

Enhanced Water Mass Mixing in Fram Strait in 2020 and elevated Circulation Timescales of Atlantic-derived Waters in 2021 based on Transient Tracers I-129 and U-236

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Abstract.

The Arctic Ocean is undergoing rapid climate-driven change, making it increasingly important to quantify the circulation pathways and transit times of Atlantic Water entering and exiting the basin. Fram Strait, the primary gateway for Arctic–Atlantic exchange, provides a key location for assessing the evolving connectivity between the Arctic and the subpolar North Atlantic. Here, we combine the anthropogenic radionuclide tracer pair Iodine-129 (¹²⁹I) and Uranium-236 (²³⁶U), with binary mixing and Transit Time Distribution (TTD) modelling to investigate the origin and transit history of surface Polar Water and mid-depth Atlantic Water sampled in Fram Strait between 2016 and 2021. Our results reveal significant interannual variability. Waters outflowing the Fram Strait in 2020 exhibited a higher degree of mixing and a stronger influence from Amerasian Basin sourced waters compared to 2016 and 2021. We further identify a distinct water parcel on the Greenland Shelf whose tracer signature indicates a long-path circulation originating from the Canada Basin. In contrast, waters sampled in 2021 exhibit generally longer transit times, consistent with either slower circulation or longer transport pathways. These results demonstrate substantial year-to-year variability in Arctic Ocean export through Fram Strait and highlight the need for sustained tracer observations to constrain changes in the circulation of the heat-bearing Atlantic Water layer and its role in Arctic–North Atlantic exchange.

1 Introduction

1.1 Fram Strait – The Primary Gateway Between the Arctic Ocean and the Subpolar North Atlantic

The Fram Strait serves as the principal gateway for water mass exchange between the Arctic Ocean and the subpolar North Atlantic (Timmermans and Marshall, 2020). On its eastern side, the West Spitsbergen Current (WSC) (Fig. 1) transports relatively warm and saline Atlantic Water into the Eurasian Basin (Aagaard et al., 1987; Timmermans and Marshall, 2020), acting

20 as a major heat source for the Arctic Ocean (Schauer et al., 2004; Besczynska-Möller et al., 2011; Smedsrud et al., 2022), continuously gaining relevance as the Atlantic Water supply to the Arctic Ocean intensifies (Wang et al., 2020). This process has received increasing attention due to Atlantification — the progressive intrusion of increasingly warm Atlantic Water into the Arctic Ocean, a process also observed in the Fram Strait itself (Tesi et al., 2021) — which is now observed to extend beyond the Eurasian Basin into the Amerasian Basin (Ingvaldsen et al., 2021; Polyakov et al., 2023, 2025b). The enhanced presence of
25 Atlantic Water weakens vertical stratification, promotes upward heat flux, and accelerates sea ice melt, contributing to Arctic Ocean Amplification, whereby temperatures rise more than two times faster than the global average (Box et al., 2019; Shu et al., 2022). These unprecedented changes driven by anthropogenic climate change are projected to result in a seasonally ice-free Arctic Ocean by mid-century (Kim et al., 2023; Jahn et al., 2024), with significant implications for circulation dynamics (Meredith et al., 2019).

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On the western side of the strait, the East Greenland Current (EGC) transports water masses from the central Arctic Ocean to the subpolar North Atlantic with its core situated at the Greenland shelf slope (Rudels et al., 2002). The EGC reflects the integrated outcome of upstream Arctic circulation and is composed of two primary layers in the upper part: a surface layer of cold, low-salinity Polar Water and a mid-depth layer of relatively warmer and saltier Arctic Atlantic Water (Rudels et al.,
35 2005; Håvik et al., 2017). These layers play a critical role in modulating freshwater and heat fluxes to the North Atlantic (Le Bras et al., 2021) and the freshwater flux has shown to vary temporally (Karpouzoglou et al., 2022, 2024). The Arctic Ocean is an important player in contributing waters to the Atlantic Meridional Overturning Circulation (AMOC) (Zhang and Thomas, 2021; Weijer et al., 2022; Dey et al., 2024) and therefore ultimately influencing stratification and deep water formation of the AMOC's lower limb.

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The surface Polar Water typically occupies the top 250 m in the Arctic Ocean and is characterized by potential densities (σ_0) below 27.7 kg m^{-3} , which includes the polar mixed layer and the halocline (Rudels et al., 1994; Rudels, 2009, 2022). Polar Water comprises a mixture of Pacific Water entering through the Bering Strait (orange arrows in Fig.1) and mainly Atlantic Water (green arrows in Fig.1). The Atlantic component, entering as the Norwegian Coastal Current (NCC), follows a
45 coastal path through the Barents Sea and into the Nansen Basin, eventually reaching the Fram Strait via the Transpolar Drift (Rudels and Carmack, 2022). Along this route, surface waters are modified by freshwater inputs from Siberian rivers and sea-ice melt, resulting in dilution and enhanced stratification (Bauch et al., 2011; Fichot et al., 2013), although north of Svalbard this stratification can be episodically disrupted by Atlantic Water entering through Fram Strait (McPherson et al., 2026a). The Pacific component enters through Bering Strait, becoming part of the Beaufort Gyre and ultimately joining the Transpolar Drift
50 towards Fram Strait (Rudels, 2022; Rudels and Carmack, 2022). The fraction of Pacific Water in the Polar Water outflowing Fram Strait is of great interest to improve predictions of freshwater export from the Arctic Ocean to the subpolar North Atlantic as the Pacific Water is relatively fresh compared to Atlantic Water and contributes substantially to the freshwater budget (Rabe et al., 2013; Carmack et al., 2016). The Pacific Water fraction has traditionally been estimated using nutrient-based tracers (Jones et al., 1998, 2003; Falck et al., 2005; Dodd et al., 2012), with the composition in the Fram Strait varying substantially

55 over time (Dodd et al., 2012) and showing a dependence on the Arctic Oscillation index (Steele et al., 2004). However, these
nutrient-based tracers do not act in the conservative way as previously assumed (Bauch et al., 2011; Alkire et al., 2015, 2019).
Modeling efforts rendered minor fractions of Pacific Water, which needs to be validated by observations (Lique et al., 2010).
The recent exploration of the new Gallium tracer has shown to deliver promising results to disentangle Atlantic and Pacific
60 ploration of alternatives as a state of the art tracer is still missing. Using Iodine-129 (^{129}I) and Uranium-236 (^{236}U) as a new
tracer tool to disentangle Pacific and Atlantic contributions has shown to be another promising approach that will be used here
again in a qualitative way (Wefing et al., 2022).

The Arctic Atlantic Water layer in the Arctic Ocean, typically residing between 400 m and 700 m depth (Rudels, 2009),
65 is a blend of two main Atlantic Water pathways: a long route through the Canada and Makarov Basins (Smith et al., 1999;
Woodgate et al., 2001; Karcher et al., 2012), and a shorter route through the Amundsen and Nansen Basins (purple arrows
in Fig.1) (Woodgate et al., 2001; Aksenov et al., 2011; Rudels et al., 2015; Wefing et al., 2019). Compared to Polar Water, Arctic
Atlantic Water in the Fram Strait is characterized by a density range of $\sigma_0 \approx 27.7$ to 27.97 kg m^{-3} (Rudels, 2009, 2022) and
a temperature maximum (Coachman and Barnes, 1963; Rudels et al., 1996; Rudels, 2022). The Arctic Atlantic Water consists
70 of two main branches: Fram Strait Branch Water (FSBW), which entered the Arctic Ocean via the Fram Strait, and Barents Sea
Branch Water (BSBW), which entered the Arctic Ocean via the Barents Sea. Both branches converge in the St. Anna Trough
and form the Arctic Ocean Boundary Current circulating cyclonically along the continental slope (Smith et al., 2011; Rudels
et al., 2015). These relatively warm branches are of considerable scientific interest, primarily due to their increasing potential
to transfer heat upward, and thereby act as a positive feedback loop of accelerating sea ice melting (Polyakov et al., 2020a;
75 Wang et al., 2024b). Consequently, quantifying and assessing the variability of circulation timescales and pathways for Arctic
Atlantic Water is paramount not only for understanding the influence of this heat-bearing layer on the Arctic climate system in
its recent past and future but also for comprehending its ecological impact (Greene et al., 2008; Polyakov et al., 2023; Brown
et al., 2025; Årthun et al., 2025b).

80 Apart from the described outflowing layers, the Fram Strait is a two-way gateway, facilitating also inflowing water masses
within the WSC, which makes it a region of intense mixing. A portion of the WSC recirculates within the strait and merges
with the EGC, forming Recirculating Atlantic Water (Quadfasel et al., 1987; Hattermann et al., 2016; Rudels, 2022). Approx-
imately half of the Atlantic Water entering the strait between 76° N and 81° N is estimated to recirculate as Recirculating
Atlantic Water (Marnela et al., 2013; de Steur et al., 2014), with a southern and northern recirculation at around 78.5° N and
85 80° N , respectively (black arrows in Fig.1b) (Hofmann et al., 2021). This process is partly driven by mesoscale eddies that
can extend westward into the EGC (Hattermann et al., 2016). The Recirculating Atlantic Water subducts beneath the Polar
Water as it joins the EGC due to its greater density and mixes with the Arctic Atlantic Water at mid-depth (Marnela et al.,
2013; Hattermann et al., 2016; von Appen et al., 2016; Hofmann et al., 2021). Efforts to understand the recirculation in the
Fram Strait have mostly gone into describing the existence, magnitude, location, seasonality of recirculation (Quadfasel et al.,

90 1987; Manley, 1995; Marnela et al., 2013; von Appen et al., 2016; Hattermann et al., 2016; Wekerle et al., 2017; Hofmann
et al., 2021) and most recently into describing episodic northward flows within the EGC (McPherson et al., 2026b) (preprint).
However, temporal dynamics of the recirculation regime are still not well understood but have implications for the Atlantic
Water heat reaching the Northeast Greenland shelf (McPherson et al., 2023). Here we study the origin of the water masses in
the Fram Strait in different years, which brings further light into temporal recirculation processes.

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Given its role as the primary outflow pathway, the western Fram Strait offers a valuable observational window into upstream
Arctic processes and their integrated effects. Both Polar Water and Arctic Atlantic Water are central to the transmission of Arc-
tic changes to the subpolar North Atlantic. Advancing our understanding of the circulation timescales, mixing and provenance
of these water masses is therefore essential for assessing the evolving dynamics of the Arctic–Atlantic system. To this end, we
100 employ the transient radionuclide tracers ^{129}I and ^{236}U , which provide powerful constraints on the circulation timescales of
Arctic–Atlantic waters.

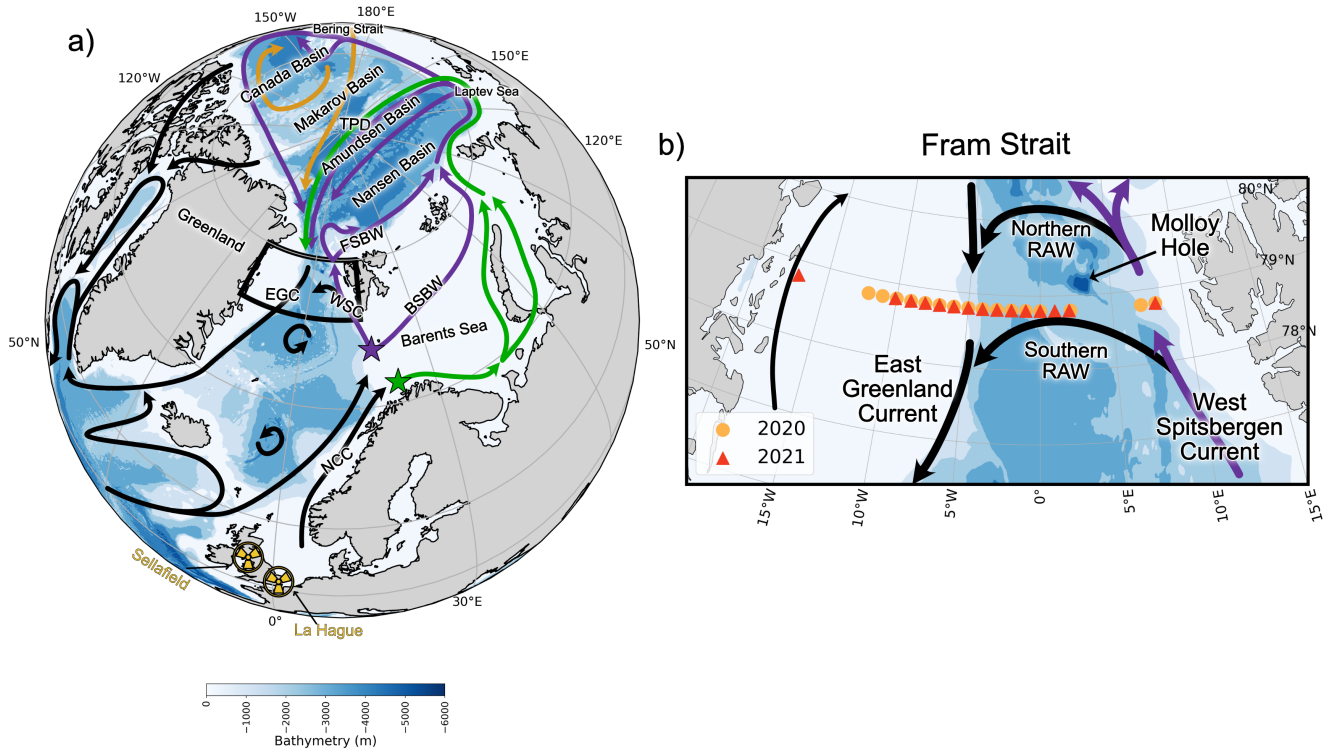


Figure 1. a) Overview map of the main circulation regime in the Arctic and North Atlantic. The location of the Nuclear Reprocessing Plants is shown as yellow radioactive symbols. The location of the two radionuclide input functions are indicated as green (surface layer) and purple (mid-depth layer) stars (the input functions are shown in Fig. 2). The surface Atlantic-derived circulation regime is indicated as green arrows with orange arrows showing the Pacific influence from Bering Strait. The mid-depth circulation regime is shown as purple arrows. Black arrows depict the general surface circulation regime outside of the input function domain. Inlet b) depicts a more detailed circulation regime in the Fram Strait, showing the stations sampled during 2020 and 2021 for the purpose of this work. RAW: Recirculating Atlantic Water.

1.2 Anthropogenic Radionuclides ^{129}I and ^{236}U as Tracers of Circulation Timescales and Mixing in the Fram Strait

The anthropogenic radionuclides ^{129}I and ^{236}U serve as tracers for assessing the dynamics of the Arctic-Atlantic system and quantifying circulation timescales in the Fram Strait (Wefing et al., 2019, 2021, 2022). While initially focused on the Fram Strait, applications have recently expanded to the central Arctic Ocean, the Canadian Basin, the Iceland, Labrador, Barents and Kara Seas (Payne et al., 2024; Leist et al., 2024; Dale et al., 2024; Pérez-Tribouillier et al., 2025; Wefing et al., 2025). Both isotopes are long-lived (half-lives: ^{129}I = 15.7 Ma; ^{236}U = 23.4 Ma) and behave conservatively in the open ocean (Edmonds

et al., 2001; Christl et al., 2012; Casacuberta et al., 2014).

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Their presence in the marine environment arises from: (i) Nuclear Reprocessing Plants (NRPs), with liquid discharges from the Sellafield (UK) and La Hague (France) facilities being the primary source for both isotopes (see Fig. 1)(Raisbeck et al., 1995; Snyder et al., 2010; Christl et al., 2015b); and (ii) atmospheric weapon tests, with global fallout from atmospheric nuclear detonations that took place in the 1950s - 60s being a secondary, substantial source, especially for ^{236}U (Sakaguchi et al., 2009). The input functions for both tracers entering the Arctic Ocean have been defined (Casacuberta et al., 2018) and updated (Wefing et al., 2021) at the Barents Sea Opening (Fig. 1). Crucially, the signal timing differs: the major injection of ^{129}I occurred in the late 1990s, whereas the peak ^{236}U input occurred earlier, in the 1960s (Fig. 2).

Historically, various tracer-based approaches have been employed to estimate the circulation timescales of Atlantic-derived waters. Rudels et al. (2000) inferred a return pathway of FSBW of under 10 years based on temperature anomalies. Subsequently, multi-tracer studies using ^{129}I , Cesium-137 (^{137}Cs), and chlorofluorocarbon-11 (CFC-11) estimated tracer ages of 8–10 years for surface layers and 9–15 years for mid-depth waters near the North Pole (Smith et al., 2011). Similarly, Stöven et al. (2016) applied a Transit Time Distribution (TTD) model using chlorofluorocarbon-12 (CFC-12) and sulfur hexafluoride (SF_6) in the Fram Strait, deriving mean ages of 7 ± 6 years for Polar Water and 32 ± 15 years for Arctic Atlantic Water. While effective for ventilation timescales, the application of gaseous tracers in surface layers is constrained by air-sea exchange, a process likely to intensify with diminishing sea ice (Bates et al., 2006). Conversely, radionuclides offer a robust alternative for investigating lateral circulation timescales (Wefing et al., 2021; Casacuberta and Smith, 2023; Payne et al., 2024).

