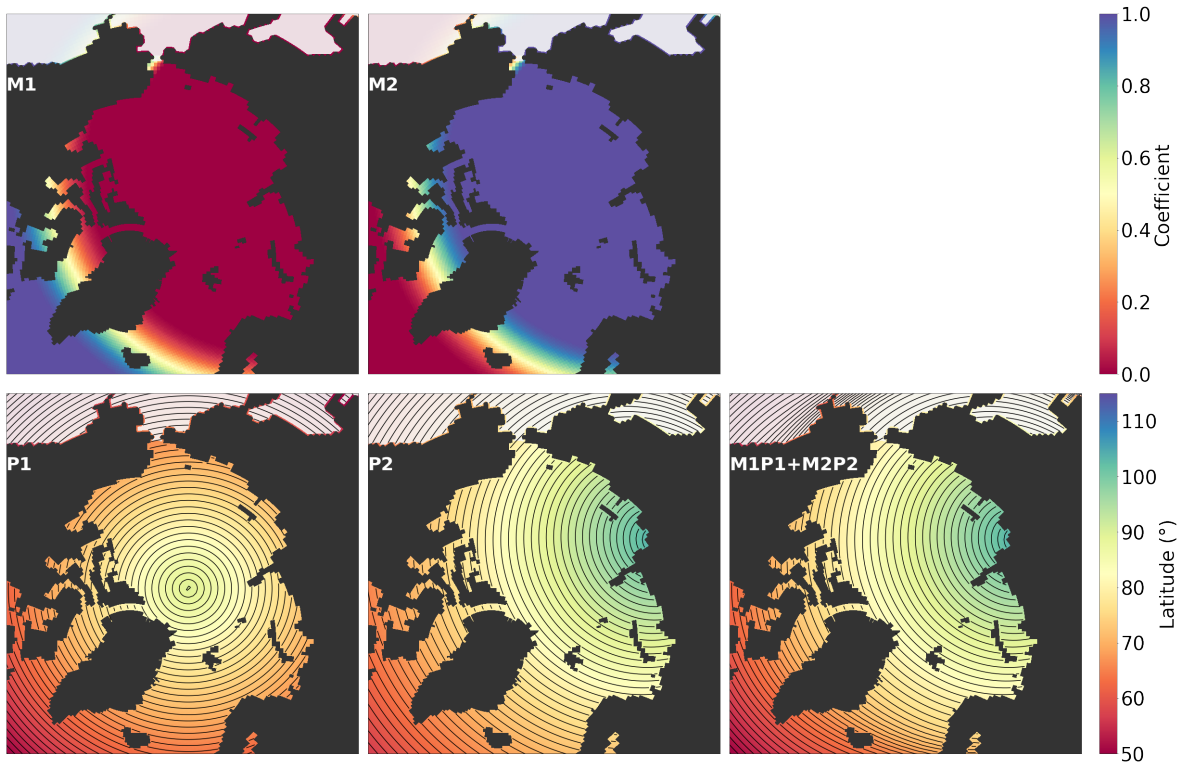


Figure 14 – Method used to create the short axis. The original latitude, P1, are combined with the displaced latitudes, P2, using a spatially dependent coefficient, M1 and M2, to create a transition region. The resulting paths are M1P1+M2P2.

1.7.2 Surface Fluxes

To calculate the buoyancy flux, we use the SHF and SFWF diagnostic of the CESM model. To compare and combine them, we use:

$$SBF = \frac{g\alpha SHF}{10\rho_{sw}cp_{sw}} + \frac{g\beta SFWF}{10^5\rho_{fw}}, \quad (1.2)$$

where g is the gravitational acceleration in cm/s^2 , α is the thermal expansion coefficient in $1/K$, SHF is the total surface heat flux including short waves in W/m^2 , ρ_{sw} is the density of seawater in g/cm^3 , cp_{sw} is the specific heat of seawater in $erg/g/K$, β is the saline expansion coefficient, SFWF is the virtual salt flux in fresh water flux formulation in $kg/m^2/s$ and ρ_{fw}

is the density of fresh water in g/cm^3 giving SBF in m/s^3 . SBF has dimensions of m^3/s^2 .

The total freshwater transport is the sum of the continental ice runoff, IOFF_F, the precipitation, PREC_F, converted in m^3/s and the transport at Bering Strait.

1.7.3 Ice cover

The temporal evolution of the sea ice cover is shown in Fig. 15

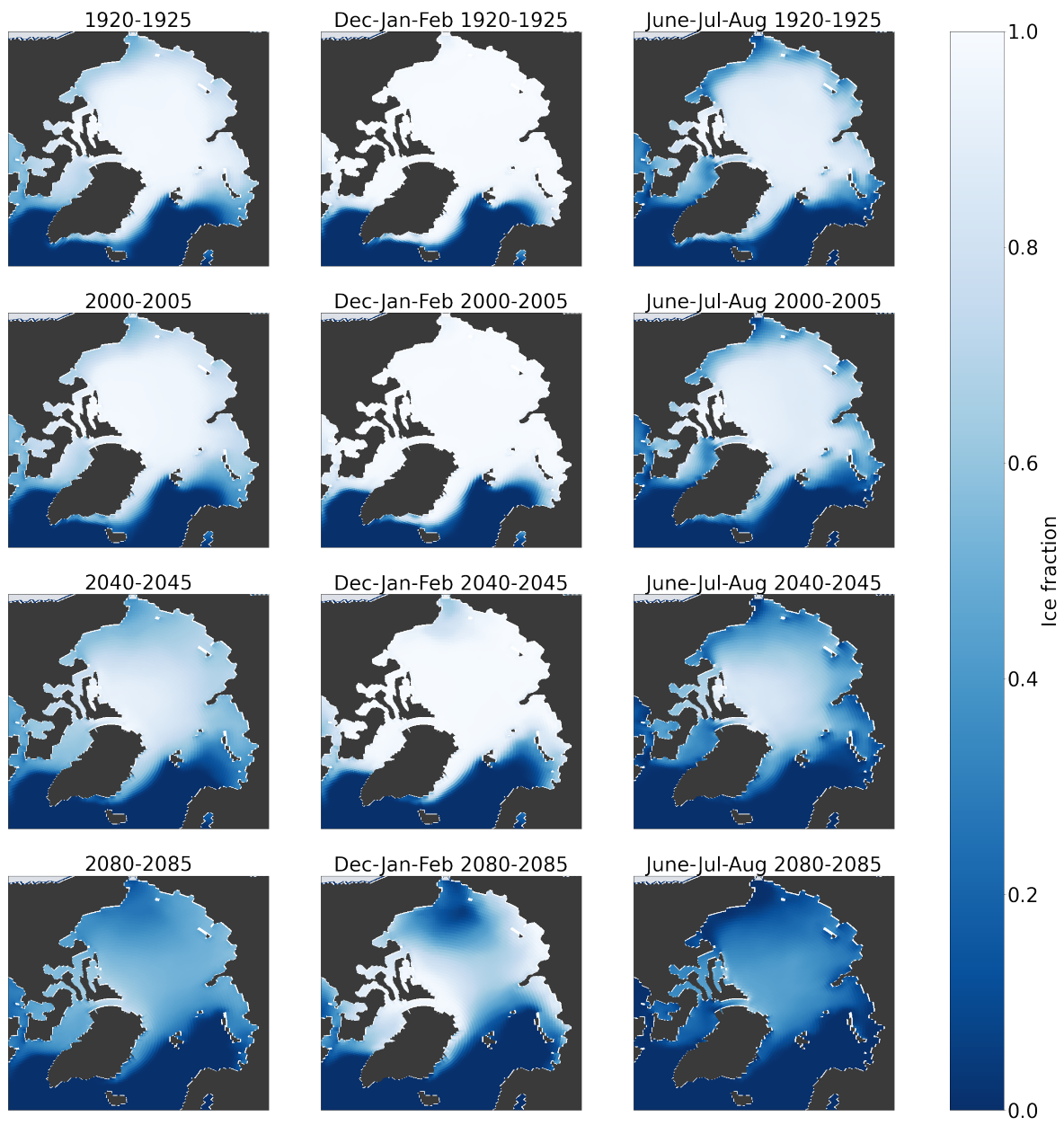


Figure 15 – Five-years averaged annual sea ice cover (left), five-years averaged winter ice cover (right) (middle) and five-years averaged summer ice cover (right).

ARTICLE 2

MÉCANISMES DE FORÇAGE DE LA CIRCULATION ESTUARIENNE DOUBLE ARCTIQUE ET LEUR INFLUENCE SUR LE BILAN DE CHALEUR

Ce article, intitulé “*Drivers of the Arctic double estuarine circulation and their control on the basin heat budget*” est en cours de préparation pour être soumis au *Journal of Physical Oceanography*. Il est le résultat du travail de moi-même, du Professeur Louis-Philippe Nadeau et de l’agent de recherche Simon Senneville. Comme première auteure, ma contribution regroupe l’élaboration de l’idée originale, la revue de littérature, le développement du modèle numérique, la réalisation des expériences numériques, l’analyse et l’interprétation de celles-ci et la rédaction de l’article. Le Professeur Louis-Philippe Nadeau, deuxième auteur, a participé à l’élaboration du modèle, à l’analyse et l’interprétation des résultats ainsi qu’à la rédaction de l’article. Enfin, Simon Senneville, a fourni un grand soutien technique en développant le modèle numérique et en réalisant des expériences numériques. Il a également participé au processus de révision de l’article.

2.1 Résumé en français

Le transport de chaleur océanique dans le bassin arctique est gouverné par sa circulation de renversement et contribue à l’amplification arctique, un mécanisme de rétroaction qui amplifie le réchauffement aux hautes latitudes, mais également plus au sud. Il est donc de plus en plus important de comprendre les impacts associés à la circulation de renversement arctique. Alors que les flux de chaleur estimés aux entrées de l’Arctique indiquent que la mer de Barents et le détroit de Fram sont responsables du plus important apport de chaleur par voie océanique, cette perspective fondée sur les frontières de l’océan Arctique offre une vision limitée de la dynamique interne de ce dernier. Comprendre cette dynamique requière de diagnostiquer la circulation de renversement arctique que l’on peut conceptualiser comme

un système à deux cellules: une cellule de type estuarienne peu profonde superposée à une cellule convective plus profonde similaire à l'AMOC. Ensemble, ces deux cellules forment la circulation estuarienne double. Nous utilisons un modèle idéalisé de l'océan arctique pour explorer les mécanismes de forçage dirigeant cette circulation ainsi que leur influence sur les flux de chaleur. Nous présentons l'espace des paramètres des régimes dynamiques de la circulation sous l'influence de flux de sel imposés. Ces flux de sel représentent le cycle saisonnier de la formation et fonte de glace de mer sur le plateau continental eurasién. Pour de faible flux de sel, le système se trouve dans un régime estuarien. Pour de forts flux de sel, un régime similaire à celui de l'AMOC s'établit. Un régime transitoire de type double estuarien se manifeste entre ces deux limites. Une forte corrélation positive émerge entre les flux de chaleur au détroit de Fram et l'intensité de la cellule convective. Cette corrélation se manifeste également dans l'anomalie du contenu thermique au-dessus de l'halocline. En revanche, la température moyennée sur toute la profondeur présente une structure plus complexe avec un refroidissement associé à la transition d'un régime estuarien à un régime double estuarien alors que les eaux denses formées sur le plateau continental remplissent graduellement les abysses. Des expériences de résilience ont démontrées que les cellules estuarienne et convective sont sensibles au flux de masse d'eaux douces entrant dans le bassin arctique à la surface et à la stratification dans les mers nordiques. La diffusivité influence majoritairement la cellule estuarienne, alors que l'anticyclone de Beaufort exerce sa plus grande influence sur la cellule de type convective.