The first ^{129}I transect in the Fram Strait was obtained in 2002 (Alfimov et al., 2004, 2013). At that time, the high-concentration signal from the 1990s had not yet transited the central Arctic Ocean. The signal, therefore, served primarily as an indicator of recirculating Atlantic Water at Fram Strait. More recently, Wefing et al. (2019) combined ^{129}I and ^{236}U to model circulation timescales from the northern Norwegian coast, estimating tracer ages of 12–19 years for surface Polar Water and 16–23 years for mid-depth Arctic Atlantic Water based on binary mixing between reprocessing-tagged Atlantic Water and the global fallout background. Wefing et al. (2021) refined this using a TTD approach incorporating flow-field mixing (Haine and Hall, 2002), yielding updated mean ages of 24–55 years for Arctic Atlantic Water. Other combinations, such as ^{236}U with Colored Dissolved Organic Matter (CDOM), have yielded tracer ages of 7–27 years for the upper 200 m of the western shelf (Lin et al., 2023a).

Despite this progress, the Fram Strait circulation regime remains poorly constrained due to two main limitations. First, methodological inconsistency hinders direct comparison across the water column: previous studies have alternated between tracer ages (Wefing et al., 2019; Lin et al., 2023a; Körtke et al., 2024) and TTDs (Stöven et al., 2016; Wefing et al., 2021), yielding divergent estimates of surface and mid-depth water ages. Second, the absence of time series observations of transient

tracers in Fram Strait does not allow the study of temporal variability in circulation timescales.

145 Motivated by these gaps, this study combines new ^{129}I data from 2020 and 2021 with ^{236}U data from Lin et al. (2023a) to
address three specific objectives. We calculate tracer ages for the surface Polar Water layer (binary mixing model) and extend
the TTD method to the surface layer. This establishes a uniform methodological framework, allowing for a direct comparison
between surface and mid-depth layers. Then, we qualitatively delineate the proportional Amerasian water contribution in the
Fram Strait surface layer from the Eurasian water using a mixing model and TTD results. Finally, we evaluate the temporal
150 stability of the circulation regime by comparing the new 2020 and 2021 estimates with data from 2016 published in Wefing
et al. (2019).

2 Materials and Methods

2.1 Sample Collection, Processing and Measurements

155 Water samples were collected during two oceanographic expeditions to the Fram Strait aboard the RV *Kronprins Haakon*,
conducted from 24 August 2020 to 13 September 2020, and from 31 July 2021 to 20 August 2021 (Fig. 1b). These expeditions
were part of the annual monitoring program coordinated by the Norwegian Polar Institute (de Steur, 2021, 2022). A total of 119
samples for ^{129}I analysis were collected in 2020, and 115 in 2021. Concurrently, a separate research group collected samples
for ^{236}U measurements published in Lin et al. (2023a). For intercomparison purposes and adding additional data in the WSC,
160 we also processed a subset of our samples for ^{236}U measurements, amounting to 22 from 2020 and 11 from 2021. The results
of this intercomparison are presented in Fig. A1 and demonstrate good agreement. Consequently, all ^{236}U concentrations used
in this study refer to the dataset published by Lin et al. (2023a). ^{129}I and ^{236}U concentrations measured as part of this study,
along with the ^{236}U concentrations from Lin et al. (2023a), can be found on zenodo (Scheiwiller, 2026).

165 Seawater was sampled using 12 L Niskin bottles mounted on a conductivity-temperature-depth (CTD) rosette. Samples were
transferred into pre-rinsed 1 L plastic cubitainers and stored at the Norwegian Polar Institute in Tromsø following the expedi-
tion. The 2020 samples were processed at the Norwegian Polar Institute, while the 2021 samples were shipped to ETH Zurich
for further analysis. Approximately 200 mL aliquots were used for ^{129}I processing, with the remaining volume allocated for
 ^{236}U measurements. A subset of the ^{129}I data from 2021 has been published in Pérez-Tribouillier et al. (2025). In addition to
170 this data, we incorporated ^{129}I and ^{236}U concentrations from the Fram Strait in 2016 into our analysis (Wefing et al., 2019).
The stations from 2016 are shown in Fig. A2 together with stations from 2020 and 2021.

After the expedition, ^{129}I samples were processed according to the method described in Casacuberta et al. (2016). The sea-
water samples (200 mL) were spiked with approximately 1.5 mg of Iodine-127 (^{127}I). The uncertainty related to the spike was
175 5–6 % and <1 % for 2020 and 2021, respectively. The higher uncertainties in 2020 compared to 2021 were due to a pipette mal-

functioning. After spiking, both Iodine isotopes were isolated from the seawater matrix and precipitated as silver iodide (AgI). The silver iodide powder was pressed into titanium holders and inserted into the 500 kV *Tandy* Accelerator Mass Spectrometry (AMS) system at the Laboratory of Ion Beam Physics at ETH Zurich to measure the $^{129}\text{I}/^{127}\text{I}$ atom ratio (Vockenhuber et al., 2015), resulting in a statistical uncertainty of 1–2 %. The known spiked amount of the stable isotope ^{127}I was then used to derive the ^{129}I concentration. An in-house standard, C2, with known concentrations of 5.055×10^{-12} at at^{-1} $^{129}\text{I}/^{127}\text{I}$ (diluted) and 38.995×10^{-12} at at^{-1} $^{129}\text{I}/^{127}\text{I}$ (concentrated) was measured alongside the samples. The relative difference between the measured C2 concentrations to the nominal value was used to normalize the value of the samples. MilliQ water blanks (18.2 M Ω high-purity water) from the laboratory were used to correct for the background level of ^{129}I . They were included in every processing run, amounting to 21 blanks and 7 blanks for the samples from 2020 and 2021, respectively. Additionally, an internal standard was measured repeatedly throughout the year to account for the combined uncertainty of the entire procedure, from chemical separation to final measurement. The standard deviation of these replicate measurements yielded an uncertainty of 4 %. The spike, measurement, and internal-standard uncertainties were then propagated, yielding total uncertainties of 7–10 % in 2020 and 4–5 % in 2021 for the ^{129}I concentrations. Further details on the sample processing and measurements are given in Casacuberta et al. (2016) and Payne et al. (2024).

In total 33 ^{236}U samples were processed according to the method of Christl et al. (2015a) and compared to Lin et al. (2023a) (see Fig. A1). The seawater samples (1 L) were spiked with approximately 1 pg of Uranium-233 (^{233}U). Uranium isotopes were concentrated with Fe co-precipitation and isolated from the seawater matrix with ion exchange chromatography (UTEVA resin columns). Uranium was oxidized and pressed into targets for measurements with the 300 kV *MILEA* AMS system at the Laboratory of Ion Beam Physics at ETH Zurich. The ratios of $^{233}\text{U}/^{238}\text{U}$ and $^{236}\text{U}/^{238}\text{U}$ were measured and the known amount of spiked ^{233}U was then used to derive the ^{236}U concentration. An in-house standard, ZUTRI, was measured alongside to normalize the samples with known ratios of $(33,170 \pm 830) \times 10^{-12}$ at at^{-1} for $^{233}\text{U}/^{238}\text{U}$, and $(4,055 \pm 203) \times 10^{-12}$ at at^{-1} for $^{236}\text{U}/^{238}\text{U}$. MilliQ water blanks (18.2 M Ω high-purity water) from the laboratory were used to correct for the background level of ^{236}U . They were included in every processing run, amounting to 3 blanks in total. Total uncertainties resulted in 3–6 % for ^{236}U concentrations. For further details on the processing and the measurements, we refer to Christl et al. (2015a, 2023) and Payne et al. (2024).

2.2 Water Mass Classification in the Fram Strait

To characterize the seawater samples according to their hydrographic properties, we use the classification scheme of Rudels et al. (2005) and Rudels (2022). Based on that, we defined 7 different water masses, which are listed in Table 1. Polar Water I and Polar Water I warm include the Pacific-derived water, the Polar Mixed Layer, shelf water, and the upper halocline. Polar Water I warm is further subcategorized due to seasonal summer heating and ice melt contribution (Rudels et al., 2000). Polar Water II consists of the Atlantic-derived lower halocline and the winter mixed layer from the Nansen Basin and Barents Sea. Atlantic Water is part of the WSC, arriving from the North Atlantic Ocean. The Recirculating Atlantic Water originates from

the WSC (same properties as Atlantic Water) but has left the core of the WSC. The Arctic Atlantic Water is also of Atlantic origin, but has already circulated through the Arctic Ocean and so it has changed its properties. Finally, we broadly categorize samples in deeper layers below the Arctic Atlantic Water into Intermediate and Deep Waters. As these layers are not the main objective of this study, we refrain from assigning more detailed categories. Samples that did not meet the defined boundaries were labeled as a mixture of two water masses based on their hydrographic profile.

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Samples categorized as Polar Water I were used to estimate circulation timescales with two models: (i) the binary mixing model (section 2.3.2) and (ii) the TTD model (section 2.3.4). For Arctic Atlantic Water samples only the TTD model (section 2.3.3) was used to estimate circulation timescales. Both models rely on an input function that defines the concentrations of ¹²⁹I and ²³⁶U entering the Arctic Ocean in a given year.

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Table 1. Categorization of samples into water masses according to the definition of Rudels (2022).

# in 2016	# in 2020	# in 2021	Water Mass	Pot. Density (kg m ⁻³)	Pot. Temp. (°C)
19	51	40	Polar Water I	$\sigma_0 \leq 27.20$	$\theta \leq 0$
2	5	8	Polar Water I Warm	$\sigma_0 \leq 27.20$	$0 < \theta$
9	14	19	Polar Water II	$27.20 < \sigma_0 \leq 27.70$	$\theta \leq 0$
22	14	9	Atlantic Water	$27.70 < \sigma_0 \leq 27.97$	$2 < \theta^a$
13	18	14	Recirculating Atlantic Water	$27.70 < \sigma_0 \leq 27.97$	$2 < \theta^a$
28	8	20	Arctic Atlantic Water	$27.70 < \sigma_0 \leq 27.97$	$0 < \theta \leq 2$
39	1	6	Intermediate & Deep Waters	$27.97 < \sigma_0$	-

^aSamples were divided into Atlantic Water and Recirculating Atlantic Water based on the geographical location, whereas Atlantic Water samples were at either 7° E or 8° E, and Recirculating Atlantic Water samples were further to the west.

2.3 Estimation of Circulation Timescales and Mixing

2.3.1 Input Function of ¹²⁹I and ²³⁶U into the Arctic Ocean

To characterize the time-dependent input of the radionuclide tracers ¹²⁹I and ²³⁶U into the Arctic Ocean, two input functions were defined (Fig. 2a,b): (i) the Atlantic Layer input function, located at 74° N and 19° E, and (ii) the NCC input function, located at 71° N and 22° E (Wefing et al., 2021). Their geographical positions are illustrated in Fig. 1, both located at the Barents Sea opening (Casacuberta et al., 2018; Wefing et al., 2021). These input functions serve as the upstream boundary conditions for estimating circulation timescales and mixing processes at downstream locations, such as the Fram Strait.

We distinguish between surface and mid-depth input functions, as they correspond to distinct water masses with different tracer concentrations (Casacuberta et al., 2018). The surface input function reflects the propagation of the NCC, which becomes diluted through interaction with shelf waters carrying the global fallout tracer signal. Wefing et al. (2022) quantified

this effect as a dilution factor of 2 for the surface input function prior to reaching the central Eurasian Basin and accordingly diluted it 50:50 with the global fallout background signal. We adopt the same approach here, applying a 50:50 dilution with background levels of 1.16×10^7 at L^{-1} for ^{129}I and 6.25×10^6 at L^{-1} for ^{236}U (Snyder et al., 2010; Chamizo et al., 2022; Dale et al., 2024). The mid-depth input function represents the continuation of the Atlantic Layer, comprising both the FSBW and the BSBW. These branches originate from Atlantic Water that undergoes progressive cooling and densification as it advects through Fram Strait and the Barents Sea, respectively (Rudels and Carmack, 2022). Upon reaching the Nansen Basin near the St. Anna Trough, the BSBW subducts beneath the FSBW as a result of its longer cooling trajectory. As the combined flow continues eastward along the Siberian continental slope, enhanced isopycnal mixing progressively homogenizes the vertical structure, forming a unified mid-depth layer (Rudels and Carmack, 2022). Both input functions present uncertainties (shaded areas in Fig. 2, which originate from uncertainties in reprocessing plant uranium releases (Christl et al., 2015b) and in the fractions of La Hague, Sellafield, and global fallout contributions (Casacuberta et al., 2018). Here, we have increased the uncertainties for the mid-depth ^{129}I input function by a factor of four compared to Wefing et al. (2021). This more conservative approach reflects the higher variability revealed by newly available data from 2018 published in Pérez-Tribouillier et al. (2025) and data from this study (2020 and 2021), which exceeded the range suggested by the initial uncertainty estimate.

Based on the water mass classification described in section 2.2, Polar Water I samples are considered to originate from the surface input function and are part of the Polar Mixed Layer. Polar Water II samples, which include the lower halocline, represent a transitional layer between Polar Water I and Arctic Atlantic Water. These samples may exhibit partial influence from deeper Arctic Atlantic Water tracer signals. For this reason, Polar Water II samples are excluded from the core analysis of surface and mid-depth layers. The Arctic Atlantic Water samples are interpreted as a downstream continuation of the mid-depth input function, encompassing both FSBW and BSBW components (Wefing et al., 2021). For a detailed description of the input function construction, we refer to Wefing et al. (2021).

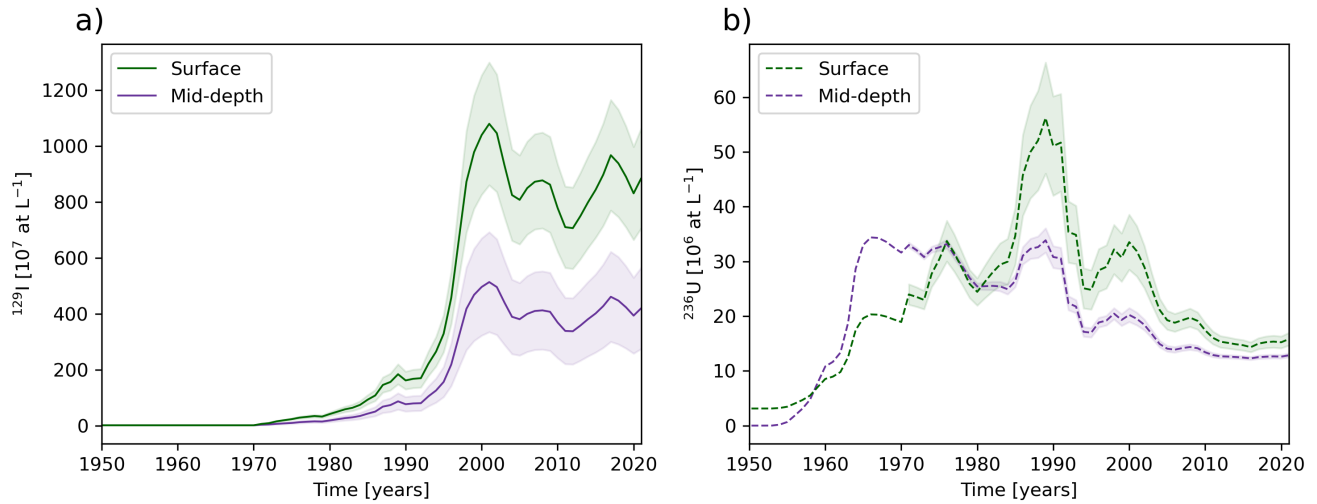


Figure 2. Input functions for ^{129}I (a) and ^{236}U (b) at the entrance of the Arctic Ocean, shown for the Polar Water layer at the surface (green lines) and the Arctic Atlantic Water layer at mid-depth (purple lines), with uncertainties shown as shaded areas. These uncertainties were propagated from uncertainties in reprocessing plant uranium releases (Christl et al., 2015b) and in the fractions of La Hague, Sellafield, and global fallout contributions (Casacuberta et al., 2018). Compared to Wefing et al. (2021), the uncertainty for the mid-depth ^{129}I input function was increased by a factor of four. This more conservative approach reflects the higher variability revealed by newly available data from 2018 published in Pérez-Tribouillier et al. (2025) and data from this study (2020 and 2021), which exceeded the range suggested by the initial uncertainty estimate.