2.2 Abstract

The Arctic overturning circulation governs ocean heat transport into the basin and contributes to Arctic Amplification, a positive feedback that accelerates warming both at high latitudes and farther south. This makes understanding its associated impacts an increasingly important area of research. While heat-flux estimates at the Arctic gateways indicate that the Barents Sea and Fram Strait deliver the largest oceanic heat input, this gate-based perspective offers limited insight into the internal dynamics of the Arctic Ocean. Understanding these dynamics requires diagnosing the Arctic overturning circulation, which can be conceptualized as a two-cell system: a shallow estuarine cell overlying a deeper convective overturning cell, together forming the so-called double estuarine circulation. Here, we use an idealized model of the Arctic Ocean to investigate the forcing mechanisms governing this circulation and their influence on heat fluxes. We map the parameter space of overturning dynamical regimes under a prescribed salt-flux forcing that represents the seasonal cycle of sea-ice growth and melt over the Eurasian shelves. The system exhibits an estuarine regime under weak salt flux, an convective regime under strong salt flux, and a transitional double estuarine regime between these two limits. A strong positive correlation emerges between heat flux through Fram Strait and the strength of the convective cell, which manifests in the heat content anomaly above the halocline. In contrast, the depth-averaged temperature exhibits a more complex structure, with a cooling associated with the transition from the estuarine to the double estuarine regime as dense shelf-formed waters gradually fill the abyss. Robustness experiments demonstrate that both the estuarine and convective cells are sensitive to surface freshwater mass flux into the Arctic and to stratification in the Nordic Seas. Diffusivity primarily influences the estuarine cell, whereas the atmospheric Beaufort High exerts its strongest control on the convective cell.

2.3 Introduction

It is now widely recognized that the Arctic plays a critical role in the global climate system (McGuire et al., 2006; Zhao et al., 2019; Dethloff et al., 2006). Under global warming, increased atmospheric and oceanic heat fluxes into the Arctic have contributed to the decline in sea-ice extent, particularly in summer (Sirevaag et al., 2011; Steele et al., 2010). The volume of multi-year sea ice has also decreased, leaving the remaining ice thinner and more susceptible to break-up (Bi et al., 2020). A more ice-free Arctic Ocean amplifies global warming through the ice–albedo feedback and enhances heat fluxes from the ocean to the atmosphere (Hall, 2004; Hudson, 2011). Central to the oceanic heat transport into the Arctic basin is the Arctic overturning circulation, which has received increasing attention over the past decades (Timmermans and Marshall, 2020; McClelland et al., 2012; Jones, 2001). Yet the processes that control the dynamics of this overturning remain poorly understood.

To assess the oceanic poleward heat transport in the Arctic, most studies have focused on heat and mass fluxes through the various gateways of the Arctic Ocean and have evaluated their relative contributions (Arthun et al., 2019; Docquier and Koenigk, 2021; Oldenburg et al., 2024). These studies generally show that the Barents Sea Opening is the largest source of oceanic heat import into the Arctic due to strong local deep water formation, which weakens stratification and facilitates the inflow of Atlantic waters. The second largest import of oceanic heat occurs through Fram Strait, another key region of deep water formation (Rudels and Carmack, 2022b; Oldenburg et al., 2024). They also found that, under global warming, the increase in heat transport into the Arctic arises primarily from the warming of Atlantic waters rather than from an increase in volume transport through the gateways (Koenigk and Brodeau, 2014), Bitz et al. (2006) and Bretones et al. (2022).

While this gateway-budget approach is valuable for quantifying the exchange between the Atlantic, the Nordic Seas, and the Arctic, it provides limited insight into the internal dynamics of the Arctic Ocean. Additional progress can be achieved by extending the Atlantic

Meridional Overturning Circulation (AMOC) into the Arctic basin. However, the polar coordinates commonly used to diagnose the AMOC are poorly suited to the geometry of the Arctic, requiring alternative coordinate systems.

Using a coordinate axis aligned from the Nordic Seas toward the Eurasian shelves, Bitz et al. (2006) and Bretones et al. (2022) showed that the Arctic overturning is dominated by a single convective cell north of Fram Strait, sustained by dense-water formation along the Eurasian shelves. In contrast, Shink et al. (in prep.) demonstrated that the diagnosed circulation is highly sensitive to the choice of coordinates. When projecting the overturning along an axis from the Nordic Seas to Bering Strait, they found a shallower estuarine cell overlying the convective cell, consistent with a double estuarine circulation. The present study focuses specifically on this circulation regime.

Few studies have directly examined how individual forcing mechanisms influence each branches of the double estuarine circulation, in part because of the difficulty of diagnosing the circulation itself. The estuarine cell is generally understood to arise from mechanisms analogous to those operating in classical estuaries: a surface salinity gradient between the relatively fresh inflow through Bering Strait and the surface outflow toward the Atlantic, the inflow of saline Atlantic waters at depth, and surface freshwater inputs (Nummelin et al., 2016; Timmermans and Marshall, 2020; Burchard et al., 2019). This horizontal salinity gradient promotes strong vertical mixing, which in turn drives the associated upwelling circulation (Walin, 1977; Burchard et al., 2019).

The convective cell appears to be forced by a combination of processes. A first key process is the surface buoyancy loss that occurs in winter, when brine rejection during sea-ice formation generates dense water (e.g., Aagaard et al., 1985; Carmack and Wassmann, 2006). Since this process occurs predominantly on the Eurasian shelves, the resulting dense waters sink to depths across the shelf break and flow out of the Arctic through Fram Strait, forming the lower branch of the convective cell (Rudels, 2009; Macdonald and Carmack, 1991). The North Atlantic also influences the Arctic through direct inflow and through its stratification,

which becomes increasingly imprinted in the Arctic as the basin undergoes atlantification (Weijer et al., 2022; Spall, 2013; Rudels and Carmack, 2022b). Enhanced stratification shoals the Atlantic Water layer, increasing its interaction with surface waters and promoting sea-ice melt (Lique and Steele, 2013; Sirevaag et al., 2011). Finally, the atmospheric Beaufort high also modulates the Arctic overturning, particularly via the Beaufort Gyre, which accumulates freshwater and strengthens the halocline, thereby insulating warm intermediate waters from the surface (Timmermans and Marshall, 2020).

Since winter surface buoyancy loss is a primary driver of the convective cell, its strength is expected to vary strongly over the seasonal cycle (Stigebrandt, 1981; Carmack and Wassmann, 2006; Shink et al., in prep.). Shink et al. shows that the seasonal variability actually projects onto both overturning cells, as changes in the convective cell feed back onto the estuarine cell. Shink et al. also confirms that the strength of the convective cell correlates with negative surface buoyancy fluxes in winter, whereas the estuarine cell exhibits a much weaker sensitivity to surface freshwater input.

Previous studies suggest that, under continued global warming and strengthening external forcing, the Arctic's double estuarine circulation may undergo a regime transition and evolve toward either a fully estuarine or a fully convective state (Haine, 2021). This possibility is at least partly consistent with trends emerging in global climate models (e.g., Bitz et al., 2006; Bretones et al., 2022; Shink et al., in prep.). Such an uncertain trajectory underscores the need for a deeper understanding of the mechanisms governing the Arctic overturning circulation in order to better anticipate its future evolution.

The goal of this study is to identify and quantify the primary drivers of the Arctic double estuarine overturning circulation. Using an idealized model of the Arctic Ocean with imposed open boundaries at the major gateways, we examine the sensitivity of the circulation to surface buoyancy fluxes, vertical diffusivity, wind stress, Bering Strait inflow, and the stratification imposed at Fram Strait. We first map the circulation regimes that emerge under a prescribed surface salt-flux forcing designed to represent the seasonal cycle of sea-ice forma-

tion and melt over the Eurasian shelves. At weak surface salt flux, the system supports a single estuarine cell, whereas at strong surface salt flux it transitions to a predominantly convective regime. Between these limits, a broad portion of parameter space exhibits a transitional double estuarine circulation. We relate the strength of the convective cell to the poleward ocean heat transport into the Arctic basin. This heat transport is expressed in the upper-ocean heat content, while the depth-averaged heat content reveals a more complex structure associated with convection-driven surface cooling.

The paper is organized as follows. Section 2 describes the integration pathways and methodology. Section 3 presents the results, and conclusions are given in Section 4.

2.4 Numerical models and experiment design

2.4.1 Model setup

Our objective is to construct the simplest configuration that nonetheless reproduces the essential structure of the double estuarine circulation. We use the hydrostatic version of the MITgcm (Marshall et al., 1997) in an idealized Arctic Ocean domain (Fig. 16). The geometry consists of a Cartesian grid covering a rectangular domain of approximately $2800 \text{ km} \times 4700 \text{ km}$, with a bowl-shaped bathymetry representing the Arctic Ocean connected to a half-bowl to the south that represents the Nordic Seas. This type of idealized geometry has been widely used in conceptual and process-oriented models of the Arctic circulation (Stigebrandt, 1981; Aagaard et al., 1985; Spall, 2013; Lique et al., 2015; Haine et al., 2015; Lambert et al., 2016). The two basins are linked by a narrow opening that mimics Fram Strait (Fig. 16). Open boundaries are specified at the northern and southern edges of the domain to represent Bering Strait and the connection to the wider Nordic Seas, respectively. The model employs a horizontal resolution of $50 \text{ km} \times 50 \text{ km}$ and vertical grid spacing that varies from 40 m at the surface to 131 m at depth.

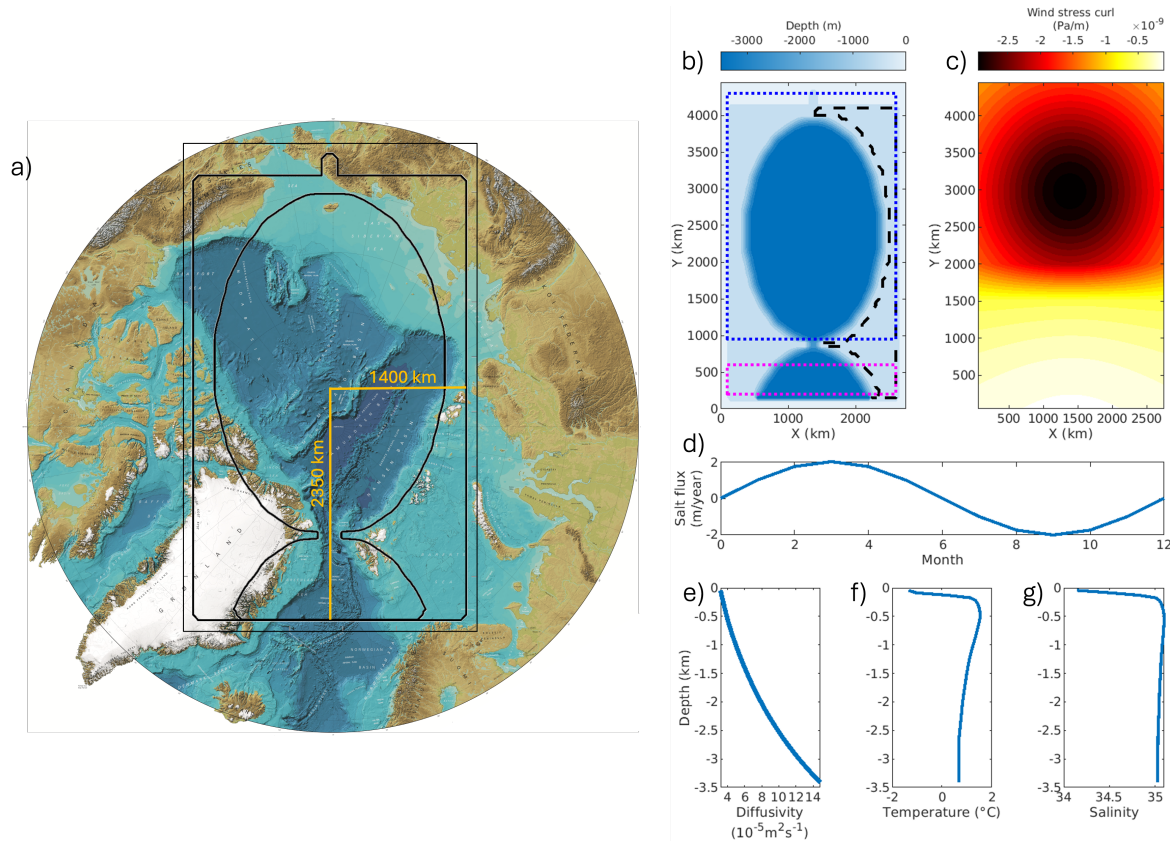


Figure 16 – GCM configuration. a) Map comparison of the Arctic bathymetry with projection parameters set to polar stereographic on the WGS 84 ellipsoid, and with true scale at 75° N taken from Jakobsson et al. (2020) and the model domain limit, 0m and 500m isobaths. b) shows the bathymetry. Over the region constrained by a dashed line is applied a seasonal salt flux following the pattern presented on d) is applied, where A the amplitude is 2 and Q is 0. The dotted blue region is restored at -1.5°C at the surface. The dotted pink line encompasses the sponge region. c) presents the curl of the wind stress applied at the surface. e) shows the vertical diffusivity profile of the idealized model. f) and g) show the imposed salinity and temperature at the southern boundary.