255 2.3.2 The Binary Mixing Model

Tracer ages for surface Polar Water I samples collected in 2016, 2020, and 2021 were estimated using a binary mixing model, following the methodology described by Wefing et al. (2021) and Payne et al. (2024). The model assumes mixing between two endmembers: (i) the time-dependent surface input functions of ^{129}I and ^{236}U , driven by the NRPs input to the ocean surface, and (ii) a background concentration of these tracers carried by Pacific and Atlantic waters that have not been in contact with the NRPs signal (see Fig. 1). The background levels were estimated to be 1.16×10^7 at L^{-1} for ^{129}I and 6.25×10^6 at L^{-1} for ^{236}U (Snyder et al., 2010; Chamizo et al., 2022; Dale et al., 2024). These are estimates of the background levels updated from the ones in Wefing et al. (2019). Therefore, the tracer ages were recalculated for the published 2016 dataset to be consistent when comparing the three years (2016, 2020, and 2021) used in this study. Sea ice meltwater and river water has been observed to carry low concentrations of ^{129}I and ^{236}U (Casacuberta et al., 2016, 2018). As these sources are not explicitly represented in the binary mixing model, tracer concentrations of all samples were corrected to a reference salinity of 34.8 to account for dilution effects associated with freshwater inputs. The reference salinity of 34.8 was calculated as the mean salinity in the Arctic Ocean (Aagaard and Carmack, 1989) and has since often been used as a reference (e.g., Karpouzoglou et al., 2022;

270 The salinity corrected concentration c_{sal} is calculated in the following:

$$c_{sal} = \frac{c_{init} \times Sal_{ref}}{Sal_{sample}} \quad (1)$$

where c_{init} is the initial measured concentration, Sal_{sample} is the salinity of each individual sample, and Sal_{ref} is the reference salinity 34.8.

The binary mixing model is illustrated in Fig. 3a, where each line represents the mixing trajectory between the background
 275 concentration and the input function for a specific year. The line closest to a given sample indicates the best-fit input year, thereby defining the tracer age of that sample as the difference between sampling year and input year. An example is shown in Fig. 3a that plots closest to the mixing line from 2000 and therefore corresponds to a tracer age of 16 years, assuming the example was sampled in 2016. The binary mixing model grid with the data from 2016, 2020, and 2021 plotted on top is shown in Fig. A3. Throughout this study, the term “tracer age” refers specifically to estimates derived from the binary mixing model,
 280 while “circulation timescale” is used more broadly to encompass age estimates from all applied methods. The uncertainty of tracer ages was assessed by calculating the minimum and maximum age estimates based on total uncertainties for both ^{129}I and ^{236}U and resulted in relative uncertainties of 4–21 %, 4–85 %, and 4–15 % for 2016, 2020, and 2021, respectively. These elevated upper bounds arise from the youngest tracer ages in the dataset. Towards more recent input years, the input functions of both tracers exhibit weaker temporal dynamics, so that the mixing trajectories of adjacent input years lie closer together. For
 285 young samples, the measurement uncertainty of the tracer concentrations therefore translates into a broader range of possible input years, inflating the relative age uncertainty. This effect is amplified by the low absolute ages, which form the denominator when the uncertainty is expressed in relative terms.

The binary mixing model assumes purely advective transport and therefore neglects the mixing of water parcels that entered
 290 the Arctic Ocean in different years and carry temporally variable tracer signals. Although the surface layer is known to be strongly advective (e.g., Smith et al., 2011), this assumption remains a limitation of the approach. Since mixing is not represented, the resulting tracer age uncertainties likewise exclude any contribution from mixing processes.

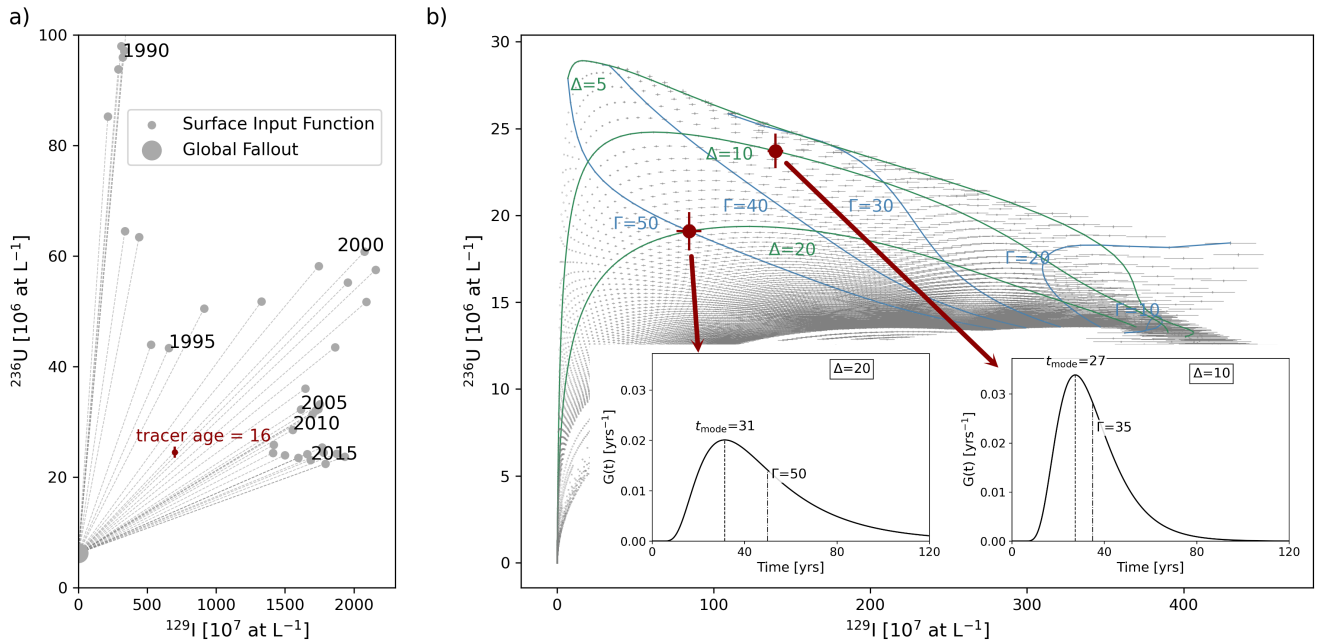


Figure 3. a) The binary mixing model grid with a constant global fallout background level (large gray point) and a time varying input of ^{129}I and ^{236}U to the surface Arctic Ocean (small gray points). An example datapoint from the year 2016 with its uncertainty is shown in red. b) The TTD model grid of ^{129}I and ^{236}U concentrations. Each gray dot represents a possible Γ - Δ combination based on the mid-depth input function with the uncertainties as gray lines. Isolines show the distribution of Γ 's (blue lines) and Δ 's (green lines). Two example datapoints are plotted on top of the grid (red). The insets show the probability density functions of these two examples.

2.3.3 The Transit Time Distribution Model at Mid-Depth

To address this limitation, the TTD model is applied as a complementary method. The TTD model provides a framework for estimating oceanic transport timescales and mixing processes by accounting for multiple pathways contributing to a given sampling location (Haine and Hall, 2002; Waugh et al., 2004; Hall and Haine, 2002). Although the mathematical formulation is one-dimensional, the model effectively incorporates the convergence of multiple water mass branches, making it particularly suitable for complex regions like the Arctic Ocean.

300

The TTD model, originally adapted to the oceanographic case by Haine and Hall (2002), has since been adopted in Arctic tracer studies (e.g., Tanhua et al., 2009; Smith et al., 2011; Wefing et al., 2021; Smith et al., 2022). Following the approach described in Wefing et al. (2021), we applied the TTD model to mid-depth Arctic Atlantic Water samples. We use the so-called “Smith’s TTD approach” (Raimondi et al., 2024), which leverages the transient input of ^{129}I and ^{236}U into the Arctic Ocean (see section 2.3.1) to derive both the age spectrum and degree of mixing for each sample. A brief overview of the model is

305

provided here; for further methodological details, see Wefing et al. (2021) and Raimondi et al. (2024).

The tracer concentration at a given location x and time t is given as:

$$c(x, t) = \int_0^{\infty} c_0(t - t') G(x, t') dt' \quad (2)$$

Here, $c_0(t)$ represents the time-dependent input function, and $G(t')$ is the Green's function, which describes the probability
 310 distribution of transit times. The Green's function is modeled as an inverse Gaussian distribution, capturing both advective
 and diffusive transport components in the ocean (Haine and Hall, 2002). It is interpreted as the probability that a water parcel
 originating at the source arrives at the sampling location x after a time t (Haine et al., 2025):

$$G(x, t) = \sqrt{\frac{\Gamma^3}{4\pi\Delta^2 t^3}} \exp\left(-\frac{\Gamma(t - \Gamma)^2}{4\Delta^2 t}\right) \quad (3)$$

The shape of the distribution is governed by two parameters: Γ , the mean transit time (mean age), and Δ , which characterizes
 the width of the distribution and thus the degree of mixing. A third parameter, the mode age t_{mode} , represents the most probable
 315 transit time and is calculated as:

$$t_{mode} = \frac{1}{\Gamma} (\sqrt{9\Delta^4 + \Gamma^4} - 3\Delta^2) \quad (4)$$

This parameter has been suggested as a more representative measure of lateral transport than the Γ age (Smith et al., 2011;
 Wefing et al., 2021).

In practice, we assumed a range of possible Δ and Γ values (1–2000 years) and constrained their ratio Δ/Γ to lie between
 320 0.1 and 1.8, following Wefing et al. (2021) and Raimondi et al. (2024). This resulted in a grid of possible ^{129}I and ^{236}U
 concentrations (Fig. 3b), where each grid point corresponds to a specific Γ – Δ pair. Sample concentrations were then mapped
 onto this grid, and the closest match determines the best-fit parameters for each sample. Two examples with the corresponding
 probability density functions are shown in Fig. 3b.

325 To estimate uncertainties, we performed a Monte Carlo simulation with 100 iterations. This included propagated uncertain-
 ties from tracer measurements. Additionally, uncertainties in the input function were incorporated into the simulation. These
 are visualized in Fig. 3b as uncertainty bars for the data points and as uncertainty envelopes around the grid points.

The TTD parameters in Arctic Atlantic Water were recalculated for the published 2016 dataset in Wefing et al. (2021) to be
 330 consistent when comparing the three years (2016, 2020, and 2021) used in this study. This resulted in slight differences to the

published dataset due to updated background concentrations of the tracers. The grids with the data plotted on top are shown in Fig. A4 for all years.

2.3.4 The Transit Time Distribution Model at the Surface

In addition to applying the TTD model to mid-depth Arctic Atlantic Water, for the first time here, we extended its use to surface Polar Water I samples. This novel application enables direct comparison of transit times across vertical layers in the Fram Strait and provides a more accurate representation of mixing processes along surface pathways.

Surface waters in the Fram Strait are influenced by a complex mixture of Atlantic Water, Pacific Water, shelf water, and sea ice meltwater. Shelf water, sea ice meltwater and the Atlantic/Pacific component that carry the global fallout tracer signal all dilute the Atlantic Water signal that carries the NRPs signal. To account for these effects, we applied two corrections to the measured ^{129}I and ^{236}U concentrations. First, we corrected for freshwater dilution by normalizing tracer concentrations to the reference salinity of 34.8 (Aagaard and Carmack, 1989), consistent with the correction applied in the binary mixing model (Eq. 1). Second, we accounted for the tracer-free global fallout contribution by subtracting the background tracer levels. Identical to the binary mixing model, the background concentrations of 1.16×10^7 at L^{-1} for ^{129}I and 6.25×10^6 at L^{-1} for ^{236}U were used (Snyder et al., 2010; Chamizo et al., 2022).

The corrected tracer concentrations (c_{corr}) were calculated using the following formula:

$$c_{corr} = \frac{c_{sal} - c_{GF} \times f_{GF}}{f_{Atl}} \quad (5)$$

where c_{sal} is the salinity corrected concentration (Eq. 1), c_{GF} is the global fallout concentration, and f_{GF} and f_{Atl} are the fractional contributions of global fallout water of Pacific and (old) Atlantic origin and the Atlantic Water branch carrying the tracer signals, respectively, with $f_{GF} = 1 - f_{Atl}$.

Both f_{GF} and f_{Atl} from Eq. 5 are not well constrained. We assume that the global fallout contribution (f_{GF}) in surface waters is mainly of Pacific origin. Previous studies based on nutrient analysis reported a wide range of Pacific Water fractions, from 40 % to 80 % between 2002 and 2011, with some years showing values below 10 % (Jones et al., 2008; Dodd et al., 2012). An independent method based on neodymium (Nd) isotopes estimated the Pacific Water fraction to be below 30 % in 2012 (Laukert et al., 2017). In this work, we tested different fractions within this range; the proportions that best fit our data were 15 %, 25 %, and 15 % for 2016, 2020, and 2021, respectively (see Fig. A5, which shows the TTD grid obtained with corrections ranging from 0 % to 30 % in 5 % steps).

We additionally explored an alternative correction method using nutrient-based estimates of Pacific Water fractions for each individual sample, based on data from Graeve and Ludwichowski (2017) for 2016 and Lin et al. (2023a) for 2020 and 2021. This approach produced unrealistic tracer concentrations that fell outside the defined TTD model grid and we therefore excluded it from the final analysis (see Fig. A6). The mean Pacific Water fraction derived from nutrient-based estimates of each year was 36 %, 36 %, and 34 % for Polar Water I in 2016, 2020, and 2021, respectively, which are higher than our best-fit values. This is expected, as this method is known to overestimate Pacific Water fractions due to denitrification in shelf sediments (Bauch et al., 2011; Alkire et al., 2019).

For all years (2016, 2020, 2021), TTD parameters in Polar Water I based on the best-fit fractions were calculated as part of this study. The grids with the data plotted on top are shown in Fig. A4.

3 Results

3.1 Distribution of ^{129}I in the Fram Strait in 2020 and 2021

^{129}I concentrations are represented as depth profiles in Fig. 4, grouped into five main regions along the section. Two general ^{129}I concentration patterns emerged in 2020 and 2021. First, ^{129}I concentrations generally decreased with depth across the entire study area. Notable exceptions to this trend were observed at stations from the Greenland Shelf region, where increases towards the bottom were detected at certain stations in 2020. Second, overall higher ^{129}I concentrations were observed in the 2021 sampling campaign compared to the 2020 campaign across most regions. A summary of the ^{129}I concentration ranges measured in both years across all five regions is provided in Table 2.

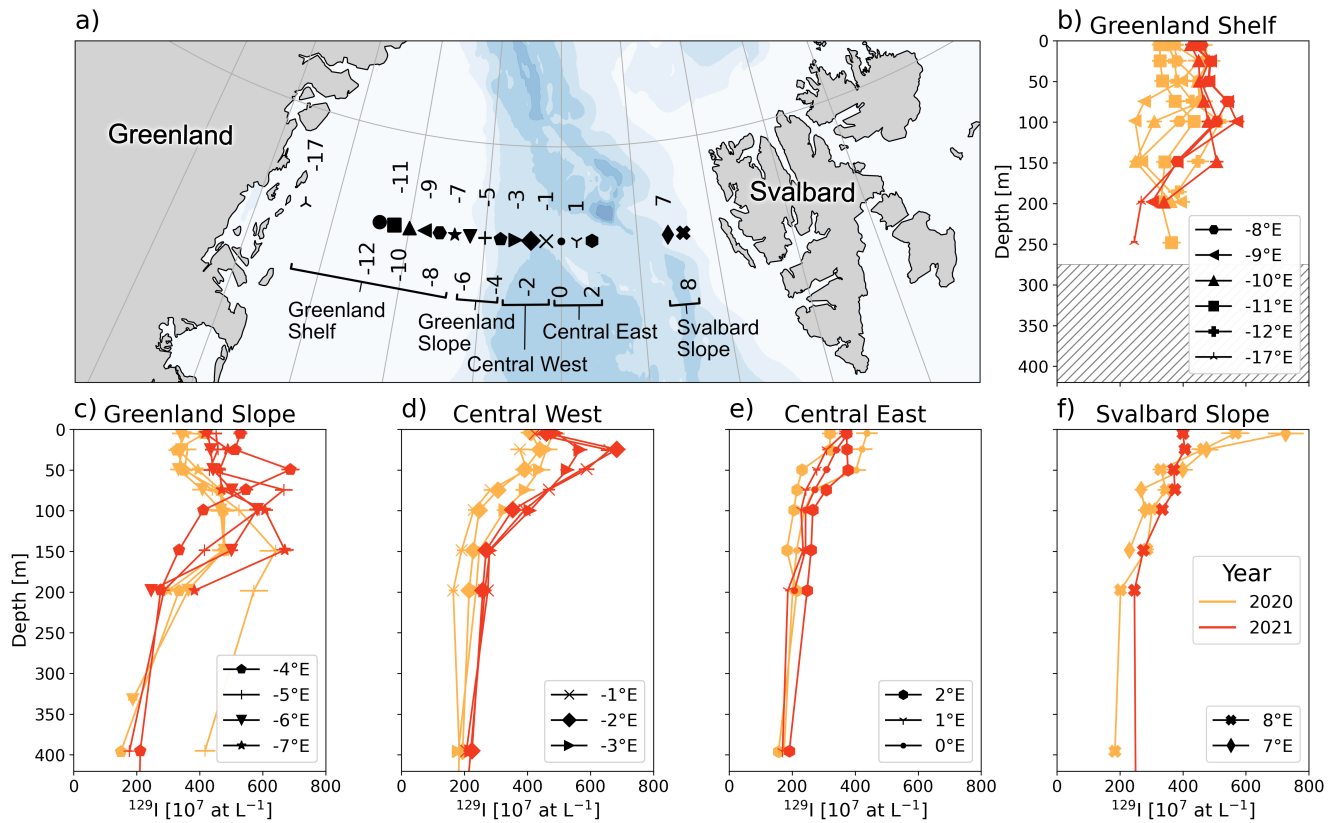


Figure 4. Depth profiles (0–400 m) of ^{129}I concentrations in 2020 (orange) and 2021 (red). The stations are divided into five regions (indicated on the map in panel a): Greenland Shelf (b, -17° E to -7° E), Greenland Slope (c, -6° E to -4° E), Central West (d, -3° E to -1° E), Central East (e, 0° E to 3° E), and Svalbard Slope (f, 7° E to 8° E). The hatched area in panel b) indicates bottom depth.

Table 2. Summary of ^{129}I concentrations measured along the Fram Strait section in 2020 and 2021. For each region and year, the table lists the longitudinal extent, the number of stations, the sampled depth range, and the minimum and maximum ^{129}I concentrations.

Region	Longitude	Year	# Stations	Depth range (m)	^{129}I ($\times 10^7$ at L^{-1})
Greenland Shelf	-17° to -8° E	2020	5	5–200	248 ± 20 to 517 ± 41
		2021	4	5–200	303 ± 13 to 571 ± 24
Greenland Slope	-7° to -4° E	2020	4	5–400	150 ± 12 to 641 ± 50
		2021	4	5–1000	177 ± 7 to 688 ± 29
Central West	-3° to -1° E	2020	3	5–2500	15 ± 2 to 481 ± 38
		2021	3	5–1000	173 ± 10 to 683 ± 29
Central East	0° to 3° E	2020	2	5–400	157 ± 12 to 438 ± 34
		2021	3	5–400	169 ± 7 to 376 ± 16
Svalbard Slope	$7^\circ/8^\circ$ E	2020	2	5–400	184 ± 15 to 727 ± 57
		2021	1	5–1000	21 ± 1 to 405 ± 17

The "Greenland Shelf" region extended from -17° E to -8° E with five stations in 2020 and four stations in 2021 (Fig. 4b). Stations -11° E and -12° E were only sampled in 2020, while station -17° E was only sampled in 2021. At stations -9° E, -10° E, and less pronounced at station -11° E in 2020, we observed a minimum in concentrations around 75–150 m depth, and then an increase again towards the bottom samples. These low ^{129}I concentrations between 75 and 150 m depth coincided with high ^{236}U concentrations reported by Lin et al. (2023a) and are discussed in section 4.4.

The "Greenland Slope" region between -4° E and -7° E was in the core southward flowing EGC and comprised the same four stations in 2020 and 2021 (Fig. 4c). In both years, we observed elevated concentrations below the surface (between 50 and 200 m) and then decreasing to the seafloor. These elevated concentrations were at a deeper depth in 2020 (100–200 m) compared to 2021 (50–100 m).

The "Central West" region spanned from -3° E to -1° E, with three stations in 2020 and the same in 2021 (Fig. 4d). The overall lowest concentration was observed in 2020 at 2500 m depth (-1° E), amounting to $(15 \pm 2) \times 10^7$ at L^{-1} (not shown in Fig. 4d), a depth that was not sampled in 2021. The difference in minimum concentration between the two years therefore reflects the different maximum sampling depths rather than a genuine interannual difference in deep-water concentrations. In 2021, all stations in this region showed a sub-surface peak at around 25 m and overall higher concentrations between 25–200 m compared to 2020, whereas concentrations in surface samples (5 m depth) were similar in both years.

The "Central East" part spanned from 0° E to 3° E with two stations in 2020 and three stations in 2021 (Fig. 4e). Station 2° E was generally lower in ^{129}I concentrations in 2020 compared to 2021, whereas station 0° E showed higher concentrations in

the upper 50 m in 2020 compared to 2021.

The samples from the "Svalbard Slope" region were located in the WSC, transporting Atlantic Water into the Arctic Ocean (Fig. 4f). The region comprised two stations in 2020 (7° E and 8° E), and one station in 2021 (8° E). The highest concentrations
405 were found at the surface, higher in 2020 compared to 2021.

3.2 Distribution of ^{129}I in Water Masses in 2020 and 2021

TS-diagrams with color-coded ^{129}I concentrations as well as section plots of ^{129}I concentrations relate the tracer distribution to water masses in the Fram Strait in 2020 (Fig. 5a,b) and 2021 (Fig. 5c,d). The samples collected in 2020 generally showed
410 elevated ^{129}I concentrations in Polar Water I and Atlantic Water (Fig. 5a). Lowest ^{129}I concentrations were found in Intermediate and Deep Waters. A subset of samples fell outside the Rudels (2022) classification, located between the 27.2 kg m^{-3} and 27.7 kg m^{-3} σ_0 isopycnals and with conservative temperatures above 0° C. These were classified as mixtures of Recirculating Atlantic Water and outflowing waters such as Polar Water I, Polar Water II, or Polar Water I warm (see Table 1). Water masses indicated in the TS diagram are also shown in the cross-section of ^{129}I (Fig. 5b). On the eastern side, Atlantic Water occupied
415 the water column along the Svalbard Slope, representing northward-flowing water entering the Arctic Ocean via the WSC. In the central region, Recirculating Atlantic Water dominated between 50–200 m depth. In the western Fram Strait, where the EGC dominates, Polar Water I and Polar Water II extended from the surface to ~200 m depth, underlain by Arctic Atlantic Water reaching depths of ~1,000 m. Beneath this, Intermediate and Deep Waters were present.

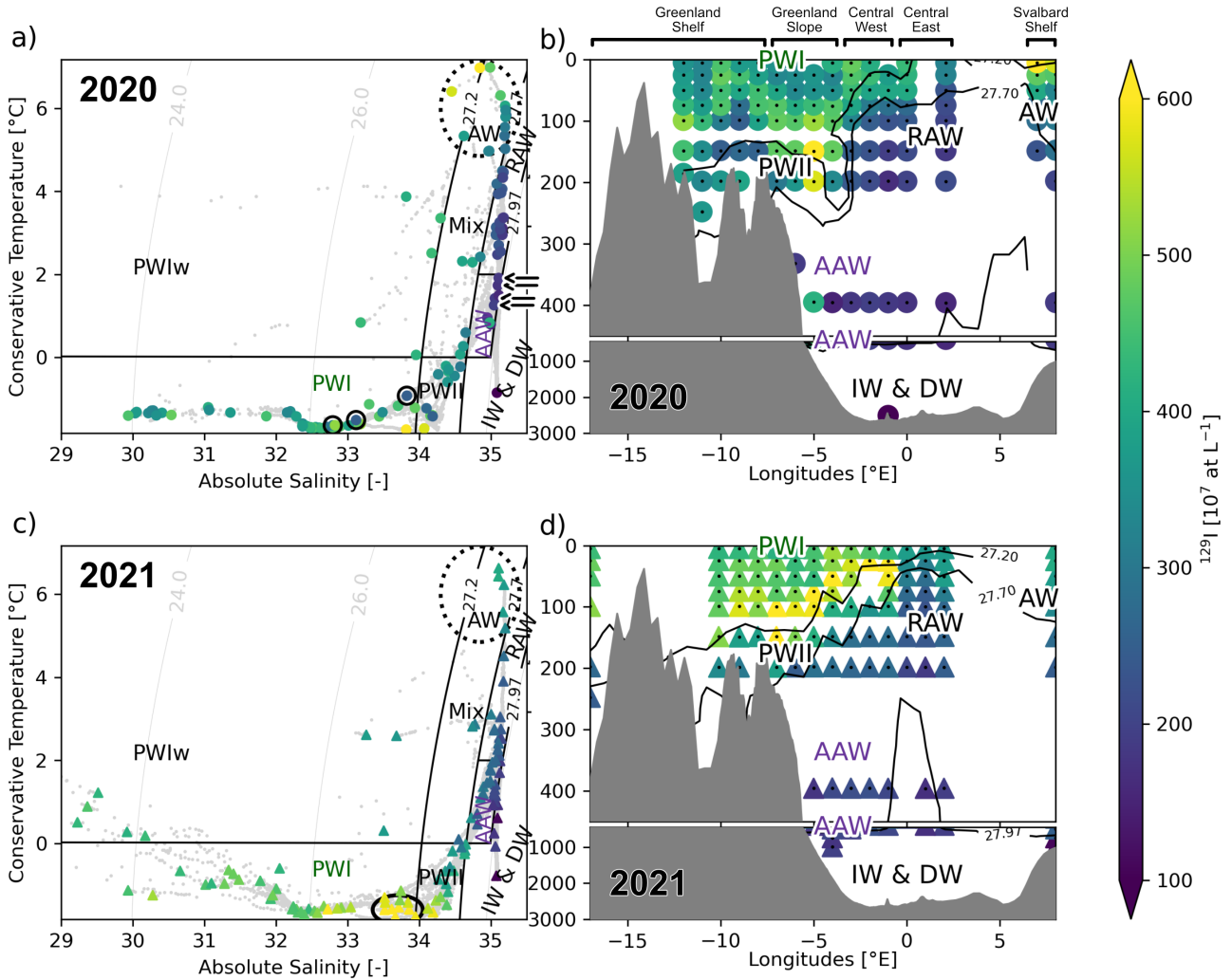


Figure 5. TS-diagrams and section plots of ^{129}I in 2020 (top, a) and b)) and 2021 (bottom, c) and d)). Panels a) and c) show Conservative Temperature vs. Absolute Salinity (TS-diagrams) with σ_0 isopycnals in light gray. Fram Strait 2020 (a) and 2021 (c) samples are plotted in the TS space as gray dots and with ^{129}I concentrations as a color-code where available. The category boundaries from Rudels (2022) (Table 1) are drawn as black lines whereas the dotted ellipse indicates Atlantic Water samples that are part of the WSC. Three black circles in panel a mark relatively low ^{129}I concentrations within Polar Water I, and four arrows indicate Arctic Atlantic Water samples (both discussed in section 4.4). The black ellipse in panel c) highlights elevated ^{129}I concentrations. Panels b) and d) show cross-sections from the Greenland Shelf to Svalbard (-17° E to 8° E) at 78.9° N with color-coded ^{129}I concentrations from 2020 (b) and 2021 (d). Isopycnals (black) for $\sigma_0 = (27.2, 27.7, 27.97)$ are overlaid. Water mass acronyms in panels a to d correspond to the water masses defined in Table 1.

420 In 2021, the ^{129}I concentrations showed overall a similar distribution in TS space and in the cross-sectional view as in 2020 (Fig. 5c,d). However, a distinct cluster of elevated ^{129}I concentrations was observed within Polar Water I near the Polar Water

II boundary in 2021 (black ellipse in Fig. 5c).

The elevated ^{129}I concentrations observed in Polar Water I and Atlantic Water in the TS diagram in both years are also evident in the sectional view. The elevated ^{129}I concentrations in Polar Water I adjacent to Polar Water II appeared as a horizontal band between $\sim 50\text{--}200$ m depth, most clearly in 2021. Surface samples in the central region included Polar Water I warm and mixed water masses, reflecting the dynamic mixing zone where inflowing and outflowing waters converge. Comparing both years, the position and orientation of the water mass front between inflow and outflow differed. In 2020, the front was more vertically aligned, with the $27.2 \text{ kg m}^{-3} \sigma_0$ isopycnal extending straight down near -3° E. In contrast, 2021 showed a more sloped front, with Polar Water I/Polar Water II reaching 0° E at the surface and gradually shifting westwards with depth. These differences were also reflected in the ^{129}I concentrations (Fig. 5b,d).

3.3 Tracer Ages of Surface Waters

Tracer age estimates for Polar Water I were derived from the binary mixing model (section 2.3.2) for the years 2016, 2020, and 2021 and are presented here as depth profiles, separated into two different regions (Fig. 6). The mean tracer age was 18 ± 1 in 2016, 22 ± 2 years in 2020, and 23 ± 1 years in 2021.

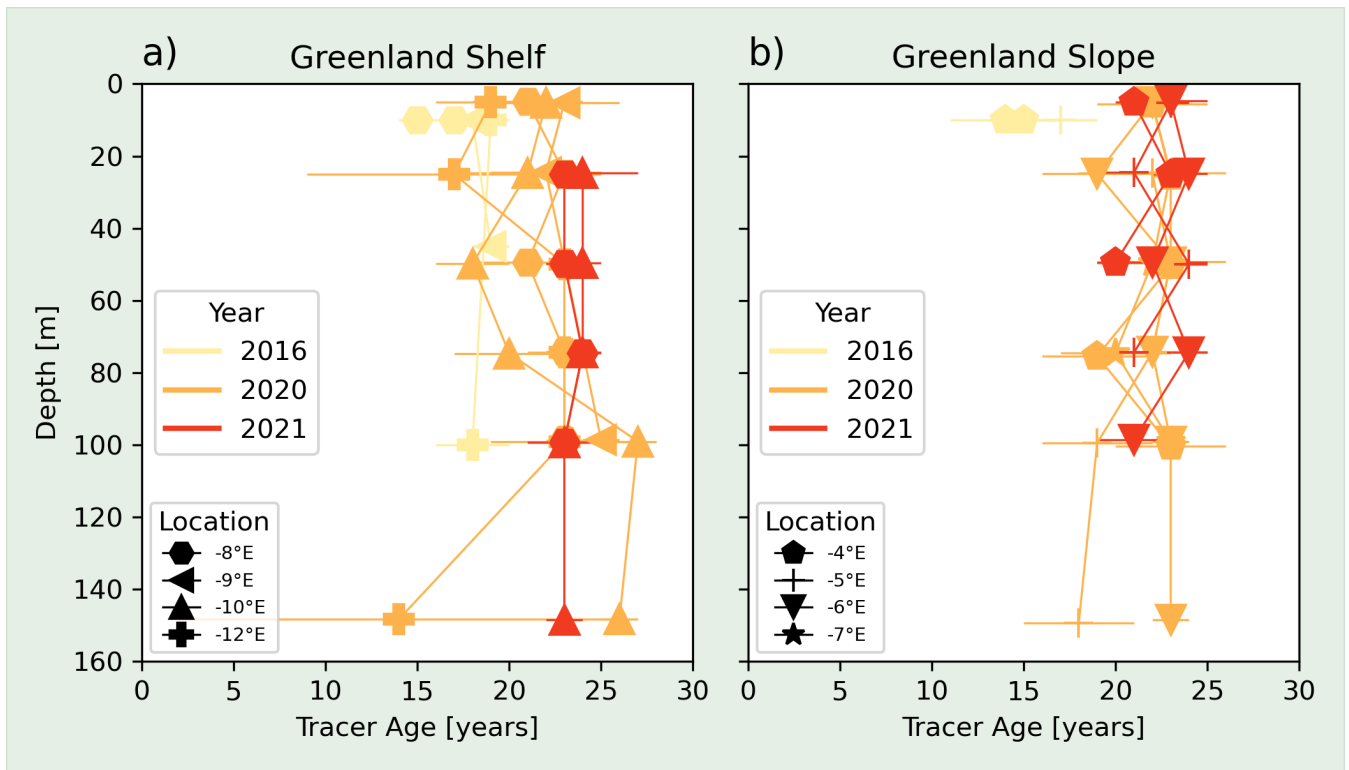


Figure 6. Tracer age results for Polar Water I at the surface (indicated by the green background color), divided in two regions of the Greenland Shelf (a) and the Greenland Slope (b). Tracer ages are shown for the years 2016 (yellow), 2020 (orange), and 2021 (red).

In the Greenland Shelf region (Fig. 6a), tracer ages were almost constant over the entire depth range in 2016 and 2021, but showed a higher variability in 2020, encompassing both the overall lowest (14 ± 12 years, -12° E at 150 m) and highest (27 ± 1 years, -10° E at 100 m) tracer age. Low tracer ages were found at shallower depths at stations -10° E and -12° E and elevated tracer ages at stations -9° E and -10° E at deeper depths. The latter samples were characterized by low ^{129}I concentrations (see section 3.1) and high ^{236}U concentrations. Regarding temporal trends, tracer ages in the upper 50 m in the Greenland Shelf region showed an overall increase from 2016 to 2021.

In the Greenland Slope region (Fig. 6b), no temporal trends were observed between 2020 and 2021, since tracer ages were quite narrowly distributed over the entire depth range in both years, largely between 19 and 24 years. Data from 2016 was only available for two samples at 10 m depth, which showed a lower tracer age compared to the same station and depth in 2020 and 2021. In the Central West region (not shown in Fig. 6), the few available Polar Water I samples had tracer ages around 20 years.

Estimates of the TTD parameters t_{mode} , Γ , and Δ for 2016, 2020 and 2021 are shown as depth profiles in Fig. 7, calculated as described in sections 2.3.3 and 2.3.4, including samples from Arctic Atlantic Water and Polar Water I.

3.4.1 Mid-depth Arctic Atlantic Water

455 The TTD parameters for the mid-depth Arctic Atlantic Water layer are visualized in Fig. 7. The ranges and means of all three TTD parameters for the three years are summarized in Table 3.