We use an f -plane approximation with a constant Coriolis parameter $f = 10^{-4}\text{s}^{-1}$, representative of high Arctic latitudes. Following McWilliams et al. (1990), the effect of baroclinic instability is parameterized using an isopycnal thickness diffusion with a constant eddy diffusivity of $1000\text{ m}^2\text{s}^{-1}$. Tracer advection is computed using the MITgcm second-order flux-limited scheme (Roe, 1985). The vertical diffusivity is horizontally uniform and follows the prescribed vertical profile shown in Fig. 16. Convection is treated through a convective adjustment scheme in which the vertical diffusivity is increased to $1\text{ m}^2\text{s}^{-1}$ whenever the water column is statically unstable. To represent the bottom frictional layer, a linear drag of 10^{-7}s^{-1} is applied to the bottom grid point. Laplacian viscosity is applied both horizontally and vertically, with coefficients of $5 \times 10^4\text{ m}^2\text{s}^{-1}$ and $10^{-2}\text{ m}^2\text{s}^{-1}$, respectively. Density is computed using the nonlinear equation of state of Jackett and McDougall (1995).

In order to directly control the surface buoyancy flux, we omit an explicit sea-ice model. Although this may appear drastic for an Arctic configuration, it enables us to prescribe a fixed, seasonally varying salt flux intended to represent the annual cycle of ice growth and melt over the Eurasian shelves. A sinusoidal salt-flux forcing is applied over the eastern shelves of the idealized domain (Fig. 16). We refer to the amplitude of this seasonal cycle as A , and to its net annual mean as Q . The sinusoidal form is chosen as the simplest representation of seasonality. Using output from CESM-LE, we estimate a typical net annual salt flux equivalent to 2 m/year of freshwater removal over the Eurasian shelves. This value is adopted as the baseline for our parameter-space exploration and is spatially redistributed over the shelf region. To maintain a cold halocline layer in the central basin, sea-surface temperature over the idealized Arctic is restored toward -1.5°C (Fig. 16).

This forcing strategy neglects atmosphere–ice–ocean feedbacks; incorporating explicit sea-ice dynamics and thermodynamics is left for future work. Nevertheless, the results below demonstrate that these feedbacks are not required for the emergence of a double estuarine circulation, which is the primary focus of this study.

At the northern open boundary, representing Bering Strait, we impose a prescribed in-

flow (1 Sv in the reference case) with uniform temperature of -1.5°C and salinity of 30. At the southern open boundary, representing the connection to the wider Nordic Seas, temperature and salinity profiles are imposed based on horizontally averaged CESM-LE historical output over a broad region of the Nordic Seas (Kay et al., 2015). The normal velocities at this boundary are adjusted to ensure zero net volume transport across the full domain, but are otherwise unconstrained. To avoid generating artificial thermal-wind shear at the southern boundary, the imposed temperature and salinity profiles are taken to be horizontally homogeneous. A sponge layer with a northward-decreasing *tanh* profile is applied adjacent to this boundary to smooth the transition between the imposed stratification and that of the Arctic interior.

The effect of the ice–ocean mechanical coupling associated with the atmospheric Beaufort High is represented through an anticyclonic, azimuthal surface stress field centered over the northern portion of the Arctic basin (Fig. 16). Its amplitude is chosen to reproduce a barotropic gyre transport of approximately 1 Sv, consistent with estimates of the Beaufort Gyre circulation.

All simulations were integrated for a minimum of 1000 years. Experiments that had not reached equilibrium by that time were extended for an additional 1000 years. We conducted a suite of sensitivity experiments in which we varied the surface buoyancy forcing, diapycnal diffusivity, surface mechanical stress, Bering Strait inflow, and the imposed southern stratification in order to assess their influence on the overturning circulation. For the surface buoyancy forcing, we varied both the net annual salt flux, Q , and the amplitude, A , of its imposed seasonal cycle. Wind stress and diffusivity were perturbed by multiplying their baseline amplitudes by prescribed scaling factors. At Bering Strait, we independently varied the mass flux while adjusting the salinity in order to hold a fixed salt flux, and conversely varied the salt flux while holding a fixed mass flux. Finally, to diagnose the influence of horizontal salinity gradients and surface stratification, we perturbed the prescribed salinity at the southern boundary to achieve a targeted modification of the stratification. Details of each

sensitivity experiment are discussed where relevant below.

2.4.2 Diagnostics

The overturning transport is computed by vertically integrating meridional transport below a given depth. This meridional transport is in turn computed by zonally integrating the meridional velocity. Because the model uses a Cartesian grid aligned with the north–south axis of the domain, this definition of the overturning streamfunction is directly analogous to the axis used in Shink et al. (in prep.) and therefore should capture both branches of the double estuarine circulation. Using the residual meridional velocity v_{res} , the overturning streamfunction Ψ is defined as

$$\Psi(y, z, t) = -\frac{1}{(t_2 - t_1)} \int_{t_1}^{t_2} \int_{X_E}^{X_W} \int_{-H}^z v_{res}(x, y, z', t') dz' dx dt', \quad (2.1)$$

where $v_{res} = v + v_{GM}$ is the sum of the resolved Eulerian velocity v and the eddy-induced velocity v_{GM} (parameterized with the McWilliams et al. (1990) closure), t_1 and t_2 is the averaging period, X_E and X_W mark the eastern and western limits of the basin, H is the total ocean depth. The streamfunction is integrated upward from the ocean floor, with a zero boundary condition at the bottom. With this convention, the barotropic throughflow across the Arctic basin appears directly as the surface value of Ψ .

Note that the residual streamfunction may alternatively be defined as the vertically integrated meridional transport below a given density surface rather than below a fixed depth. Because the mean isopycnal slopes in our domain are typically small, the two definitions yield overturning patterns that are similar (not shown). For this reason, and for simplicity of interpretation, we retain the depth-based overturning streamfunction in this study.

In the double estuarine regime, the overturning circulation is organized into two vertically stacked, counter-rotating cells. This structure naturally defines three vertically separated transport branches: (i) U , the upper outflow branch of the surface estuarine cell, which

exports surface freshwater southward, (ii) M , the intermediate inflow of Atlantic-origin water, comprising the combined lower branch of the estuarine cell and the upper branch of the convective cell, and (iii) L , the deep outflow branch of the convective cell, which exports dense waters southward at depth. The three transport branches introduced above can then be expressed from the streamfunction as

$$\begin{aligned} U &= \Psi_{surface} - \Psi_{min}, \\ M &= \Psi_{max} - \Psi_{min}, \\ L &= \Psi_{max} - \Psi_{bottom}, \end{aligned} \tag{2.2}$$

where $\Psi_{surface}$ is as the barotropic throughflow at the surface, Ψ_{min} is the (negative) extremum associated with the upper estuarine cell, Ψ_{max} is the positive maximum associated with the convective cell, and $\Psi_{bottom} = 0$ is the value of the streamfunction at the ocean floor. All components are positive by definition. In practice, we use the streamfunction at $y = 2450$ km to evaluate each of the above contributions. We excluded the top 110 m to disregard the Bering Strait throughflow. The intensity of the estuarine and convective cells are then evaluated using the values of U and L respectively.

Although the above transport branches are defined with the double estuarine circulation in mind, the same definitions remain applicable in regimes where only a single overturning cell is present. We anticipate three primary circulation regimes: (i) an estuarine dominated regime, (ii) a double estuarine regime, and (iii) a convectively dominated regime. To characterize these regimes quantitatively, we introduce the nondimensional index

$$\Psi_{regime} = \frac{L - U}{L + U}. \tag{2.3}$$

A value of $\Psi_{regime} = 1$ corresponds to a purely convective regime, $\Psi_{regime} = -1$ to a purely estuarine regime, and intermediate values indicate a double estuarine circulation, with $\Psi_{regime} = 0$ corresponding to two overturning cells of equal strength.

2.5 Results

2.5.1 Parameter space under prescribed salt-flux forcing

Since wintertime surface buoyancy loss is widely recognized as a key forcing of the convective cell (e.g., Bitz et al., 2006; Bretones et al., 2022), we explore the parameter space defined by the net salt flux, Q , and the amplitude of its seasonal cycle, A , which together encapsulate idealized ice thermodynamics over the Eurasian shelves. Fig. 17 summarizes the response of the overturning circulation across the full (A, Q) parameter space, showing the upper outflow branch U , the deep outflow branch L , and the circulation regime index Ψ_{regime} . Results diagnosed from long (50-year) time-averaged fields are shown in the upper row, whereas those obtained from instantaneous (monthly-mean) fields are shown in the lower row.

To further illustrate the circulation regimes and their transitions, we present a representative subset of solutions diagnosed from 50-year averages of the overturning streamfunction (Fig. 18), along with time series of the overturning circulation at a fixed latitude ($y = 2450$ km) based on monthly means (Fig. 19). (The full parameter-space dependence of these diagnostics is provided in Appendix A.)