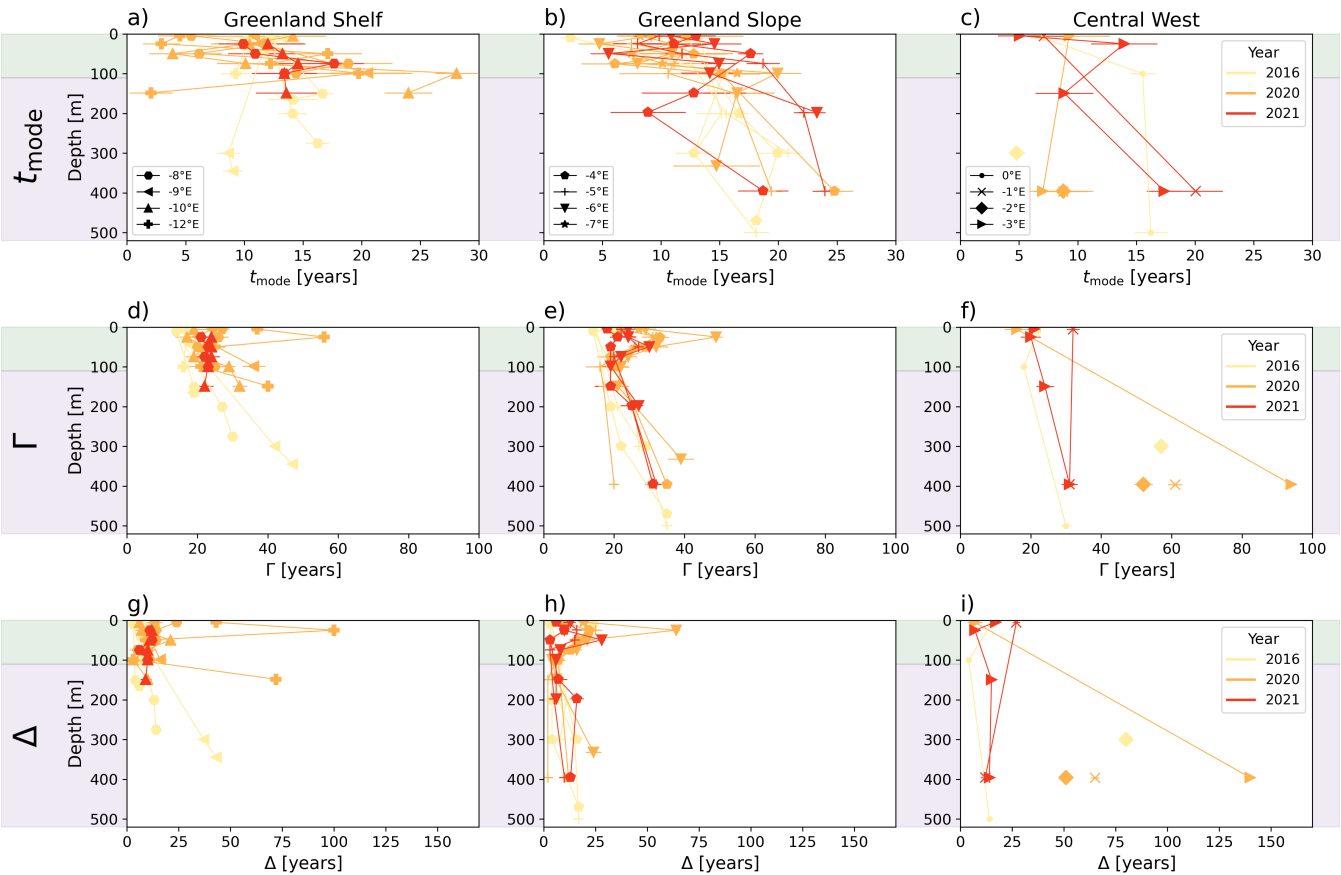


Figure 7. TTD parameters t_{mode} (a-c), Γ (d-f), and Δ (g-i) divided in the three regions of the Greenland Shelf (left), the Greenland Slope (middle), and Central West (right). Results are shown for Polar Water I and Arctic Atlantic Water for the years 2016 (yellow), 2020 (orange), and 2021 (red). The TTD calculations for the surface (green background) followed a different method than at mid-depth (purple background) (see section 2.3.3).

Table 3. Summary of TTD parameters for the mid-depth Arctic Atlantic Water layer in 2016, 2020, and 2021. For each parameter and year, the table lists the range, with the mean and 1σ uncertainty given in brackets. All values are in years.

Year	t_{mode}	Γ	Δ
2016	5 to 21 (14 ± 4)	18 to 57 (30 ± 10)	4 to 80 (18 ± 16)
2020	7 to 25 (13 ± 7)	20 to 94 (54 ± 26)	2 to 140 (56 ± 50)
2021	9 to 24 (17 ± 6)	19 to 32 (27 ± 4)	5 to 16 (11 ± 4)

Due to the limited number of mid-depth samples in 2020 (n=7) and 2021 (n=9), spatial interpretation is constrained, but a general increase of all TTD parameters with depth was observed. The Greenland Slope, associated with the core of the EGC, exhibited the most homogeneous t_{mode} , Γ , and Δ distributions (Fig. 7b, e, h), consistent with reduced influence from other water masses. Notably, samples from the Central West region in 2020 showed lower t_{mode} ages (7–9 years, Fig. 7c) and elevated Δ values (51–140 years, Fig. 7i) than those from the Greenland Slope. This pattern likely reflects the presence of younger inflowing waters, transported by the WSC and returning through Fram Strait (see section 4.2). As these younger waters mix with older waters that have transited the Arctic Ocean, the resulting broadening of the transit-time distribution yields a larger Δ .

3.4.2 Surface Polar Water I

The TTD parameters in the surface Polar Water I layer are visualized in Fig. 7. The ranges and means of all three TTD parameters for the three years are summarized in Table 4.

Table 4. Summary of TTD parameters for the surface Polar Water I layer in 2016, 2020, and 2021. For each parameter and year, the table lists the range, with the mean and 1σ uncertainty given in brackets. All values are in years.

Year	t_{mode}	Γ	Δ
2016	2 to 15 (10 ± 3)	14 to 34 (18 ± 5)	3 to 24 (9 ± 6)
2020	2 to 28 (12 ± 6)	16 to 56 (26 ± 8)	2 to 100 (18 ± 20)
2021	5 to 19 (12 ± 4)	18 to 32 (23 ± 6)	3 to 28 (11 ± 6)

The Greenland Slope showed the most consistent t_{mode} ages throughout the Polar Water I layer (Fig. 7b, green background area), similar to the consistent tracer ages in that region. In contrast, the Greenland Shelf exhibited greater variability, particularly in 2020, with t_{mode} ages increasing with depth (Fig. 7a and Fig. A7a). In 2020, a subsurface minimum was observed in t_{mode} ages at the Greenland Shelf and Slope at around 50 m depth, which corresponded to a maximum in Γ and Δ in the same depth range. Generally, Γ ages and the mixing parameter Δ were highest in 2020 throughout the Polar Water I layer (Fig. A7b,c and Table 4).

Several samples deviated from the general pattern in 2020. On the Greenland Shelf, notably low t_{mode} ages were recorded at station -12° E and elevated t_{mode} ages were observed at stations -9° E and -10° E at depths of 100—150 m, which is discussed in section 4.2. For several depths at station -12° E, as well as for one sample from station -6° E (25 m depth), high values for Γ and Δ were apparent.

4 Discussion

4.1 The Role of Mixing in Polar Surface Water Age Estimations

Circulation timescales in the surface Polar Water layer based on anthropogenic radionuclides have historically been estimated using tracer ages derived from the binary mixing model (Smith et al., 2011; Wefing et al., 2019, 2021; Lin et al., 2023a; Casacuberta and Smith, 2023). Tracer ages fundamentally exclude the mixing of water masses carrying different travel times (see section 2.3.2). In contrast, the TTD approach explicitly incorporates the effect of mixing (see section 2.3.4) and provides more comprehensive metrics like the mean age (Γ), the most probable age (t_{mode}) and the degree of mixing (Δ).

This study represents the first application of the TTD model to the surface layer using anthropogenic radionuclides, with the physical consistency of our results validated by the observed depth-dependency in t_{mode} ages (Fig. 7a-c). We observed significantly lower ages in the surface layer compared to the mid-depth layer—a difference that is mechanistically consistent with the distinct forcing mechanisms of the region: the surface layer is driven by ocean surface wind stress, whereas the mid-depth layer is largely topographically steered and inherently slower (Lique et al., 2015). These findings align closely with Lagrangian particle tracking studies (Lique et al., 2010), which reported a median surface transit of 4–9 years compared to median mid-depth returns of approximately 25 years. This pattern is further corroborated by Dörr et al. (2026), who found surface timescales of roughly 10—20 years and mid-depth timescales frequently exceeding 20 years. Similar depth-dependent timescales have been substantiated by earlier transient tracer studies (^3H , ^3He) (Ostlund and Hut, 1984; Ekwurzel et al., 2001; Pasqualini et al., 2024) and independent modeling (Karcher and Oberhuber, 2002; Popova et al., 2013; Pemberton et al., 2014). The robust agreement with historical data reinforces confidence in our application of the TTD model to the surface layer and the model’s general ability to capture the fundamental dynamics of the Fram Strait outflow.

To unify the concept of circulation timescales across both the surface and mid-depth layers, we compared the results of the binary mixing model and the TTD model (Fig. 8). Across all three years, a general correspondence was observed between the tracer age and the TTD parameters (t_{mode} , Γ). Crucially, tracer ages agree well with t_{mode} and Γ when mixing is minimal (i.e., low Δ values). This aligns with the finding of Smith et al. (2011), who observed good agreement between tracer ages and t_{mode} for samples exhibiting advective characteristics at mid-depth. The three oldest samples in 2020 (at -9° E / -10° E) showed good agreement between tracer ages and t_{mode} (highlighted in red in Fig. 8a), suggesting that this older water parcel (discussed in section 4.2) experienced little mixing (Δ between 3 and 16 years, with the lowest Δ showing the best fit). Under

510 these advective conditions, the binary mixing model provides a realistic approximation of circulation timescales.

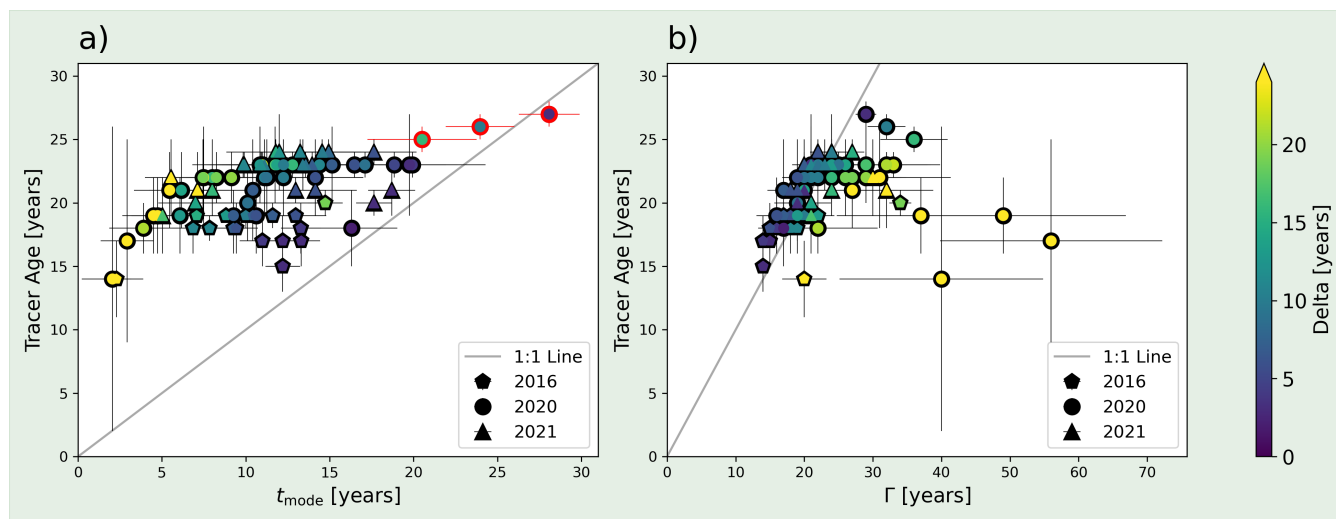


Figure 8. Scatter plot comparing surface Polar Water I (indicated by green background) tracer ages derived from the binary mixing model with t_{mode} (a) and Γ (b; note different x-axis limits) obtained from the TTD model for the years 2016, 2020, and 2021. Data points are color-coded according to their corresponding Δ values from the TTD model, representing the degree of mixing. The gray diagonal line indicates the 1:1 relationship between the two age estimates. Samples highlighted in red are further discussed in the text.

Systematic deviations from the 1:1 line (Fig. 8) correlate with elevated Δ values, confirming that mixing drives model divergence. Specifically, tracer ages underestimate Γ but overestimate t_{mode} . This bias stems from the mixing-induced elongation of the $G(t)$ tail in the probability density function (see examples in Fig. 3b with Δ 's of 10 and 20 years): tracer ages are less sensitive to the presence of older water parcels than the Γ ages (Haine and Hall, 2002), yet are more influenced by it than the t_{mode} (Vaugh et al., 2002, 2003). Although previous studies, such as Smith et al. (2011), argued that tracer ages are justified at the surface due to its predominantly advective flow and expected narrow $G(t)$ distribution, our results show that mixing (Δ) is not negligible in Polar Water in the Fram Strait, and therefore tracer ages deviate from t_{mode} . This highlights a fundamental limitation: tracer age profiles appear more uniform than the TTD parameters (compare Fig. 6 to Fig. 7), but this uniformity is misleading. Since water parcels in the Fram Strait are ultimately the result of multiple transit times due to the merging of various branches of different origin, the TTD model, which explicitly accounts for this range of possible transit times, provides a better representation of the complex ocean transport processes. As sea ice retreats, more momentum will be transferred to the ocean, leading to more turbulence and increased mixing (Morison et al., 1985; Martin et al., 2014; Muilwijk et al., 2024; Brown et al., 2025). Therefore, it will become increasingly important to account for mixing when calculating surface layer circulation timescales across the Arctic Ocean and Fram Strait.

Apart from the limitation of the binary mixing model to capture circulation timescales when mixing is large, Waugh et al. (2003) showed that changes in tracer ages do not necessarily imply temporal changes in the circulation. This is because the calculated tracer ages are not fundamental timescales of the flow due to the transient nature of the circulation (Haine and Hall, 2002; Waugh et al., 2003). As also pointed out by Smith et al. (2011), binary mixing models and their related tracer ages are inherently limited in their ability to capture temporal dynamics. This was recently reinforced by Kumamoto et al. (2024), who related an observed increase in tracer ages to model limitations rather than actual longer circulation timescales by comparing the results to TTD estimates. Since the TTD model captures that each water parcel is ultimately the result of a distribution of multiple transit times (Haine and Hall, 2002; Waugh et al., 2003), we suggest that TTD parameters are the superior tool to characterize the flow field and its temporal evolution. Therefore, the subsequent discussions on the origin of water masses (section 4.2) and on the temporal dynamics of circulation (section 4.5) will rely exclusively on the TTD parameters.

4.2 Variability of Water Mass Origin in Polar Water between 2016 and 2021

Unraveling the provenance of Polar Water outflowing through Fram Strait is essential for understanding upstream dynamics and the freshwater export. To this end, we contextualize the ^{129}I and ^{236}U concentrations observed in Polar Water I in 2016, 2020, and 2021 against upstream tracer signatures from distinct Arctic Ocean regions (Fig. 9).

The Polar Water I outflow through Fram Strait reflects a composite signal originating both from the Amerasian and Eurasian Basins. In the Amerasian Basin, Pacific Water entering through the Bering Strait carries predominantly the global fallout tracer signature and mixes with Atlantic-derived waters originating from the Eurasian Basin. These Atlantic-derived waters transport diluted signals from NRPs, and their contribution to the Polar Water in the Amerasian basin remains minor ($< 5\%$) (Payne et al., 2024), as illustrated in Fig. 9 (black markers with orange contours). In contrast, waters entering with the WSC, carry a tracer signature shaped by a mixture of Atlantic Water and NCC water. Part of this inflow recirculates within Fram Strait as Recirculating Atlantic Water and subducts beneath Polar Water due to its higher density and joins the EGC at intermediate depths (Hattermann et al., 2016; von Appen et al., 2016; Hofmann et al., 2021). Therefore, we do not consider this inflow to be an influencing factor of the observed signal in Polar Water, apart from samples located close to the front between outflowing Polar Water and inflowing WSC waters (approximately east of the prime meridian). In the Eurasian Basin, Polar Waters do not reflect a pure NCC signature but rather a diluted signal resulting from mixing with Atlantic waters (originating from the North Atlantic Current) that carry global fallout tracer concentrations. These upstream characteristics (black markers with green contours in Fig. 9) exhibit pronounced temporal variability between 2011 and 2021, driven by the transient nature of the tracer input functions (Fig. 2). A similar temporal evolution is observed in the Amerasian Basin (2015–2021), along with a spatial gradient marked by increasing ^{129}I concentrations toward Fram Strait. This pattern reflects the progressive influence of Eurasian Basin waters mixing into the Makarov Basin (Fig. 9).

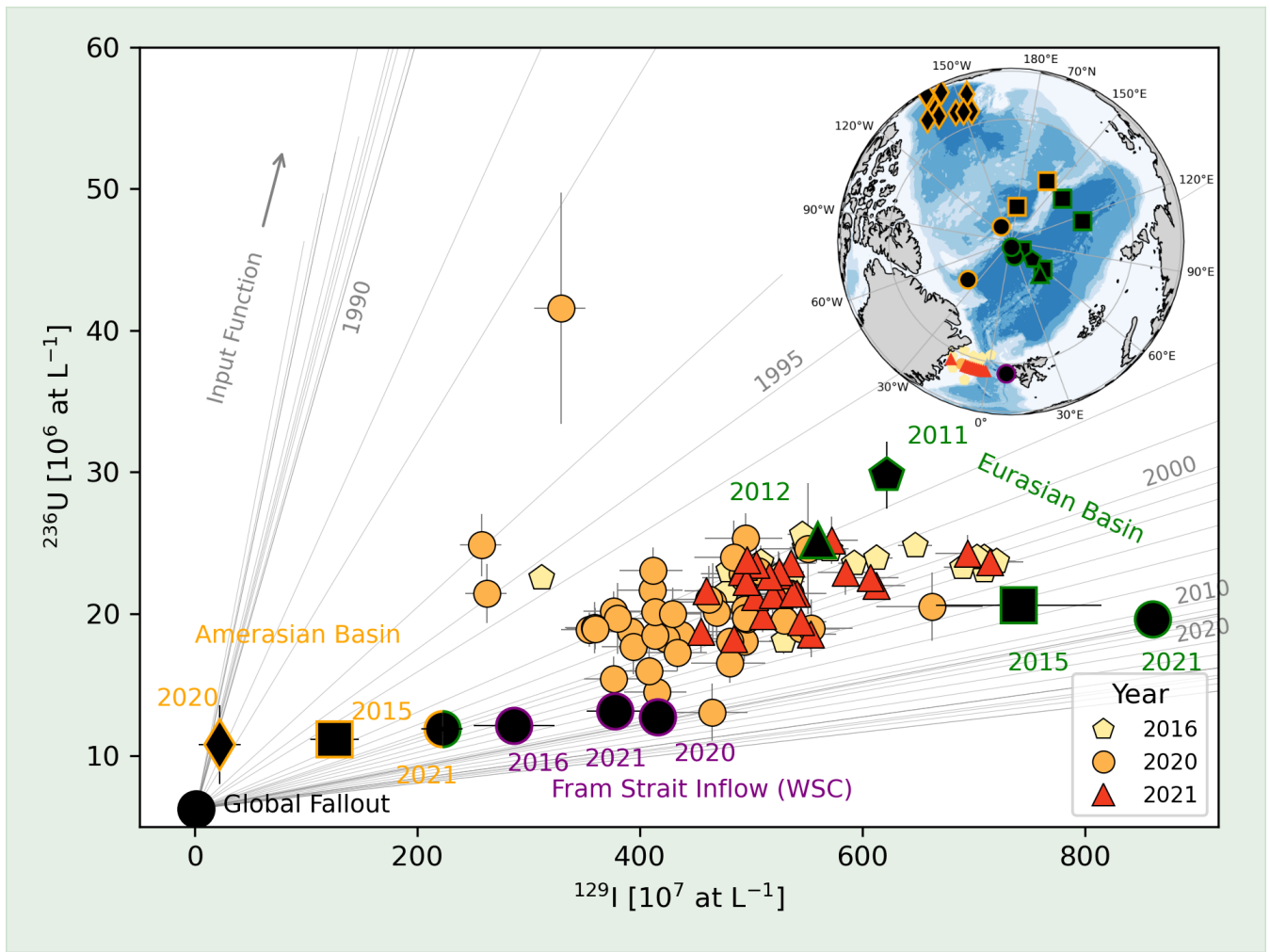


Figure 9. Polar Water I samples from 2016 (yellow), 2020 (orange), and 2021 (red) are displayed in the ^{129}I against ^{236}U tracer space. Previous measurements with the same densities as Polar Water I from 2011, 2012, 2015, 2020, and 2021 are shown as yearly averaged concentrations separately for Amerasian and Eurasian stations. Amerasian-origin waters (orange outline, station locations indicated in the inset map) are represented by samples taken from the Canadian and Makarov Basins, sampled during different expeditions in 2015 (stations 96, 101, and 134 from Casacuberta et al. (2018), squares), 2020 (all stations from Payne et al. (2024), diamonds), and 2021 (stations 28 and 46 from Wefing et al. (2025), circles). Eurasian-origin waters (green outline, station locations indicated in the inset map) are represented by stations in the Amundsen and Nansen Basins, sampled in 2011 (station 212 from Casacuberta et al. (2016), pentagon), 2012 (station 378 from Casacuberta et al. (2016), triangle), 2015 (stations 68, 81, 117, and 125 from Casacuberta et al. (2018), squares), and 2021 (stations 16 and 20 from Wefing et al. (2025), circles). The global fallout tracer signature is indicated as a black circle ($1.16 \times 10^7 \text{ at L}^{-1}$ for ^{129}I and $6.25 \times 10^6 \text{ at L}^{-1}$ for ^{236}U (Snyder et al., 2010; Chamizo et al., 2022)). Inflowing WSC waters (purple outline) are represented as averaged tracer signatures of the upper 100 m at $7^\circ \text{ E} / 8^\circ \text{ E}$ in 2016, 2020, and 2021. The surface NCC input function is shown as grey lines with some specific years for orientation. The inset map shows the geographical location of the samples and upstream tracer signatures.

560 The Polar Water I samples from 2016, 2020, and 2021 fall predominantly between the Amerasian and Eurasian tracer signatures, indicating that the outflow represents a mixture of these two sources (Fig. 9). The relative contribution of these two basins, however, vary interannually. In 2016 and 2021, Polar Water I samples cluster closer to the Eurasian-origin stations, indicating a dominant Eurasian influence. This finding is consistent with Wefing et al. (2022) and is further corroborated by Karpouzoglou et al. (2022), who observed relatively low freshwater transport through Fram Strait in 2016, consistent with a Eurasian Basin source given that Amerasian-origin waters typically carry a more pronounced freshwater signature derived from Pacific inflow and the Beaufort Gyre. In contrast, 2020 exhibits a marked shift toward the Amerasian-origin endmember which is also reflected in a higher Pacific fraction that we use in the surface TTD calculations (see section 2.3.4). We exclude local dilution by sea-ice melt or meteoric water as the primary driver of this shift, as all samples were salinity-normalized (to $Sal_{ref} = 34.8$, see section 2.3.2). Instead, this anomaly aligns with the trend reported by Wefing et al. (2022), who attributed a shift in tracer signals between 2016 and 2018/2019 to an increasing Amerasian contribution. Our data suggest that this trend continued into 2020, likely linked to the post-2019 release of freshwater accumulated in the Beaufort Gyre (Wang et al., 2024a), a primary reservoir of Amerasian water that showed an increasing freshwater content in the 2000s and 2010s (Giles et al., 2012; Proshutinsky et al., 2019; Lin et al., 2023b; Wang et al., 2024b). Our observation is also consistent with the decadal-mean reconstruction of Planat et al. (2025), who find the upper-layer outflow at Fram Strait over 2005–2017 (a period that includes our 2016 sampling) to be predominantly of Eurasian origin.

Distinguishing between Atlantic- and Pacific-origin waters in the surface Arctic Ocean is complex and requires multiple complementary approaches. While nutrient-based tracers have historically been used to track Pacific Water variability in Fram Strait (Falck et al., 2005; Jones et al., 2008; Dodd et al., 2012), their conservative behavior has been questioned in several studies (Bauch et al., 2011; Alkire et al., 2015, 2019). In this context, ^{129}I provides an additional, independent constraint, owing to the approximately eight-fold concentration contrast between Amerasian surface waters, which reflect the low-level global fallout signature from the Pacific, and Eurasian surface waters, which carry a strong Atlantic ^{129}I signal. Future assessments of water mass origin in Fram Strait would benefit from a comprehensive multi-tracer approach. Combining ^{129}I and ^{236}U with other parameters such as CDOM, oxygen-18 ($\delta^{18}\text{O}$), neon (Ne), and Nd could significantly improve source disentanglement (Dodd et al., 2012; Lin et al., 2023a; Heuzé et al., 2023; Pérez-Tribouillier et al., 2025). For instance, Nd has already been used to distinguish Pacific from Atlantic waters in both the Fram Strait (Laukert et al., 2017) and the central Arctic Ocean (Paffrath et al., 2021). Gallium has also been investigated as a water mass tracer in the central Arctic Ocean (McAlister and Orians, 2015; Whitmore et al., 2020). Ultimately, integrating ^{129}I and ^{236}U with these tracers (especially CDOM, $\delta^{18}\text{O}$, Ne, and Nd) would provide a more robust framework for delineating freshwater sources than the salinity normalization method employed here (Dodd et al., 2012; Granskog et al., 2012; Pérez-Tribouillier et al., 2025).

4.3 Divergence from Advective Flow in 2020: Enhanced Mixing and Age Broadening Across the Water Column

The shift in water mass provenance towards a higher contribution of Amerasian Basin waters in Polar Water I in 2020 is coupled with an increase in mixing, as revealed by the TTD parameters (Fig. 10a-d). Similar temporal trends are observed for samples from the mid-depth Arctic Atlantic Water (Fig. 10e-h), despite the limited sampling resolution. The year 2020 is characterized by a substantial increase in the mixing parameter Δ (Fig. 10a and e), as well as the mean age Γ (Fig. 10b and f). This indicates that the Amerasian-influenced water parcels in Polar Water I as well as the Arctic Atlantic Water experienced a more complex transit history with a broader range of ages compared to the more coherent, advective flow of 2016 and 2021. This interpretation is reinforced by elevated Δ/Γ ratios in 2020 (Polar Water I mean: 0.6, Arctic Atlantic Water mean: 0.8; see Fig. 10c and g), pointing to less advective transport during this period. Finally, the 2021 reversal to Eurasian-dominated source waters is consistent with a return toward the cyclonic circulation regime that Smith et al. (2021) link to a positive Arctic Oscillation. They document the opposite transition between the mid-1990s and 2015 (from a cyclonic regime to an anticyclonic one) and flagged the winter 2020 Arctic Oscillation as anomalously high. Since the cyclonic mode lags the Arctic Oscillation by roughly one year (Morison et al., 2021), we interpret the high Arctic Oscillation in 2020 as driving a return to cyclonic conditions, which correspondingly favors the enhanced presence of Eurasian-dominated source waters seen in 2021. Morison et al. (2021) further show that this mode reaches intermediate depths, consistent with the 2020–2021 changes we observe in the mid-depth Arctic Atlantic Water layer.

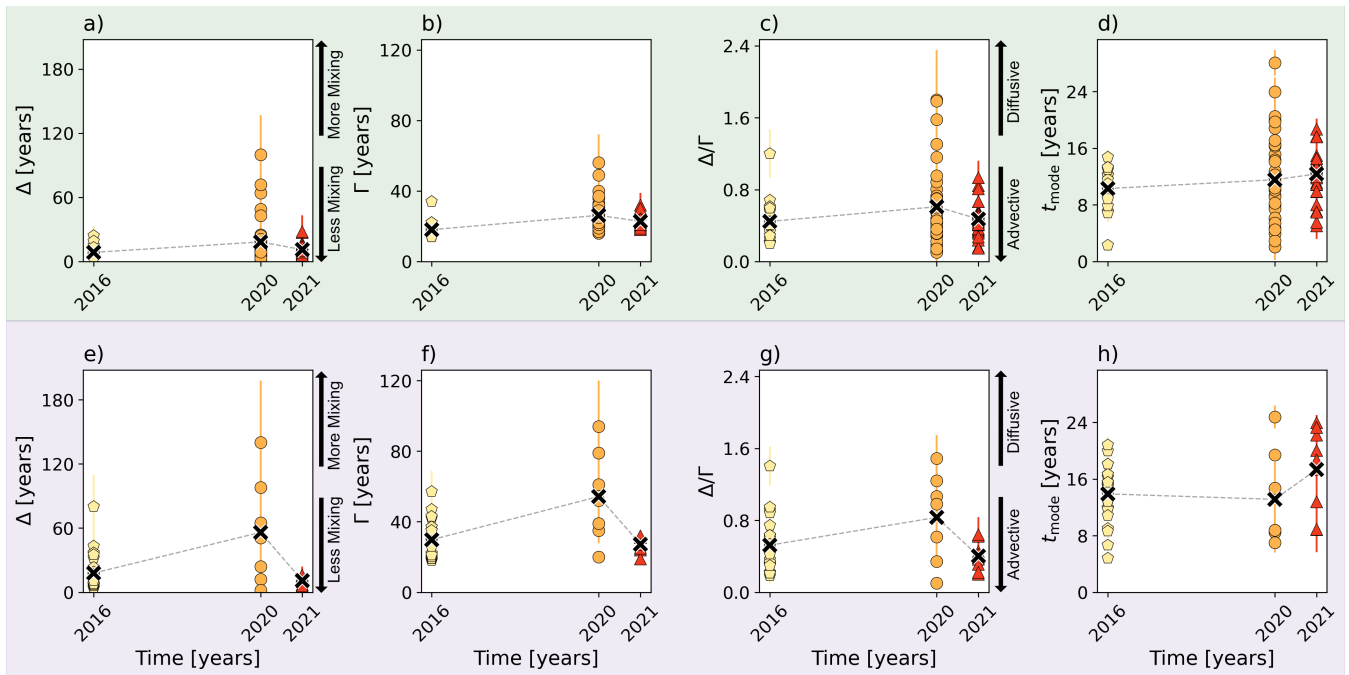


Figure 10. The results of the TTD model are shown for the Polar Water I surface layer (a-d, green background) and Arctic Atlantic Water mid-depth layer (e-h, purple background) for three years, 2016 (yellow), 2020 (orange), and 2021 (red). Four TTD parameters are shown: Δ (a,e), Γ (b,f), Δ/Γ (c,g), and t_{mode} (d,h). The black crosses indicate the respective average values. The gray dashed lines connect successive averages to guide the eye and indicate the temporal evolution.

4.4 Convergence of Distinct Water Masses in 2020: Identifying Canada Basin and Shelf Signatures

610 The 2020 dataset features both highest and lowest ^{236}U concentrations measured across all years discussed in this study and a larger range of ^{129}I concentrations compared to 2016 and 2021 (Fig. 9). This is also reflected in the large range of the TTD parameters obtained for 2020, both for Polar Water I and Arctic Atlantic Water (Fig. 10). A more detailed analysis of the highly heterogeneous 2020 dataset reveals substructures within the outflow, highlighting the complexity of the region.

615 Three spatially clustered samples in the Fram Strait from 2020 (at -9° E to -10° E , 100–150 m depth) exhibited a distinctive combination of high ^{236}U and low ^{129}I concentrations (Fig. 9). Comparison with the time-dependent input function (Fig. 2) reveals that such elevated ^{236}U levels are associated with old source waters. This signature identifies a distinct water parcel with a significantly greater age (tracer age: 25–27 years; t_{mode} : 21–28 years) than the surrounding water. We attribute this to a parcel that followed a longer circulation route through the Amerasian Basin.

620

Conversely, the samples with the lowest ^{236}U concentrations in 2020 ($13 \pm 2 \times 10^6$ at L^{-1}), together with relatively low ^{129}I concentrations (around 400×10^7 at L^{-1}), likely reflect waters that recirculated on the Greenland Shelf, potentially within the

East Greenland Coastal Current, where they underwent dilution. This interpretation is supported by their geographical location on the shelf (primarily at -12° E). Eddy transport of Recirculating Atlantic Water onto the shelf, a mechanism described by Hattermann et al. (2016), offers an alternative explanation, but such water typically subducts beneath the cold, fresh Polar Water. The absence of both the elevated temperatures characteristic of this water mass and the low salinities indicative of fresh meltwater further supports the shelf recirculation hypothesis (circled in Fig. 5a). As station -12° E was not sampled during the 2021 expedition (see Fig. 4), the apparent absence of this low-concentration signal in 2021 might result from spatial sampling bias rather than the temporal disappearance of this shelf feature.

630

Low t_{mode} ages obtained for Arctic Atlantic Water samples in central Fram Strait in 2020 were coupled with high Δ values (Fig. 7c,i and Fig. 8a) and point to mixing with Recirculating Atlantic Water, carrying very young tracer signals. These findings are also supported by the distribution in TS-space (black arrows in Fig. 5a).

635 Large ranges in t_{mode} ages observed within the surface Polar Water I (2–28 years) and mid-depth Arctic Atlantic Water (5–25 years) masses (Fig. 10d,h) underscore the convergence of different water masses from the central Arctic Ocean and the Nordic Seas, the latter recirculating in Fram Strait as Recirculating Atlantic Water. Rather than a unimodal approach, this complex regime is likely better represented by bimodal or multimodal TTDs capable of capturing multiple peaks from mixing water masses. Such approaches have been increasingly advocated for other regions (Haine and Hall, 2002; Peacock and Maltrud, 2006; Chouksey et al., 2022) and were recently applied to ^{129}I and ^{236}U to distinguish branches from the central Arctic Ocean and Canada Basin (Wefing et al., 2025). Within the Fram Strait, the additional influence of the WSC suggests that a trimodal TTD structure may be necessary to fully resolve the regional circulation. However, this is beyond the scope of this study and should be further elaborated in future work with more data available to resolve the different contributions.