Starting from the steady-state limit $A = 0$, for which the instantaneous and long time-averaged solutions are identical, the circulation undergoes a clear transition as the net salt flux Q increases (see $A = 0$ in Fig. 17, 18 and 19). At weak or negative Q , the system resides in a purely estuarine regime. As Q increases, a narrow window of double estuarine circulation emerges, before the system transitions to a convectively dominated regime at sufficiently large positive salt flux. Accompanying this transition, the stratification beneath the pycnocline in the Arctic basin increases steadily, as indicated by the black isopycnals in Fig. 18. As expected from the experimental design, the strength of the convective cell increases smoothly with Q in the convective regime. In contrast, the estuarine cell exhibits only a weak sensitivity

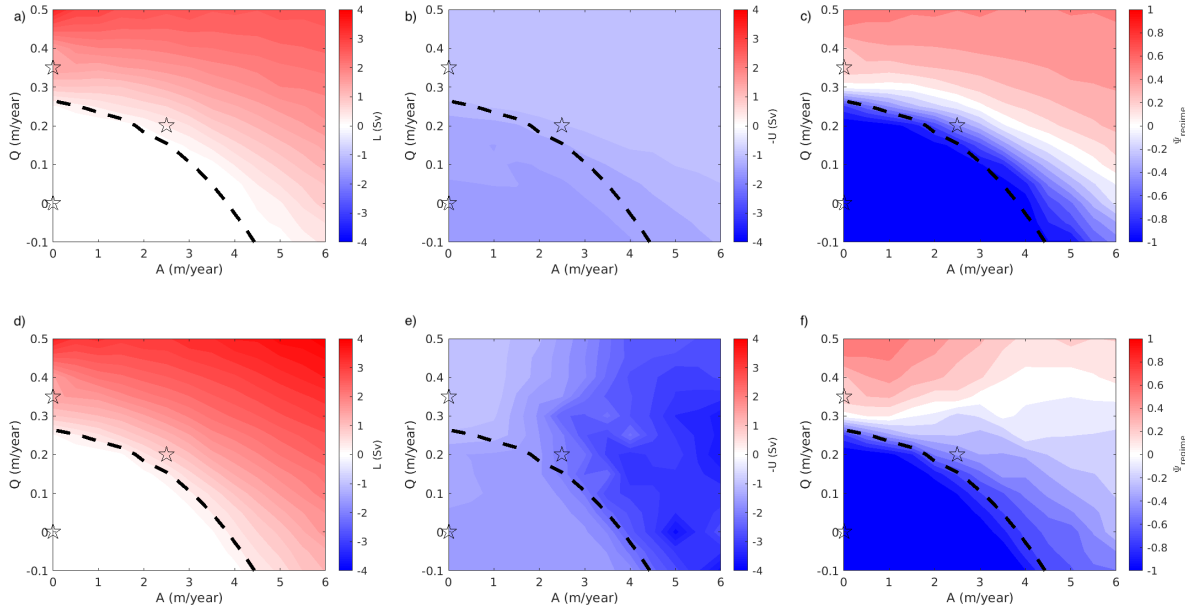


Figure 17 – Parameter space of overturning dynamical regimes under a prescribed salt-flux forcing. (left) Intensity of the convective cell, (center) intensity of the estuarine cell, and (right) circulation type Ψ_{regime} for the full A - Q parameter space. a), b) and c) were created using long time-averaged fields and d), e) and f) were created using instantaneous fields. The black dashed line indicate the convection threshold, Q_{conv} . The stars identify the reference estuarine ($A = 0$ and $Q = 0$), double estuarine ($A = 2.5$ and $Q = 0.2$) and convectively dominant double estuarine ($A = 0$ and $Q = 0.35$) reference circulations used in later analysis.

to Q within the estuarine-dominated regime. The rapid transition between the estuarine and double estuarine regimes occurs when the imposed salt flux exceeds a critical threshold, Q_{conv} , associated with the onset of deep convection. This transition is linked to the formation of a convective cell at the middle of the basin as identified by the dashed black lines in Fig. 17.

As the amplitude of the seasonal cycle, A , increases, the threshold Q_{conv} is progressively shifted toward smaller values of Q , and the range of parameters supporting a double estuarine regime gradually widens (Fig. 17), indicating a change in the underlying dynamics.

Considering first the long time-averaged solutions, increasing A at fixed Q produces an effect similar to increasing Q at fixed A (upper panels of Fig. 17). In both cases, the

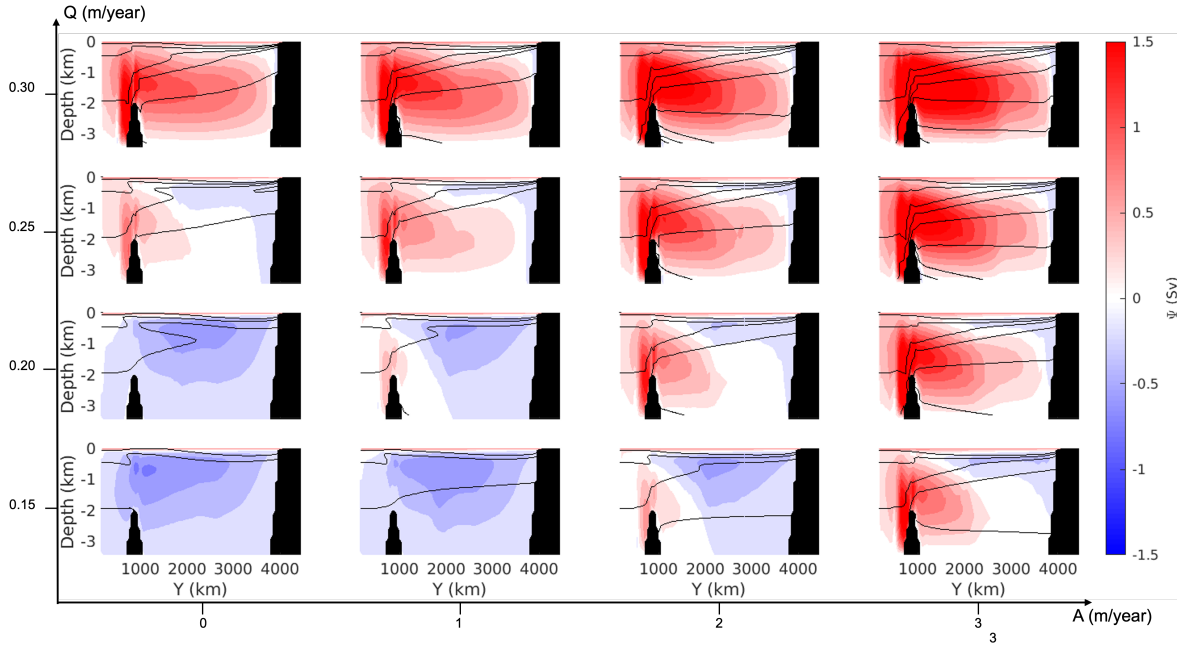


Figure 18 – 50 years averaged overturning circulation obtained for various amplitude (A) and average (Q) of the seasonal cycle of the surface salt flux. The black solid lines are the isopycnals of σ_2 ranging from 37.25 to 37.48 on a log scale.

overturning circulation transitions from an estuarine-dominated to a convectively dominated regime, accompanied by increasing stratification beneath the pycnocline as A increases. This behavior arises because increasing A amplifies the wintertime salt flux, thereby enhancing dense-water formation on the shelves and favoring the development of the convective cell. As a result, the strength of the convective cell remains nearly constant along curves roughly parallel to the convection onset threshold Q_{conv} . Since convection is directly driven by the positive (wintertime) component of the surface salt flux, we find that these curves (along which L remains nearly constant) empirically correspond to the 55th percentile of the positive part of the imposed salt flux in the (A, Q) parameter space.

Examining the time-dependent circulation reveals a different perspective. While the intensity of the convective cell is not substantially altered by temporal variability (compare L in

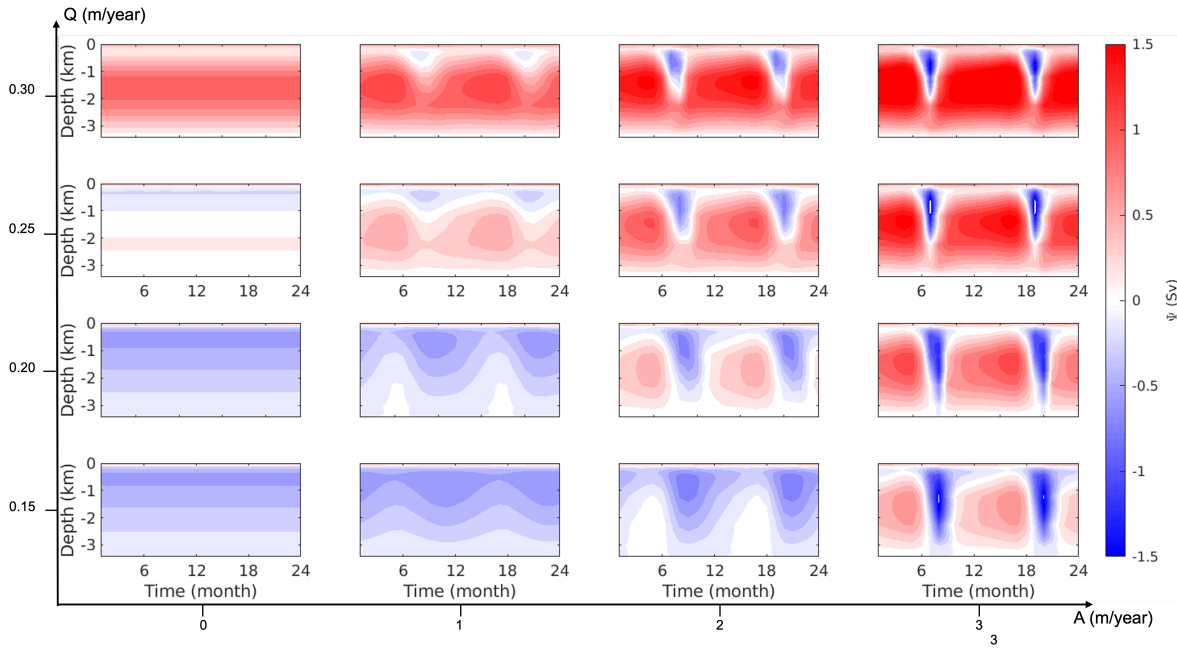


Figure 19 – Time evolution of the overturning circulation obtained for various amplitude (A) and average (Q) of the seasonal cycle of the surface salt flux. Isopycnals are omitted since their variability is negligible at monthly time scales.

Fig. 17 a and d), the estuarine cell exhibits a pronounced increase with A in the instantaneous diagnostics (see U in Fig. 17e). The strengthening of the estuarine circulation thus appears to arise through a more indirect mechanism than that of the convective cell. By reducing the summertime salt flux, larger values of A promote the re-establishment of a local halocline on the shelves; however, whether this process directly enhances the estuarine cell remains unclear. Time series of the overturning circulation show that, once the convection threshold is exceeded, the estuarine cell persists over a much smaller fraction of the annual cycle than the convective cell. This behavior suggests that the apparent strengthening of the estuarine cell is more likely associated with internal circulation dynamics, rather than being a direct response to the imposed surface buoyancy flux (Fig. 19).

To further explore the seasonal evolution of the double estuarine circulation, we exam-

ine the phase space of the overturning circulation in Fig. 20. The two transport components, the upper branch U and the lower branch L , are plotted as functions of the surface salt flux with a two-month lag, for a representative subset of simulations across the parameter space.

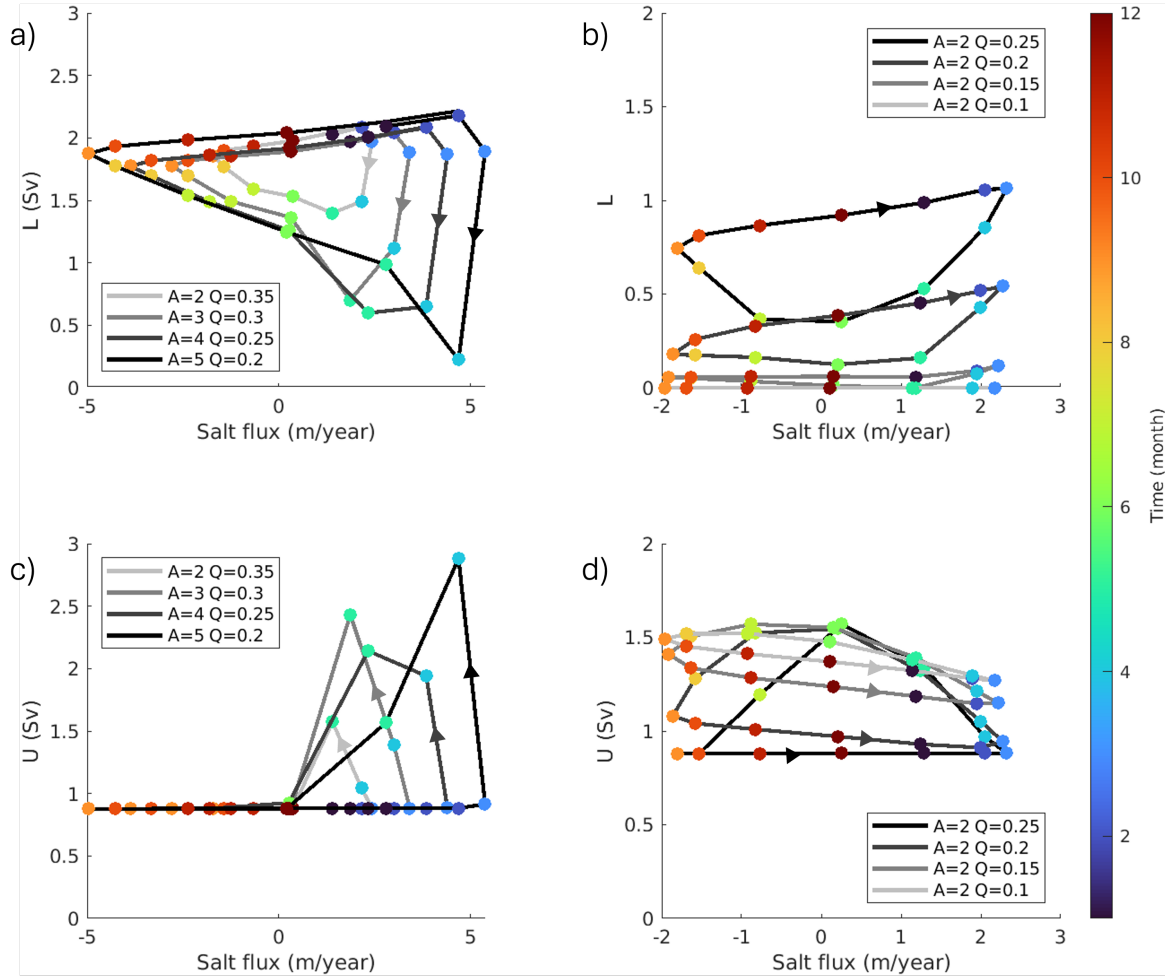


Figure 20 – Phase space of the overturning circulation. Seasonal cycle of the intensity of the estuarine cell, U , and the convective cell, L , according to the surface salt flux on the Eurasian shelves. a) and c) present double estuarine circulation for which the 55th percentile of the positive salt flux is constant. b) and d) show the double estuarine circulation for which the amplitude of the seasonal cycle is constant. Both L and U lag the salt flux by two months.

Double estuarine regimes for which $\Psi_{\text{regime}} \approx 0$ exhibit a common qualitative seasonal cycle in phase space. During winter, L increases steadily with increasing surface salt flux

along an upper branch of the trajectory. In spring, a rapid weakening of L coincides with a marked amplification of U . During summer, L gradually recovers, even while the surface salt flux remains negative.

As noted above, L remains nearly constant along curves corresponding to the 55th percentile of the positive (wintertime) salt flux in the (A, Q) parameter space, reflecting the direct control of shelf convection by the peak buoyancy loss. Solutions lying along such curves are shown in the left panels of Fig. 20. Along these trajectories, all simulations collapse onto a common upper branch for L , governed by the increase in surface salt flux. As the salt flux weakens in spring, the convective cell collapses, simultaneously with a brief but pronounced strengthening of the estuarine cell. The convective cell subsequently recovers as the estuarine cell weakens.

The role of A in amplifying the seasonal cycle is clearly illustrated in the left panels of Fig. 20. In contrast, the right panels isolate the effect of varying the mean salt flux Q . For these simulations, the estuarine cell reaches a comparable minimum strength across cases, while the intensity of the convective cell varies substantially. This behavior suggests that beyond the critical threshold for convection, the amplitude of the peak estuarine cell is primarily controlled by A (see also Fig. 17e).

2.5.2 Heat budget of the idealized Arctic Basin

Having established the possible regimes of the Arctic overturning circulation, we now examine the associated oceanic heat fluxes entering the basin and the resulting mean temperature of the Arctic Ocean (Fig. 21).

Once all experiments have been integrated to statistical equilibrium, the basin averaged temperature tendency necessarily vanishes. The heat budget of the idealized Arctic domain is therefore closed through a three-way balance between the net surface heat flux and the heat transports through the two open boundaries, $\partial_t \langle \Theta \rangle_A = Q_{\text{Surf}} + Q_{\text{Fram}} + Q_{\text{Bering}} = 0$, where $\langle \Theta \rangle_A$

denotes the average temperature of the Arctic basin.

For the analysis below, we examine anomalies relative to a reference simulation, identified by a star, located near the center of the parameter space and characterized by a double estuarine regime with $\Psi_{regime} \sim 0$ ($A=2.5$ and $Q=0.2$). Because the heat flux prescribed at Bering Strait is identical across all simulations in the (A, Q) parameter space, its anomaly is identically zero, $\Delta Q_{Bering} = 0$. Consequently, closure of the heat budget implies that anomalies in surface and Fram Strait heat fluxes must balance, i.e. $\Delta Q_{Surf} = -\Delta Q_{Fram}$.

The amplitude of the heat flux anomaly entering the Arctic through Fram Strait, ΔQ_{Fram} , shown in Fig. 21a, shows a structure that is qualitatively similar to the strength of the convective overturning branch (Fig. 17a, d). This correspondence suggests a strong correlation between L and ΔQ_{Fram} , a relationship that is quantitatively confirmed in Fig. 22a. In the estuarine regime below the convection threshold, however, ΔQ_{Fram} remains essentially constant. This does not necessarily imply an absence of correlation between the Fram Strait heat flux and the strength of the estuarine cell (Fig. 22b), because the mass transport U is itself nearly invariant across this regime (Fig. 17b, e).

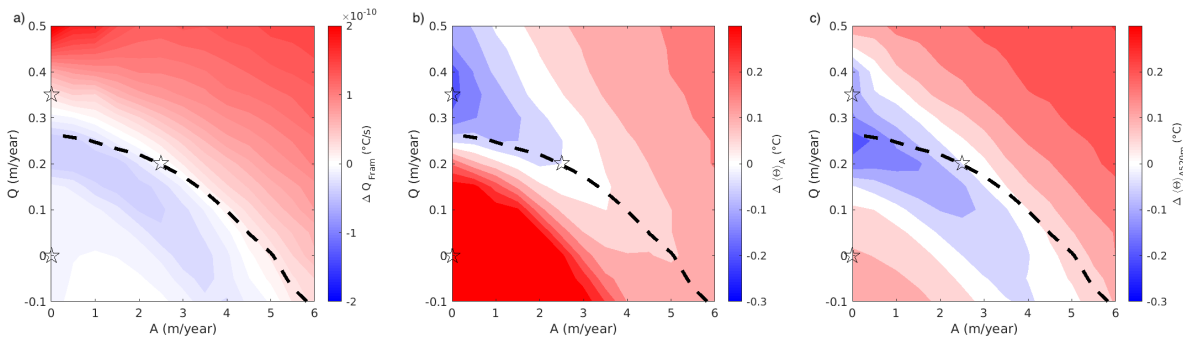


Figure 21 – Heat budget of the Arctic Basin for the (A, Q) parameter space. a), Heat flux anomaly through Fram Strait, ΔQ_{Fram} , b), anomaly of the average temperature of the Arctic basin over the full water column $\Delta \langle \Theta \rangle_A$, and c) anomaly of the average temperature of the Arctic basin above 520m (typical depth of Ψ_{min}), $\Delta \langle \Theta \rangle_{A200m}$. In all three cases, the anomaly is taken from a reference double estuarine circulation for which $A = 2.5$ m/year and $Q = 0.2$ m/year. The black dashed line is the convective threshold.

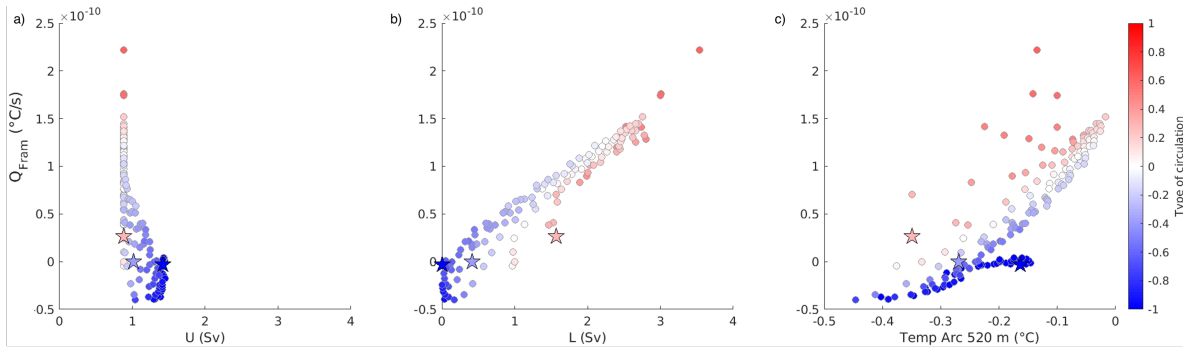


Figure 22 – Correlation between the heat flux anomaly with reference temperature -1.5°C through Fram Strait ($\Delta\langle\Theta\rangle_A$) and, a), the intensity of the estuarine cell, b), the intensity of the convective cell and c), the average temperature over the upper 520m of the arctic basin. Each point refers to a circulation in the A-Q space and their color identifies to the circulation type for long time-average fields. The stars identify the reference estuarine ($A = 0$ and $Q = 0$), double estuarine ($A = 2.5$ and $Q = 0.2$) and convectively dominant double estuarine ($A = 0$ and $Q = 0.35$) reference circulations.

Although the heat budget ensures that the system is closed at equilibrium, it does not determine the basin's equilibrium heat content, $\langle\Theta\rangle_A$, which is set by the internal dynamics of the domain. As a result, the equilibrium heat content anomaly, $\Delta\langle\Theta\rangle_A$ (Fig. 21b), bears little qualitative resemblance to ΔQ_{Fram} . Instead, $\Delta\langle\Theta\rangle_A$ is characterized by a pronounced negative anomaly aligned with the onset of the double estuarine regime near Q_{conv} . This negative feature gradually weakens as the seasonal cycle amplitude A increases, and the anomaly eventually becomes positive toward large A and Q .

To interpret this initial negative anomaly and the subsequent transition to positive heat content at higher A and Q , it is informative to examine latitude–depth sections of the zonally averaged temperature (Fig. 23). In the purely estuarine regime, the imposed temperature at the southern boundary penetrates the Arctic interior with little modification: relatively warm Atlantic origin waters fill the deep basin beneath a shallow, cold halocline. This behavior is also evident in the temperature–salinity (TS) diagram for a representative estuarine solution (Fig. 24a), where most points remain tightly clustered around the southern boundary TS

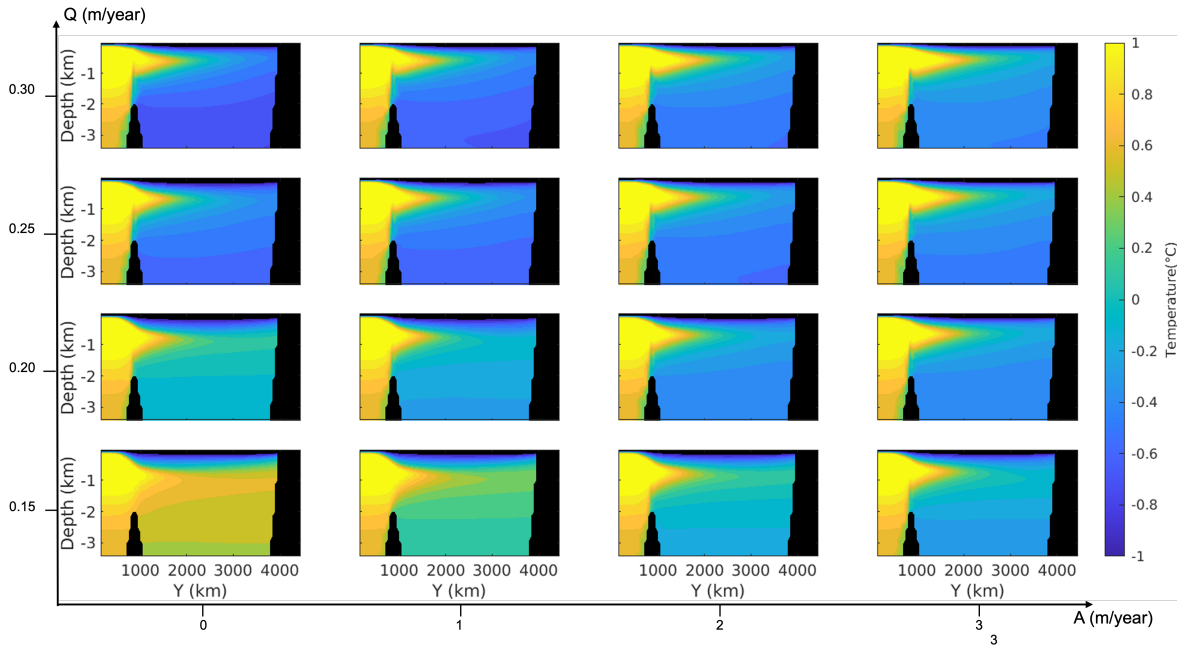


Figure 23 – As Fig.3 but for zonally averaged temperature

profile (black line).