645 4.5 Drivers of the 2021 Mode Age increase: Assessing Evidence for Circulation Slowdown and Path Extension

To address the temporal variability of circulation timescales between 2016 and 2021, we use the most probable age (t_{mode}), which serves as a robust metric for assessing the temporal evolution of advective circulation timescales, as already pointed out by Wefing et al. (2021).

650 The data indicate a tendency toward increased circulation timescales and/or longer pathways in 2021 (Fig. 10d,h) for both the surface and the mid-depth Atlantic layer. The mean t_{mode} remained relatively constant from 2016 to 2020 and increased in 2021 in both layers, with the increase more pronounced at mid-depth. This tendency is also apparent at the upper end of the mid-depth age distribution (Fig. 10h). In 2016, when sampling was densest, t_{mode} values clustered between about 10 and 20 years and rarely exceeded 20 years. In 2021, by contrast, several samples exceeded 20 years even though far fewer samples were collected ($n=9$), indicating that the oldest mid-depth waters became more prominent. Taken together, these observations are consistent with a gradual deceleration of the circulation, a lengthening of the transport pathways reaching the Fram Strait,

or both.

The t_{mode} results at the surface depend on the best-fit global fallout correction applied in this study (see section 2.3.4).
660 To test the robustness of the elevated t_{mode} values observed in 2021, we performed a sensitivity analysis using varying fall-out fractions (Fig. A8). When the same fractional correction is applied across all years, this trend is consistently observed (Fig. A8a). Notably, the surface results shown in Fig. 10 were derived using a higher fractional correction for 2020, which actually dampens this effect compared to applying a uniform fractional correction across all years. However, when fractions are varied independently between years, the tendency toward elevated t_{mode} values in 2021 diminishes for some combinations.
665 For example, applying a higher fraction of 30 % in 2020 together with a lower fraction of 10 % in 2021 yields the same mean t_{mode} for both years. This is illustrated by the minimum and maximum envelopes in Fig. A8b. These envelopes nonetheless show that, in most cases, the mean t_{mode} still increases from 2020 to 2021. This high sensitivity to the assumed global fallout fraction should be considered in future transit-time estimates, and results from surface TTD analyses should be interpreted with corresponding caution.

670

The observation of increasing circulation timescales and/or longer pathways in the surface layer presents a complex picture when viewed against the literature. Our results for Polar Water I are in line with the elevated circulation timescales reported for Fram Strait in 2021 by Lin et al. (2023a) and with a weaker upper-layer EGC (Karpouzoglou et al., 2022). Mechanistically, the increase toward 2021 may be linked to the transient period of reduced flow in the Siberian Arctic Ocean between late
675 2015 and early 2018 (Polyakov et al., 2025a). The 2016 samples were collected too soon after the onset of this anomaly to record it, whereas the lagged signal would likely reach the Fram Strait region by 2020 and 2021. This weak local slowdown nevertheless contrasts with reports that ongoing sea-ice decline increases ocean surface stress and thereby strengthens Arctic geostrophic currents (Meredith et al., 2019). Such an intensification has been inferred from satellite altimetry over 2003–2014 (Armitage et al., 2017), detected as increased near-surface currents and shear in eastern Eurasian Basin moorings (Polyakov
680 et al., 2020b), and projected by climate models (Muilwijk et al., 2024). These signals are, however, both regionally confined and partly transient: the Beaufort Gyre acceleration relaxed toward its 2003–2006 level after 2011 (Armitage et al., 2017), and the mooring-based intensification is dominated by the semidiurnal band, with little change in the mean along-slope transport (Polyakov et al., 2020b). A further explanation for the increased transit times observed in 2021 is a shift toward longer, more circuitous pathways, consistent with the pronounced variability of Arctic Ocean surface trajectories reported by Wilson et al.
685 (2021).

The findings of elevated t_{mode} ages in the mid-depth layer in 2021 are well-supported by independent evidence of a recent circulation slowdown. Our results align with reports of increased circulation timescales in the central Arctic Ocean between 2015 and 2021 (Wefing et al., 2025), increased mean ages attributed to a weakening boundary current from 2005 to 2021 (Gerke et al., 2024), and a reported tendency towards increased circulation timescales in the Fram Strait in 2021 (Lin et al.,
690 2023a). Future observations should reveal a reversal toward decreasing circulation timescales, consistent with recent evidence

of a strengthening Arctic overturning circulation (Årthun et al., 2025a).

5 Conclusions and Outlook

This study used the combination of anthropogenic radionuclide transient tracers ^{129}I and ^{236}U to investigate the origin, degree
695 of mixing, and circulation timescales of water masses exiting the Arctic Ocean through Fram Strait between years 2016 and
2021. Two models were applied to obtain estimates of circulation timescales: the binary mixing model resulting in tracer ages
(only applied to surface Polar Water I samples) and the TTD model resulting in mean (Γ) and mode (t_{mode}) ages as well as a
mixing parameter (Δ). This model has previously only been applied to radionuclide data from Arctic Atlantic Water and was
here for the first time also applied to samples from Polar Water I, after correcting for the dilution with the global fallout signal
700 from waters of Pacific origin.

The relationships between tracer ages derived from the binary mixing model and Γ and t_{mode} ages from the TTD model
were validated for Polar Water I samples exhibiting a strong advective component (low Δ values). However, several Polar
Water I samples were dominated by mixing processes (high Δ values) and the ages from the two models did not agree. Our
705 findings therefore suggest that the transport in the surface is not as advective as previously proposed (Smith et al., 2011) and
we emphasize that the TTD model, which accounts for a distribution of transit times due to mixing within the flow, is also
preferred at the surface to accurately characterize the flow field.

The tracer data revealed significant interannual variability in water mass properties. Specifically, the year 2020 was charac-
710 terized by a higher proportion of Amerasian Basin waters in the surface Polar Water layer and by a notably higher degree of
mixing throughout the water column compared to 2016 and 2021. Our observations also suggest that the central Fram Strait
experienced a substantial influence from Recirculating Atlantic Water from the WSC in 2020. Additionally, a small cluster of
three water samples on the Greenland Shelf region exhibited old t_{mode} ages (21, 24, and 28 years) consistent with an origin
in the Canada Basin. The distinct tracer signature, a low ^{129}I concentration coupled with a high ^{236}U concentration, proved
715 highly effective in identifying these small-scale substructures within the water masses, underscoring the usefulness of ^{129}I -
 ^{236}U as a tracer pair to delineate water mass origins. Finally, the TTD model suggested increased circulation timescales (t_{mode}
ages) for both the surface and mid-depth layers in 2021. The dataset, however, does not permit assessment of a long-term trend.

To address these complex temporal dynamics, further tracer-based monitoring at key locations like the Fram Strait is crucial
720 for estimating mean flow developments and complementing the high temporal resolution of current meter observations at spe-
cific locations, as highlighted by Pasqualini et al. (2024). A comprehensive multi-tracer study, incorporating ^{129}I , ^{236}U , CFCs,
 SF_6 , and tritium, would significantly reduce the uncertainties in circulation timescales. Since different tracer combinations
respond uniquely to boundary conditions and source histories (Haine and Hall, 2002; Waugh et al., 2003), integrating them

would provide a more robust estimate of transit times, independent of the source history of any single tracer. To refine the application of the TTD model to surface samples, tracer data from upstream regions of Fram Strait would be valuable to better constrain the fraction of Pacific-origin water entering the study region. Overall, strategic and synergistic tracer sampling at the Fram Strait will significantly enhance our ability to monitor, understand, and improve predictive capabilities for the Arctic Ocean circulation regime under a warming climate.

6 Open Research

Data availability.

The CTD data for 2016 was taken from Kanzow and Rohardt (2017).

The CTD data for 2020 and 2021 was taken from Dodd et al. (2022a, b).

The ^{129}I and ^{236}U data for 2016 was taken from Wefing et al. (2019).

The ^{236}U data for 2020 and 2021 was taken from Lin et al. (2023a).

A subset of the ^{129}I data from 2021 has been published in Pérez-Tribouillier et al. (2025)

The ^{129}I data for 2020 and 2021 is available on Zenodo (Scheiwiller, 2026).

Appendix A: Supporting Figures

A1 Intercomparison of ^{236}U between ETH Zurich and DTU

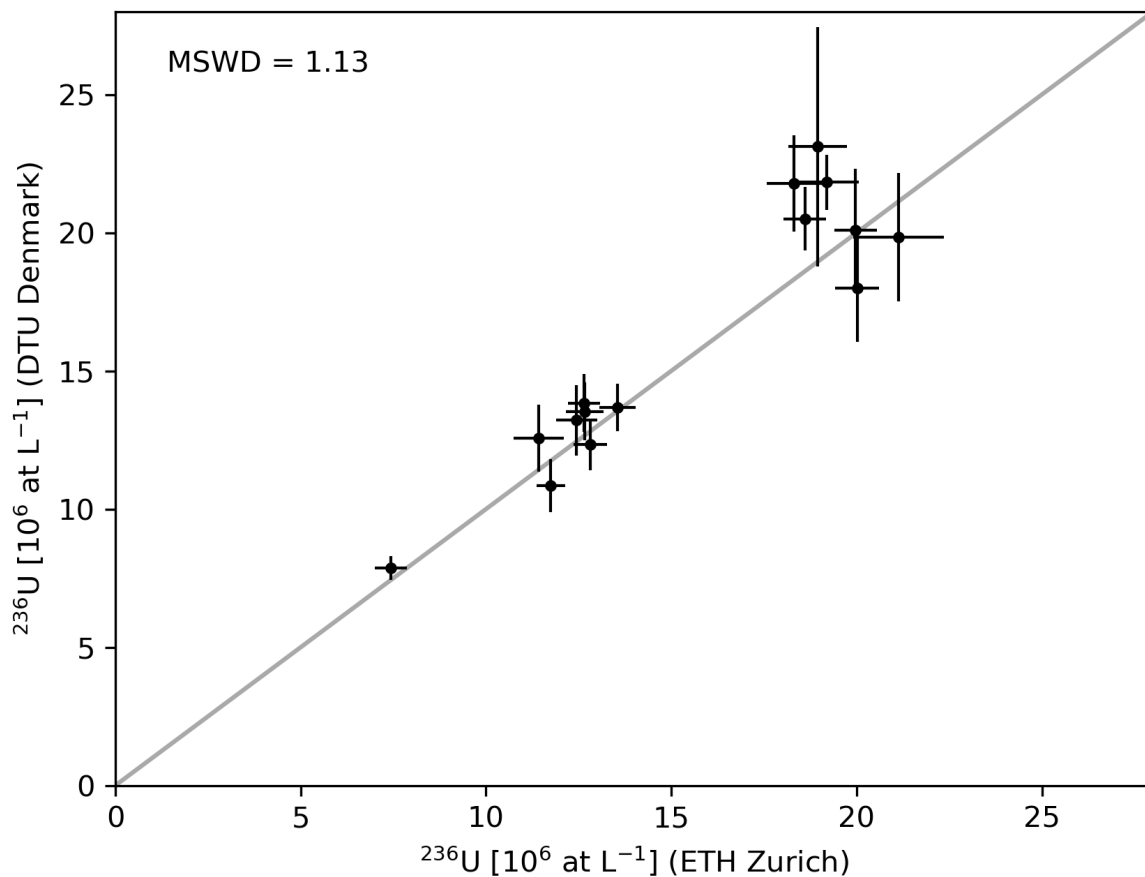


Figure A1. Comparison of ^{236}U concentrations of duplicate samples measured at ETH Zurich and the University of Vienna published by the Technical University of Denmark. The mean square weighted deviation (MSWD) is shown as a measure for agreement within stated uncertainties (1σ).

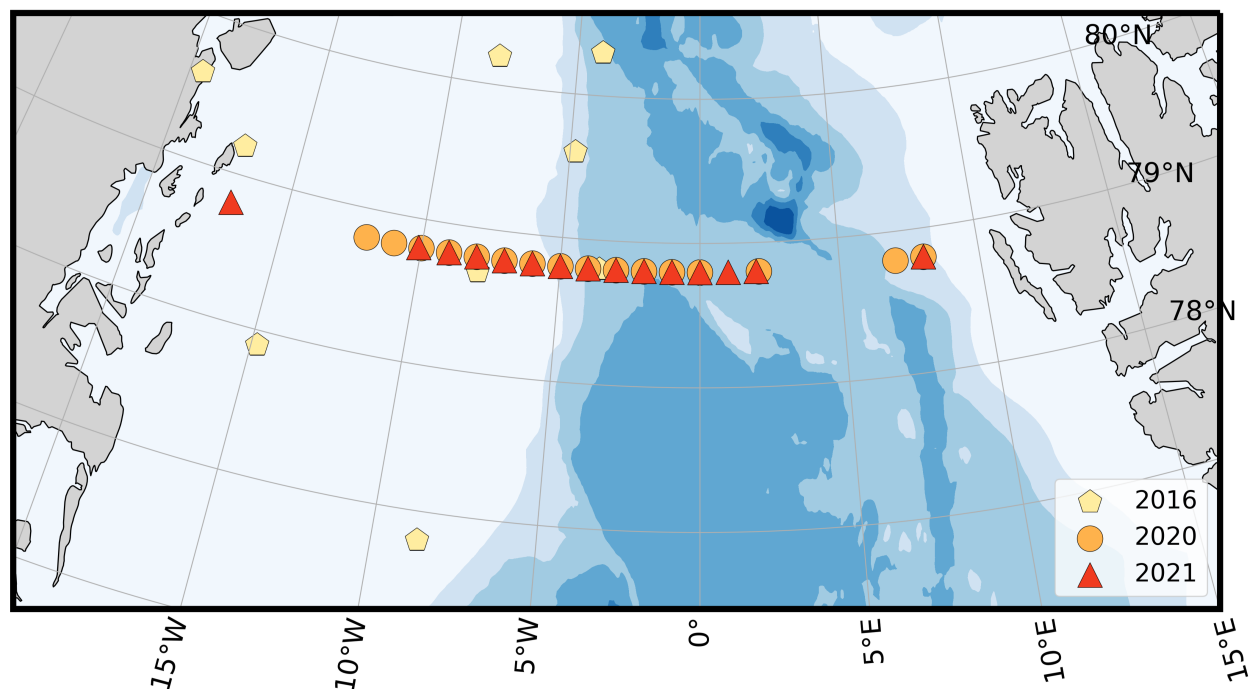


Figure A2. Overview map of all stations in 2016, 2020, and 2021. The data from 2016 is taken from Wefing et al. (2019).

A3 The binary mixing model with data from 2016, 2020, and 2021

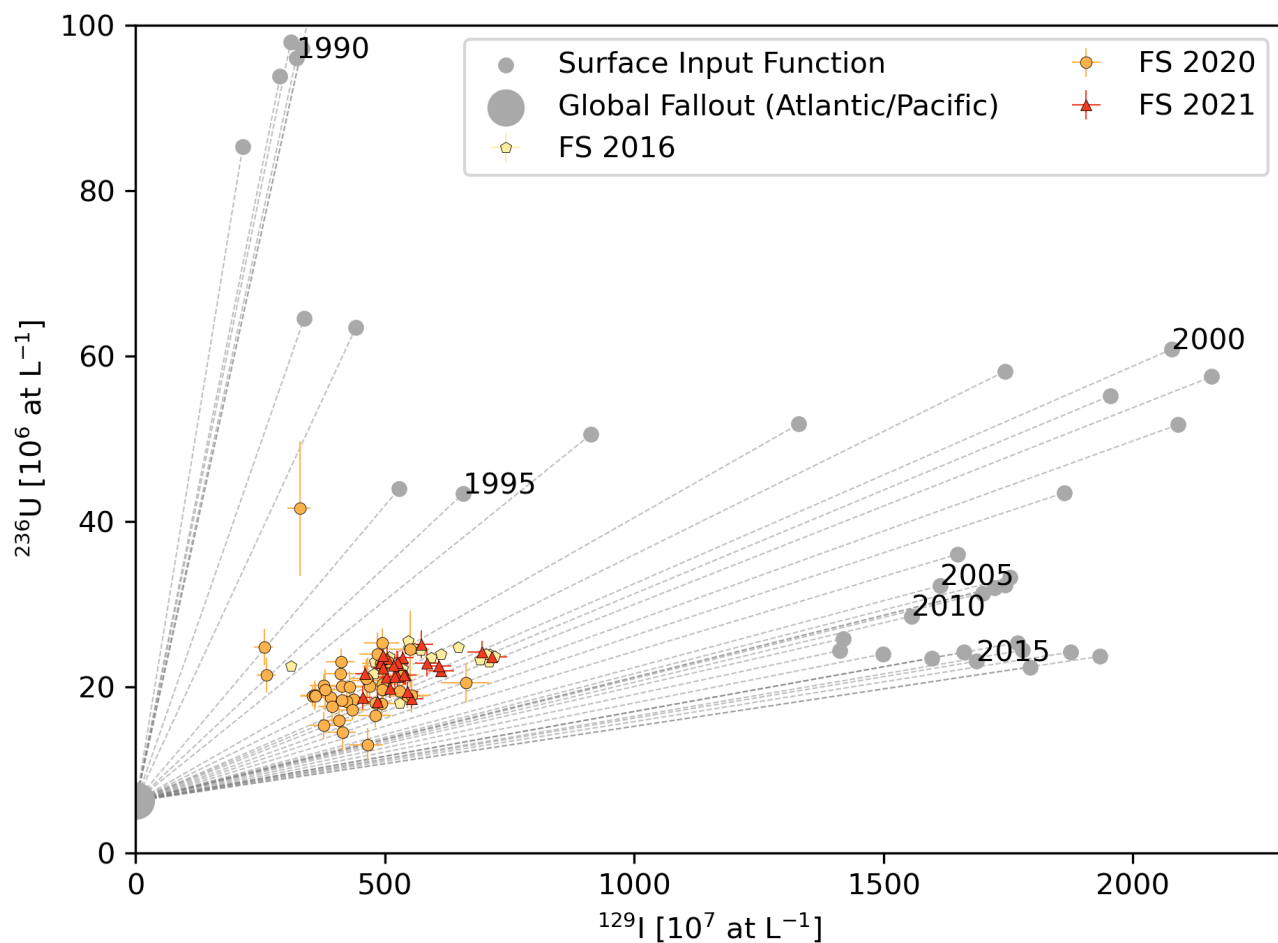


Figure A3. The binary mixing model grid with a constant global fallout background level (large gray point) and a time varying input of ^{129}I and ^{236}U to the surface Arctic Ocean (small gray points). The data for Polar Water I from 2016, 2020, and 2021 is shown on top.