Once the convection threshold is crossed, this deep warm Atlantic layer is rapidly replaced by dense, cold waters formed over the shelves through wintertime salt flux forcing (Fig. 23). The negative anomaly in $\Delta\langle\Theta\rangle_A$ just above Q_{conv} is therefore associated with this first episode of shelf driven ventilation, during which newly formed cold water floods the interior. The temperature–salinity (TS) diagram for a representative double estuarine solution in the center of this negative-anomaly region further illustrates this process (Fig. 24b): it shows the emergence of new, dense shelf-formed water masses as well as the abyssal branch arising from their mixing with ambient waters as they cascade across the shelf break. The emergence of convection shifts the entire TS distribution toward saltier and colder values due to the combined effect of surface salt flux and cooling.

This abyssal cooling also generates a sharp middepth temperature inversion below the

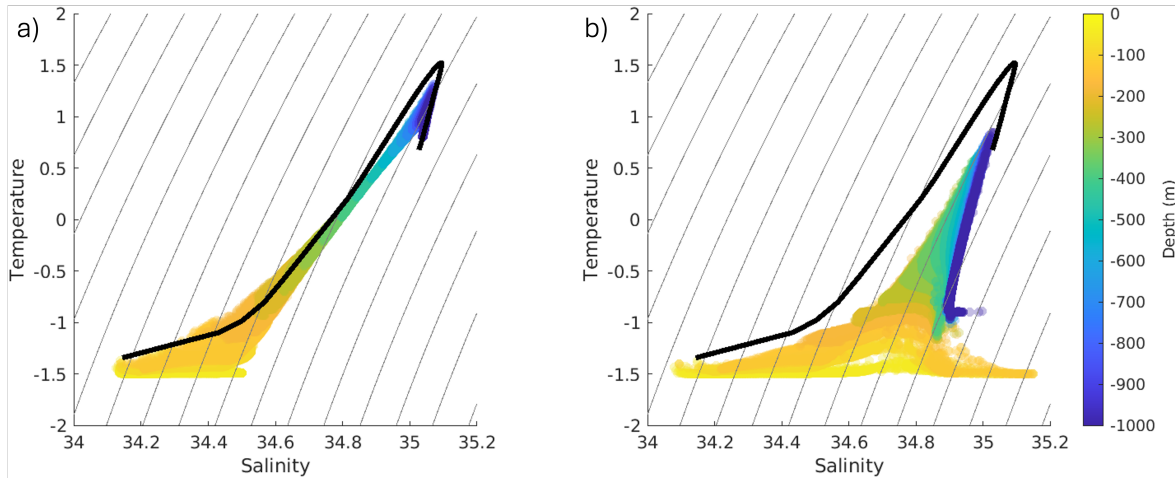


Figure 24 – Temperature - Salinity diagram for the reference estuarine circulation, left, and the convectively dominant double estuarine circulation, right in the arctic basin. The color of the points represents the depth, the black points are the imposed conditions at the southern boundary and the black lines are isopycnals of σ_2

halocline, qualitatively similar to that observed in the real Arctic Ocean (Fig. 23). As A and Q continue to increase, progressively warmer Atlantic origin waters are imported into the Arctic at middepths beneath the halocline. This inflow strengthens the temperature inversion, shoals the halocline, and ultimately produces a positive heat content anomaly in $\Delta\langle\Theta\rangle_A$ above Q_{conv} (Fig. 21b).

This interpretation is supported by the heat content of the upper 520 m halocline layer (Fig. 21c). From the perspective of the surface halocline, the heat content increases monotonically once Q_{conv} is exceeded. The negative anomaly in $\Delta\langle\Theta\rangle_{A520m}$ instead arises within the estuarine regime, where shallow shelf convection provides a modest cooling of the halocline prior to the onset of deep ventilation.

2.5.3 Robustness to other forcing mechanisms

Thus far, our analysis has focused on the role of surface salt flux in establishing the arctic overturning circulation regime and its impact on basin-wide heat transport and equilibrium temperature. However, as noted in the introduction, several other forcing mechanisms are also known to influence the arctic overturning circulation.

Since the surface salt flux is controlled by two parameters, it is convenient to combine them into a single metric to facilitate comparison with other forcing mechanisms. To this end, we use the 55th percentile of the positive part of the imposed salt flux, which was found empirically to provide a robust measure of the intensity of the convective cell and the associated circulation regime. A double estuarine circulation ($A = 2.5$, $Q = 0.2$) with $\Psi_{regime} \sim 0$ is used as a reference. The 55th percentile of the salt flux for each circulation in the A - Q parameter space is normalized by the corresponding value for this reference case. The surface salt flux in each simulation can therefore be expressed as a multiplicative factor relative to the reference double estuarine circulation.

We then examine the individual effects of Bering Strait mass and salt fluxes, vertical diffusion, southern boundary stratification, and wind stress. Three reference circulations are selected: an estuarine regime ($A = 0$, $Q = 0$), a double estuarine regime ($A = 2.5$, $Q = 0.2$), and a convectively dominated double estuarine regime ($A = 0$, $Q = 0.35$). Starting from each reference state, the corresponding forcing mechanisms are varied independently by applying a prescribed multiplicative factor. Fig. 25 highlights the forcing mechanisms that exert the strongest influence on both the estuarine and convective cells of the circulation.

Surface buoyancy forcing remains the dominant control on the convective cell, consistent with our initial hypothesis. It also exerts a strong influence on the strength of the estuarine cell once the threshold value Q_{conv} is exceeded. Mass inflow through Bering Strait plays an additional key role. It increases the mass flux which rapidly weakens the convective cell and can ultimately suppress it entirely, effectively transitioning the system to a purely estuarine

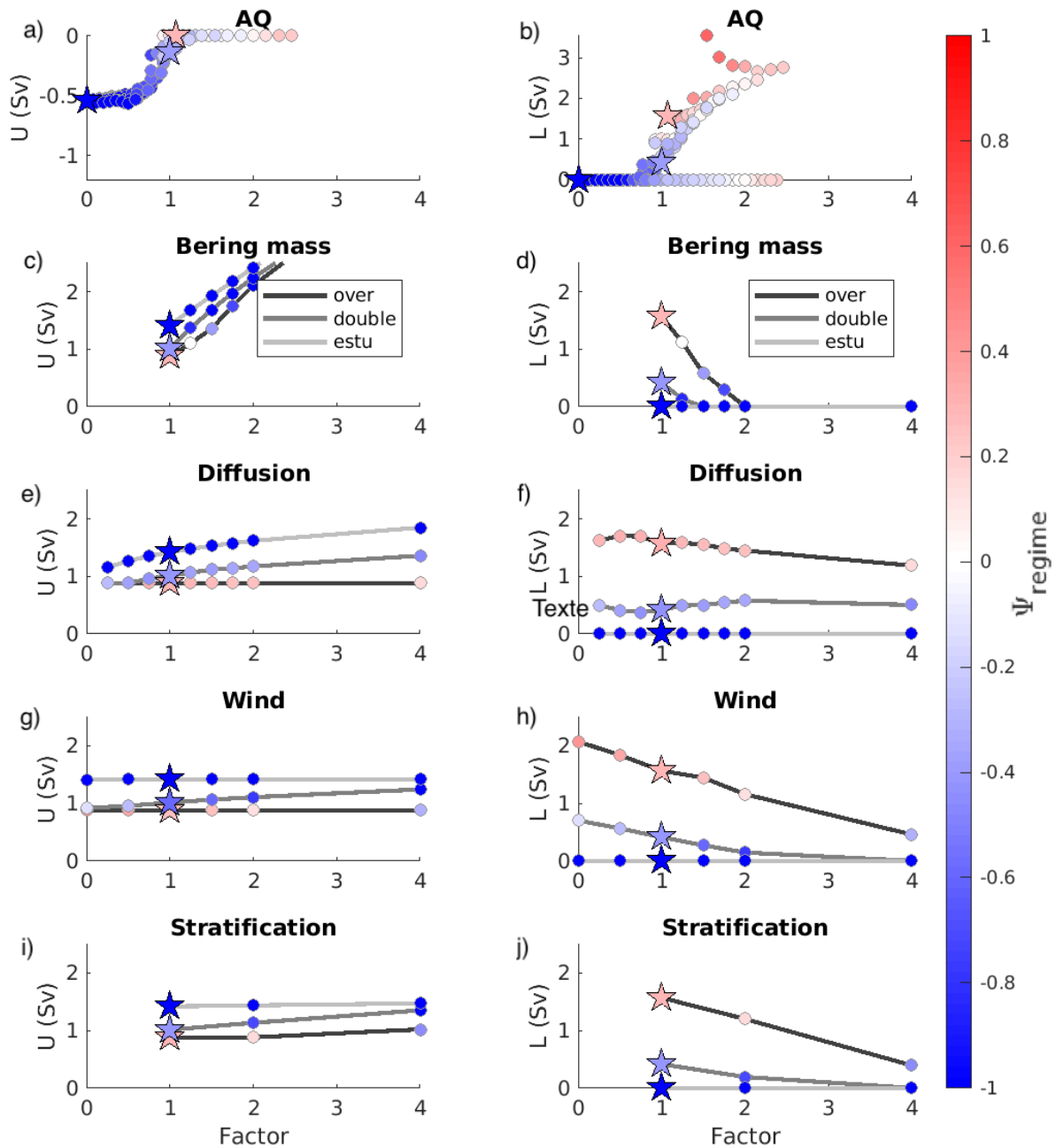


Figure 25 – Intensity of the estuarine cell, a), c), e), g) and i), and of the convective cell, b), d), f), h) and j), as a function of various forcing mechanisms for which the intensity as by multiplied by a factor compared to references experiments. The reference estuarine ($A = 0$ and $Q = 0$), double estuarine ($A = 2.5$ and $Q = 0.2$) and convectively dominant double estuarine ($A = 0$ and $Q = 0.35$) circulations are identified by a star. The color of the point refers to the circulation type for long time-average fields.

circulation. At the same time, the estuarine cell is strengthened by the enhanced mass input. This result is somewhat unexpected, as previous studies have emphasized salinity gradients as the primary driver of estuarine circulation. In contrast, we find that varying the salinity at Bering Strait while holding the mass flux constant produces no discernible change in either circulation cells (not shown). This suggests that, in the Arctic, the estuarine cell may operate in a distinct dynamical regime, one that is driven more by mass input than by salinity gradients.

As expected, diapycnal diffusion strongly affects the estuarine branch. Because estuarine circulation relies on vertical mixing, enhanced diffusivity strengthens the estuarine cell, while the convective cell remains largely insensitive to diffusion. Wind stress exerts a strong control on the convective branch by reinforcing the halocline. Increasing wind stress progressively weakens the convective circulation until it collapses. Southern boundary stratification influences both branches. Stronger stratification enhances the halocline and favors the estuarine regime, requiring larger surface salt fluxes to trigger convection.

Overall, the other forcing mechanisms mainly influence the estuarine cell of the circulation. Considering the reference estuarine circulation, we explore the influence of the mass flux at Bering Strait, vertical diffusion, wind curl and stratification at the southern boundary of the heat flux at Fram Strait, Fig. 26. The mass flux at Bering Strait doesn't influence the heat import at Fram Strait since the inflowing freshwater from the Strait stays at the surface and exits at the southern boundary without changing the circulation below. The vertical diffusivity is the forcing mechanism that has the stronger influence on heat import since it amplifies the whole estuarine cell and therefore promotes the import of Atlantic waters at depth. The wind curl doesn't affect the intensity of the estuarine cell and therefore has little impact on the heat import. Finally, the stratification at the southern boundary influences the estuarine cell and heat import in the same fashion as the vertical diffusion but its influence is not as strong as the latter.

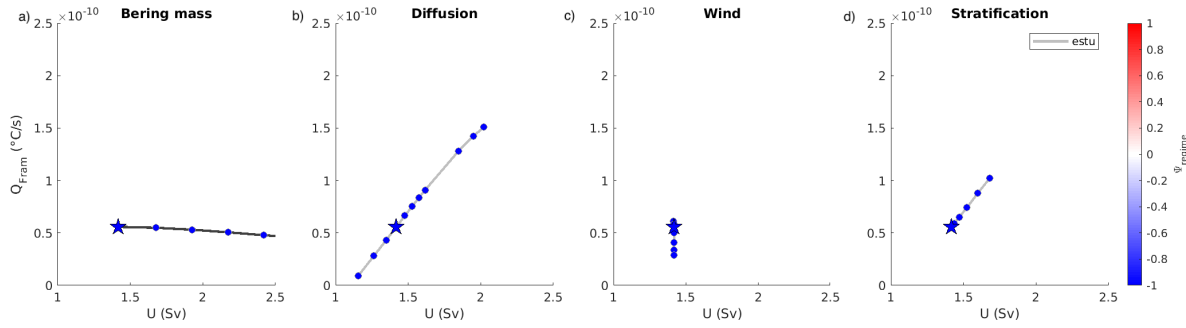


Figure 26 – Correlation between the heat flux anomaly with reference temperature -1.5°C through Fram Strait ($\Delta\langle\Theta\rangle_A$) and the intensity of the estuarine cell for various forcing mechanisms for which the intensity as by multiply by a factor compared to reference estuarine ($A = 0$ and $Q = 0$).

2.6 Conclusion

Using an idealized model, we showed that the arctic overturning circulation can take the form of a single estuarine cell, a double estuarine circulation or a single convective cell. The double estuarine form is remarkably stable and cannot collapse into a fully convective regime as long as Bering Strait continues to supply mass to the Arctic basin.

Among all forcing mechanisms, the amplitude and mean value of the seasonal salt flux over the Eurasian shelves, linked to sea ice formation, exert the strongest control on the strength of the overturning cells. Additional factors including the mass inflow through Bering Strait, diapycnal diffusivity, wind stress, and stratification in the Nordic Seas also modulate the circulation in ways comparable to the salt flux forcing by shifting the convective threshold in the A - Q space.

We further find that a single estuarine cell is associated with a weaker heat transport through Fram Strait compared with the double estuarine regime. The enhanced mass inflow that accompanies the development of a convective cell leads to a substantial increase in the heat flux entering the Arctic basin, which is reflected in the mean temperature of the upper 520

m. In contrast, the vertically averaged temperature of the full water column is warmer under a purely estuarine circulation, likely due to the presence of warm Atlantic waters below the halocline. Once a convective cell develops, the downwelling of cold surface waters cools the deeper layers and lowers the basin-averaged temperature. Despite its simplicity, our model captures the key features of the Arctic overturning circulation and provides insight into the mechanisms that govern its variability and stability.

Future work should incorporate a dynamic sea-ice model to capture the feedback between heat and salt fluxes. Doing so would make it possible to assess whether the presence of sea ice sets thresholds in heat fluxes and how these thresholds modify the arctic basin's mean temperature across the A – Q space. It would also be valuable to include the seasonal variability of the Bering Strait inflow and of surface wind stress to investigate how they interact with the system's other forcing mechanisms.

Finally, in a previous study Shink et al. (in prep.) showed that, under global warming, the double estuarine circulation shifts from a regime with $\Psi_{\text{regime}} \sim 0$ to a more convectively dominated regime as both the amplitude (A) and mean value (Q) of the surface salt flux increase. This change corresponds to a shift toward the upper-right corner of the A – Q space presented in this study. To better understand this evolution, it would be worthwhile to investigate the overturning circulation in a transient regime in which the stratification at the southern boundary and the monthly surface temperature evolve according to a global warming scenario.

2.7 Appendix A

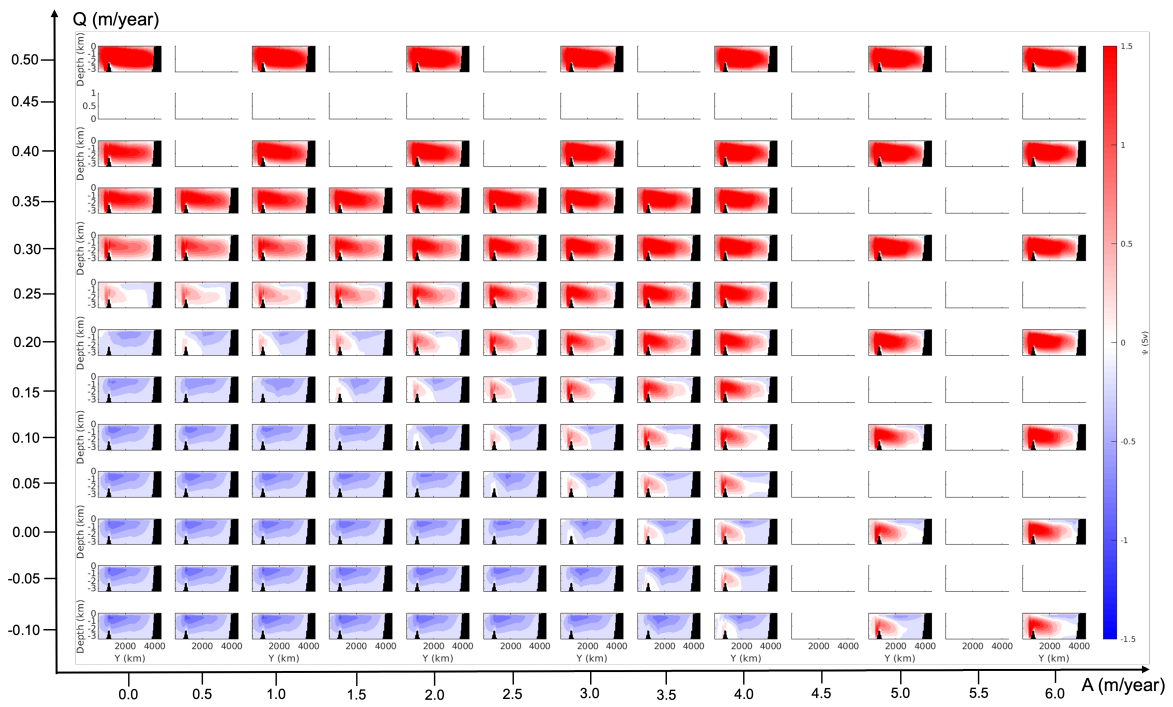


Figure 27 – 50 years averaged overturning circulation obtained for the full A - Q space. The black solid lines are the isopycnals of σ_2 ranging from 37.25 to 37.48 on a log scale.

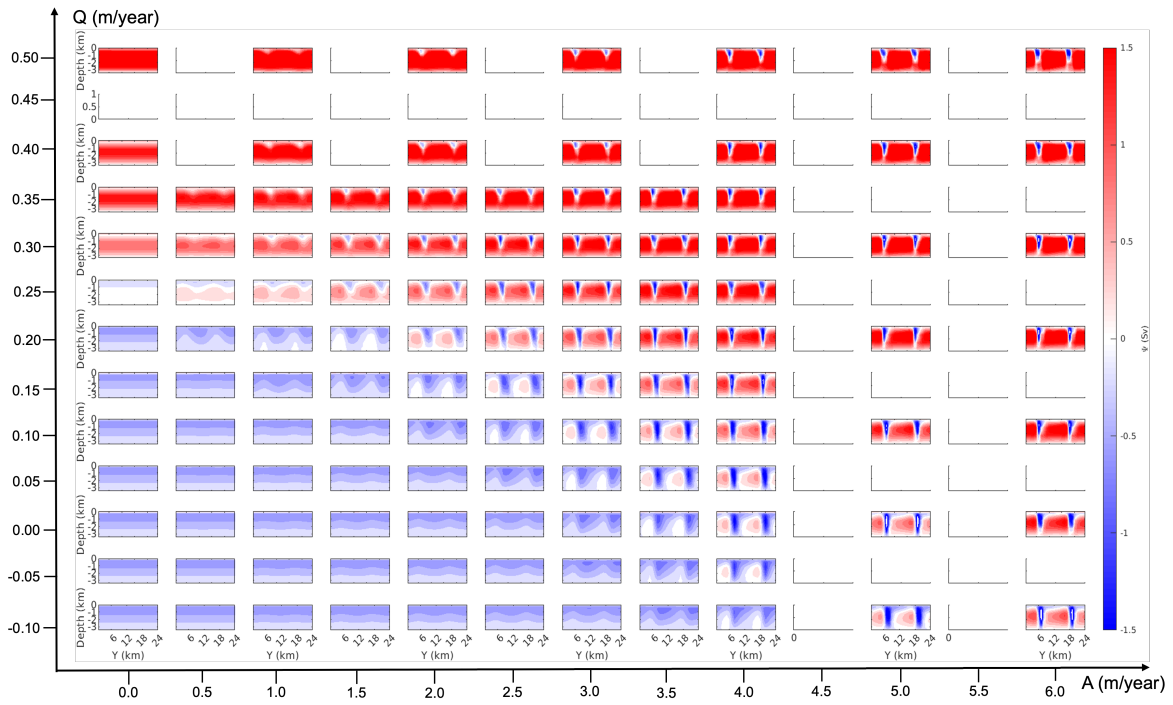


Figure 28 – Time evolution of the overturning circulation obtained for the full A - Q space. Isopycnals are omitted since their variability is negligible at monthly time scales.

CONCLUSION GÉNÉRALE

Dans le cadre de cette thèse, nous nous sommes intéressés à la circulation estuarienne double arctique et à sa dynamique. En raison de la géométrie de l'océan Arctique, les parcours d'intégration traditionnellement utilisés pour diagnostiquer la circulation de renversement entraînent une perte d'information. En effet, en suivant les parallèles de latitude, nous avons montré que la branche estuarienne de la circulation est difficilement détectable et que la branche convective est très faible.

En proposant de nouveaux parcours d'intégration et en les comparant à des parcours alternatifs trouvés dans la littérature, nous constatons que différentes informations relatives à la circulation de renversement peuvent être extraites selon la projection utilisée. Bien que la projection de la circulation entraîne nécessairement une perte d'information, en utilisant des parcours orientés vers la côte eurasienne, la cellule convective est mieux représentée, ce qui facilite l'étude de la formation des eaux profondes. En utilisant des parcours reliant le détroit de Béring au détroit de Fram, les deux cellules de la circulation peuvent être diagnostiquées distinctement.

À l'aide du modèle CESM-LE, nous avons montré que la circulation de renversement arctique présente un fort cycle saisonnier. En hiver, la cellule convective est présente sur l'ensemble du bassin arctique et la cellule estuarienne est contrainte à la surface. À l'inverse, en été, la cellule estuarienne s'étend sur tout le bassin arctique tandis que la cellule convective se retranche vers les mers nordiques.