A4 Transit Time Distribution grids with data from 2016, 2020 and 2021

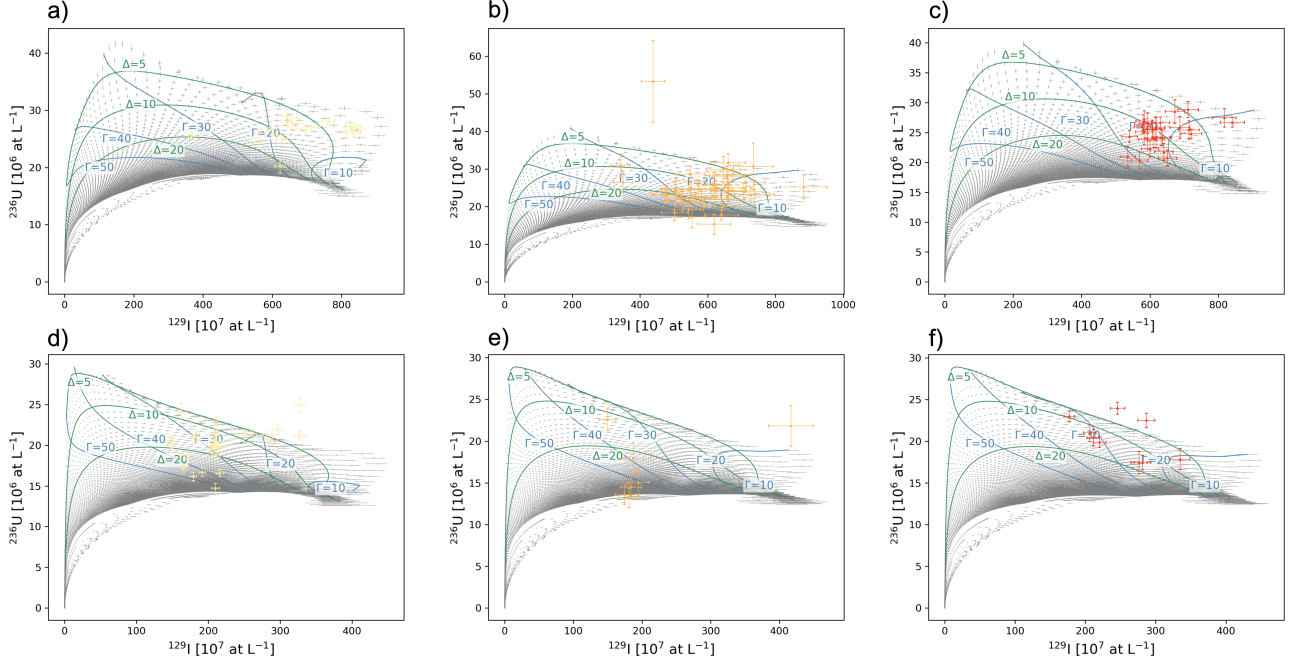


Figure A4. The TTD model grids of ^{129}I and ^{236}U concentrations for 2016 (a,d), 2020 (b,e), and 2021 (c,f) with Polar Water I data shown for the surface (a,b,c) and Arctic Atlantic Water for the mid-depth (d,e,f). Gray dots in all panels represent possible Γ - Δ combinations based on the surface (a,b,c) and mid-depth (d,e,f) input functions with the uncertainties as gray lines. Isolines show the distribution of Γ 's (blue lines) and Δ 's (green lines). The surface data was corrected according to the described method in section 2.3.4 with a 15 %, 25 %, and 15 % global fallout fractions for 2016, 2020, and 2021, respectively.

A5 Transit Time Distributions grids with different Correction Percentages for the Surface

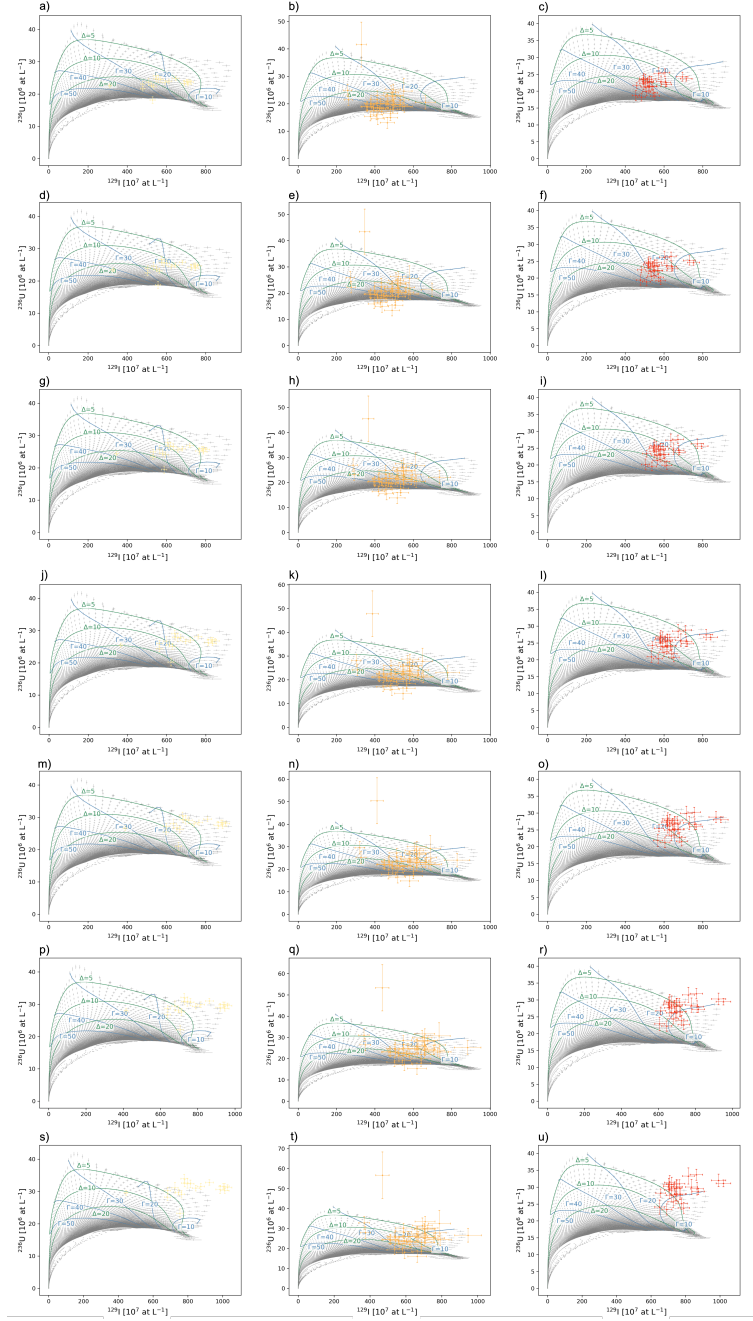


Figure A5. The TTD model grids for 2016 (a,d,g,j,m,p,s), 2020 (b,e,h,k,n,q,t), and 2021 (c,f,i,l,o,r,u) with Polar Water I data shown for the surface. Gray dots in all panels represent possible Γ - Δ combinations based on the surface input function with the uncertainties as gray lines. Isolines show the distribution of Γ 's (blue lines) and Δ 's (green lines). The surface data was corrected according to the described method in section 2.3.4 with a 0 % (a,b,c) to 30 % (s,t,u) (in 5 % increments) global fallout fraction to find the best fit.

A6 Transit Time Distributions grids with Nutrient Correction Percentages for the Surface

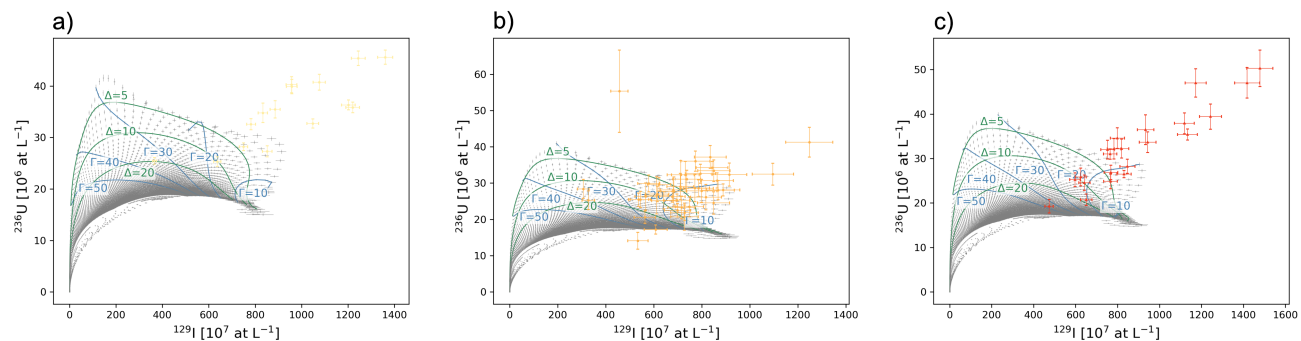


Figure A6. The TTD model grids for 2016 (a), 2020 (b), and 2021 (c) with Polar Water I data shown for the surface. Gray dots in all panels represent possible Γ - Δ combinations based on the surface input function with the uncertainties as gray lines. Isolines show the distribution of Γ 's (blue lines) and Δ 's (green lines). The surface data was corrected according to the described method in section 2.3.4 with nutrient fractions of Pacific Water for each individual sample.

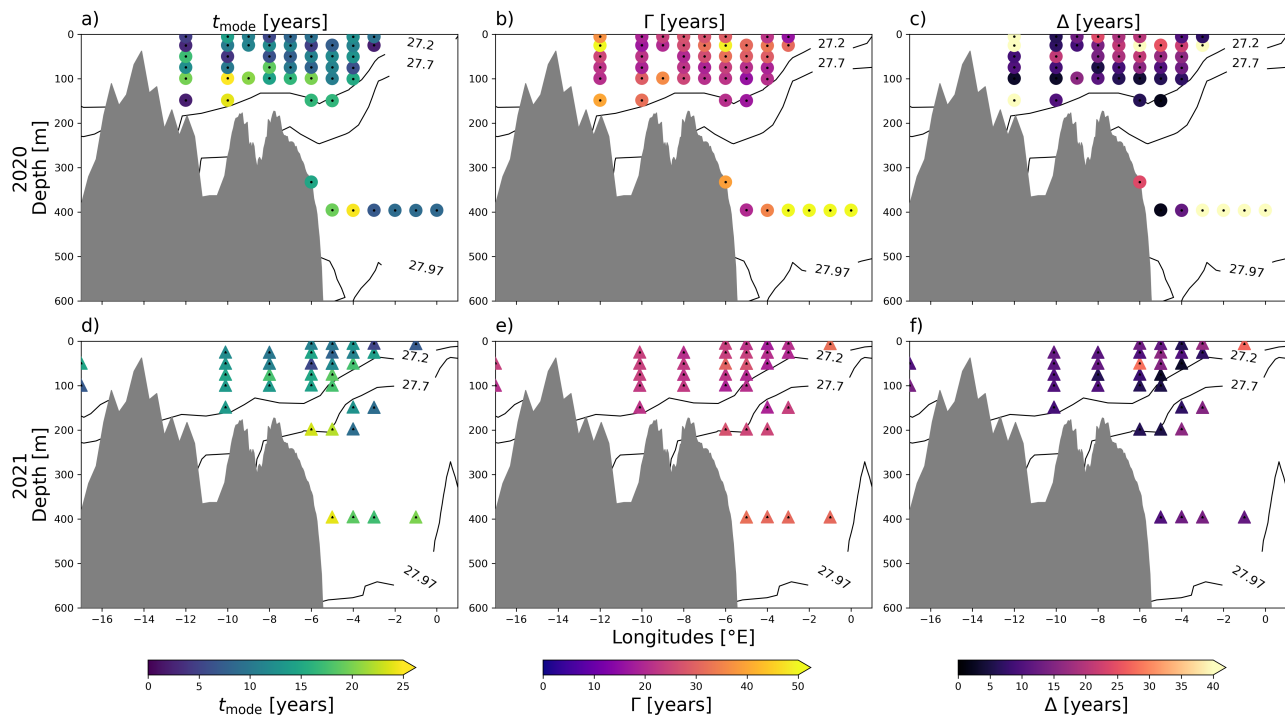


Figure A7. TTD parameters t_{mode} (a,d), Γ (b,e), and Δ (c,f) for 2020 (a,b,c) and 2021 (d,e,f). Results are shown for Polar Water I and Arctic Atlantic Water. Isopycnals (black) for $\sigma_0 = (27.2, 27.7, 27.97)$ are overlaid.

A8 Sensitivity Analysis for surface Transit Time Distribution results

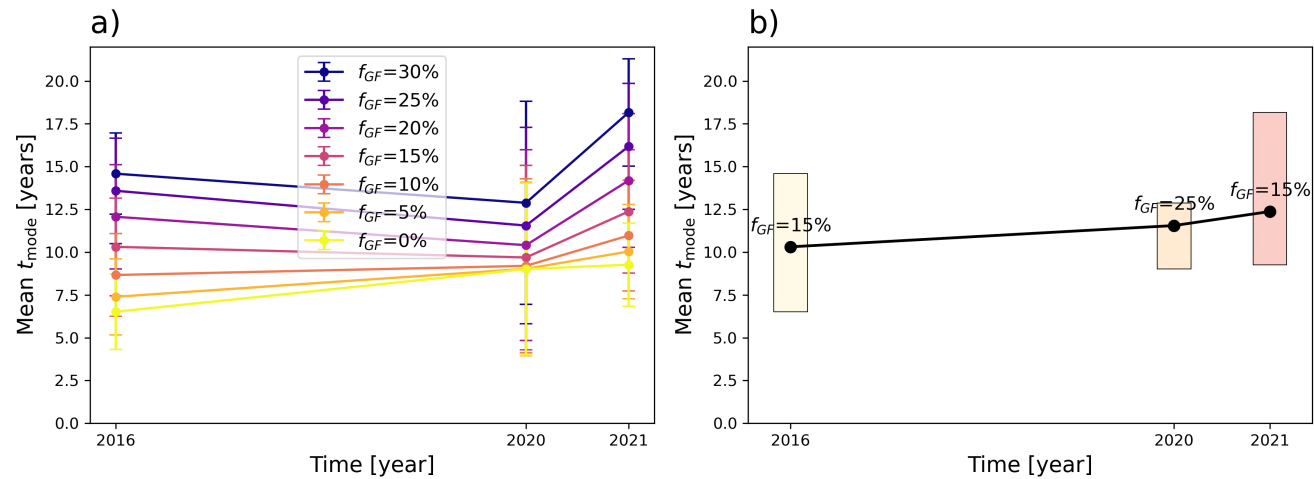


Figure A8. a) Mean t_{mode} (with unc. 1σ) in each year for global fallout fractions (f_{GF}) ranging from 0 % to 30 % in 5 % increments. b) Envelope of minimum and maximum t_{mode} values from panel a), with the black line showing the best-fit fractions used in this study. Results are shown for Polar Water I at the surface.

Author contributions.

MS performed the conceptualization, investigation, data curation, formal analysis, visualization, and wrote the original draft. AMW supported the conceptualization, investigation and data curation and performed writing (review and editing). HPT supported the measurement of data, methodology and performed writing (review and editing). CV supported the measurement of data and methodology. PAD and JPG provided resources and performed writing (review and editing). NC acquired funding, provided resources, supported the conceptualization, investigation, measurements of data, and data curation and performed writing (review and editing).

Competing interests. The authors declare that they have no conflict of interest.

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