Comme démontré à l'aide de notre modèle simplifié, ce cycle saisonnier est lié au flux de sel à la surface (représentant la formation de glace de mer dans ce cas-ci) le long de la côte eurasienne. En hiver, la formation de glace crée des eaux saumâtres très denses à la surface. En été, la fonte de glace continentale, l'apport fluvial et le transport au détroit de Béring apportent des eaux douces et peu denses à la surface. Ainsi, en hiver, les eaux

denses formées en surface sont suffisamment lourdes pour traverser localement l'halocline et générer de la convection. On note un décalage de deux mois entre l'apport en sel à la surface et la formation de la cellule convective dans notre modèle simplifié. Pour le modèle CESM, un décalage de trois mois entre le flux de sel et l'intensité de la cellule convective permet d'obtenir la meilleure corrélation (pas montré).

En été, l'apport en eaux douce restratifie la colonne d'eau ce qui permet le maintien de la cellule estuarienne et son amplification par l'ajout d'un flux de masse provenant des eaux continentales et du Pacifique. Cette relation n'a cependant pas pu être observée directement avec le modèle CESM. Cela peut être dû au fait que le temps de réponse de la cellule estuarienne à la fluctuation de flottabilité à la surface est plus long que la période de ce signal. Pour vérifier cette hypothèse, il pourrait être pertinent de caractériser le temps de réponse de la cellule estuarienne à l'aide de notre modèle idéalisé.

L'utilisation du modèle simplifié nous a ensuite permis d'isoler l'influence d'autres mécanismes de forçage de la circulation estuarienne double. Nous avons montré qu'une augmentation du flux de masse à Béring, du vent à la surface, et de la stratification dans les mers nordiques favorisent la cellule estuarienne et réduisent l'intensité de la cellule de convection. En effet, ces trois mécanismes renforcent la stratification. Les eaux saumâtres formées en surface doivent donc être plus denses pour traverser l'halocline. Si un ou plusieurs de ces mécanismes augmentent suffisamment, la cellule convective peut s'effondrer. Tel qu'attendu, l'intensité de la cellule estuarienne dépend de la diffusivité, et donc du mélange, et du flux à Béring. Nous avons toutefois montré que, contrairement à un estuaire classique, l'intensité de la cellule estuarienne arctique dépend du flux de masse à Béring et non pas du flux de sel, lequel modifie le gradient de salinité dans le domaine.

Ces conclusions permettent une meilleure compréhension de la circulation obtenue avec CESM dans un contexte de réchauffement climatique. Ici, le flux à Béring n'augmente pas de manière notable dans le temps et la diffusivité verticale reste constante ce qui implique l'invariance de l'intensité de la cellule estuarienne. L'amplitude du cycle saisonnier du flux

de flottabilité à la surface ainsi que le flux annuel moyen augmentent ce qui mène à une intensification de la cellule de convection. L'augmentation de la stratification en Atlantique Nord et dans les mers nordiques, rapproche le maximum de la cellule convective de la surface, un indicateur de l'atlantification.

Finalement, le modèle simplifié MITgcm nous indique que les circulations de type estuariennes sont plus froides au dessus de 520 m que les circulations estuariennes double ou purement convectives. En effet, pour une circulation estuarienne, les eaux froides et douces s'accumulent dans la partie supérieure de la colonne d'eau du bassin arctique ce qui diminue la température moyenne dans cette partie de la colonne d'eau. Si l'on considère la température moyenne sur l'ensemble du bassin par contre, la circulation estuarienne est plus chaude que la circulation estuarienne double, en raison de l'import d'eaux atlantiques chaudes en profondeur. La circulation estuarienne double devient plus froide puisque la formation d'une cellule convective permet d'exporter les eaux refroidies en surface vers le fond du bassin. Lorsque la cellule convective devient dominante dans le système, ce phénomène est contrebalancé par l'import d'eaux atlantiques par le détroit de Fram et la température moyenne augmente. Cette augmentation de la température est également observée avec le modèle CESM, mais dans ce cas elle résulte à la fois de l'augmentation du transport et du réchauffement des eaux atlantiques.

Nos analyses ont également mis en évidence un mécanisme de rétroaction entre l'augmentation de la stratification en Atlantique Nord qui accroît l'intensité de la cellule convective ce qui amplifie l'import d'eaux atlantiques chaudes. Ces dernières favorisent la fonte de la glace de mer particulièrement en été. Au début de la saison hivernale une plus grande surface de l'Arctique, notamment sur la côté eurasienne, est alors exposée à l'atmosphère ce qui augmente le flux de chaleur de l'océan vers l'atmosphère et permet la formation de davantage de glace de mer pendant cette saison. L'effet combiné d'un refroidissement accru et d'un rejet de sel plus important augmentent à leur tour l'intensité de la cellule convective et la stratification. Cette boucle rétroactive n'a pas pu être testée avec le MITgcm puisque

les expériences étudiées avec ce modèle étaient toutes à l'équilibre et qu'il serait nécessaire d'avoir un système transitoire pour observer cette dynamique.

Les deux modèles utilisés dans le cadre de cette thèse présentent des limites qu'il est important de prendre en considération pour interpréter correctement nos résultats. Considérons d'abord le modèle CESM. Il est connu que les eaux pacifiques ne sont pas correctement représentées dans ce modèle et que la structure verticale de la colonne d'eau pourrait être plus réaliste.

De plus, la grille Arawaka-B, sur laquelle est construit le modèle océanique, limite grandement les analyses pouvant être réalisées à l'aide des sorties du modèle. En effet, comme les vitesses sont données sur les coins des cellules, et non pas aux centres des faces, elles ne peuvent pas être utilisées directement pour calculer des quantités vectorielles. Pour pouvoir les utiliser, il faudrait interpoler les vitesses sur les faces des cellules en respectant le schéma d'advection du modèle. Nous avons tenté de réaliser cette interpolation en suivant le manuel d'utilisation du modèle, mais avons trouvé des erreurs dans celui-ci. Après avoir contacté les développeurs, nous avons été informés que les diagnostics fournis en libre accès ne permettent pas de reconstruire les vitesses sur les faces des cellules et donc que toute quantité vectorielle horizontale est inutilisable pour étudier la dynamique du modèle.

Nous avons contourné ce problème en utilisant la vitesse verticale pour calculer la circulation de renversement. Toutefois, la vitesse verticale permet uniquement de calculer la circulation dans l'espace de profondeur et non dans l'espace de densité qui nous renseigne sur la circulation baroclinique. Cette contrainte technique nous a également empêché de réaliser un bilan de chaleur puisqu'il nous a été indiqué que celui-ci ne ferme pas point par point en utilisant les diagnostics disponibles publiquement.

Tel que mentionné dans l'article 1, le déversement d'eaux denses entre les mers nordiques et l'Arctique n'est pas directement représenté dans le modèle. Pour tenir compte de ce transport, des flux de masse sont imposés au travers de la topographie. Nous avons négligé

les impacts de cette discontinuité en intégrant la circulation de renversement du nord vers le sud puis du sud vers le nord jusqu'à un point de rencontre dans les mers nordiques. Le décalage entre les deux circulations est faible, mais aurait pu être nul si les flux imposés dans la topographie avaient été mis à notre disposition.

Concernant notre modèle simplifié basé sur MITgcm, bien que nous ayons pu reproduire la structure générale de la circulation estuarienne double, certaines particularités de celle-ci pourraient être améliorées. Par exemple, la cellule estuarienne y est généralement plus faible que la cellule convective, alors qu'avec le CESM nous obtenons deux cellules d'amplitudes comparables. Cette sous-représentation est probablement liée au fait que nous n'avons pas inclus l'apport fluvial dans ce modèle. Étant donné que le flux de masse entrant est un mécanisme important pour la cellule estuarienne, un apport supplémentaire d'environ 1Sv provenant des rivières aurait un effet non négligeable.

Par ailleurs, la structure de l'halocline diffère de celle observée ou représentée dans les modèles climatiques. Des ajustements seraient donc nécessaires pour améliorer sa représentation dans notre modèle. De plus, nous avons uniquement considéré un cycle saisonnier pour le flux de sel à la surface, mais il serait pertinent de prendre en compte un cycle saisonnier pour la température à la surface, le transport à Béring et le vent au-dessus du bassin arctique. En ignorant la variation temporelle de ces forçages, nous avons simplifié le système et ne pouvons pas étudier les interactions entre différents mécanismes de forçage au cours d'une année.

La bathymétrie utilisée est simplifiée et ne permet pas de prendre en considération l'effet des crêtes océaniques, par exemple. Des tests préliminaires, non présentés dans cette thèse, indiquent que l'ajout d'une crête au centre du bassin modifie la structure de la circulation et complexifie significativement cette dernière. Dans le cadre de ce projet, l'objectif de ce modèle était de reproduire le plus simplement possible et donc cette modification de la bathymétrie n'a pas été incluse. Cependant, dans des travaux futurs pour lequel un modèle plus complexe est utilisé et couplé, par exemple, à un modèle de glace, il serait intéressant

d'inclure une crête océanique dans le bassin pour explorer la dynamique résultante.

Pour conclure, ce projet de recherche nous a permis de développer un outil innovant offrant une perspective nouvelle de la circulation estuarienne double arctique et de mieux comprendre sa dynamique, ce qui soulève de nouvelles questions de recherche qu'il serait intéressant à explorer.

Dans un premier temps, puisque nous disposons désormais d'un modèle simplifié de l'Arctique, il serait possible d'y intégrer un modèle de glace de mer également idéalisé. En incluant directement la dynamique de la glace dans nos expériences, nous pourrions explorer les phénomènes de rétroaction entre la couverture de glace, la circulation de renversement et le transport de chaleur. Avec cette nouvelle configuration, il serait possible de répéter les analyses présentées à l'article 2 afin de déterminer si les conclusions trouvées restent les mêmes ou si des nuances doivent y être apportées.

Par la suite, il serait intéressant de déterminer le temps de réponse des cellules estuarienne et convective à leur forçage respectif. Deux types d'expériences pourraient être faites à cet effet. Dans un premier scénario, nous pourrions considérer une circulation estuarienne et une circulation convective indépendamment, cesser de forcer la circulation et évaluer le temps nécessaire pour atteindre un nouvel état d'équilibre. Il est attendu que ce temps de réponse soit très long, limitant les informations que l'on peut en tirer. Pour cette raison, une seconde approche consisterait à appliquer une variation cyclique à l'amplitude de nos mécanismes de forçage et à y insérer une anomalie ponctuelle afin d'évaluer le temps de réponse du système à cette perturbation.

Cette approche aborde une autre thématique qu'il serait intéressant d'explorer soit la dynamique du système dans un régime transitoire, plus représentatif dans le contexte actuel de réchauffement climatique. Pour ce faire, nous pourrions initialiser une expérience avec une circulation estuarienne double similaire à celle diagnostiquée au début des années 1900 avec le CESM, puis faire évoluer les mécanismes de forçage en suivant les tendances d'un

scénario de réchauffement climatique et caractériser la réponse du système en temps réel.

Ensuite, les vitesses à la frontière sud du modèle MITgcm étaient libres de varier, ce qui ne nous a pas permis d'étudier l'influence du transport de masse dans les mers nordiques comme mécanisme de forçage. Il serait possible de prescrire un transport à cette frontière en imposant un gradient de densité horizontal, générant un courant dont l'intensité pourrait être contrôlée par ce gradient de densité.

Finalement, plusieurs modèles de l'Arctique avec une résolution horizontale plus fine que ceux utilisés pour cette étude sont accessibles. Les analyses réalisées avec le CESM étant limitées par la grille du modèle, il serait intéressant de reproduire nos travaux avec un modèle plus réaliste pour explorer plus en détail le bilan de chaleur par exemple et ajouter des nuances à nos interprétations.

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