A Study of Intermediate Water Circulation in the Strait of Georgia Using Tracer-Based, Eulerian, and Lagrangian Methods

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ABSTRACT: The intermediate circulation of the Strait of Georgia, British Columbia, Canada, plays a key role in dispersing contaminants throughout the Salish Sea, yet little is known about its dynamics. Here, we use hydrographic observations and hindcast fields from a regional 3D model to approach the intermediate circulation from three perspectives. First, we derive and model a “seasonality” tracer from temperature observations to age the water, estimate mixing, and infer circulation. Second, we analyze modeled velocity fields to create mean current maps and examine the advective and diffusive components of the mean flow field. Last, we calculate Lagrangian trajectories to derive transit time distributions and Lagrangian statistics. In combination, these analyses provide an overview of the mean intermediate circulation that can be summarized as follows: subducting water in Haro Strait ventilates the intermediate water primarily via an up-strait boundary current that flows along the eastern shores of the southernmost basin in 1–2 months. This inflowing water is either incorporated into the interior of the basin, recirculated southward, or transported into the northernmost basin, mixing steadily with adjacent water masses during its transit. A second, shallower ventilating jet emanates southward from Discovery Passage, locally modifying the Haro Strait inflow signal. Outside of these well-defined advective features, diffusive transport dominates in the majority of the region. The intermediate renewal signal fully ventilates the region in 100–140 days, which serves as a benchmark for contaminant dispersal time scale estimates.

SIGNIFICANCE STATEMENT: To better understand the movement of contaminants through the Strait of Georgia, this study aims to identify and diagnose the mean circulation pathways of the regional intermediate layer. We develop existing concepts of the local estuarine circulation, showing that newly formed intermediate water fills the layer in a matter of months. Additionally, we identify various along-strait currents in the lateral flow field. The resulting circulation scheme is a complicated mixture of organized currents and eddy-diffusive transport, which can be used to predict the transport and accumulation of contaminants. Future work will aim to link this circulation with regional contaminant distributions and to assess the time variability of the flow features identified here.

KEYWORDS: Seas/gulfs/bays; Ocean circulation; Intermediate waters; In situ oceanic observations; Ocean models; Pollution

1. Introduction

The Strait of Georgia, British Columbia, Canada, an important part of the Salish Sea, is a system of deep basins and waterways with an estuarine circulation (Fig. 1), the inflowing (lower) component of which occurs in the intermediate depths of the Strait. The Strait is deep enough that this Intermediate Water (IW) layer is interposed between a dynamic surface layer (Pawlowicz et al. 2019) and a (mostly) stagnant deeper layer (Masson 2006; Pawlowicz et al. 2007), and is thus shielded from direct wind, solar, and bottom forcing, and in addition is wide enough for Coriolis effects to be important (Stacey et al. 1991). While this is a fjord-type estuary for which an overall

estuarine exchange scheme is well known (Pawlowicz et al. 2007, 2019), details of transport pathways within the intermediate water layer are still obscure. Understanding the circulation of intermediate water masses is critical to understanding the regional oceanography, not only for its intrinsic interest but also to be able to predict where objects or observed parameters of concern will travel over time. One practical motivation for understanding these pathways in the Strait of Georgia system is that a major effluent outfall introduces contaminants into this layer and various regulatory agencies need to know more about the ultimate fate of these substances.

A number of approaches are possible for determining the intermediate circulation. One technique is the so-called “core method” (Wüst 1935), an approach in which maps are made of particular tracers, and arrows are drawn along tongues (i.e., crossing isopleths of these tracer fields). Tracers may be conservative (e.g., temperature or salinity), or may depict age (e.g., dissolved oxygen, or any number of biogeochemical parameters). However, these views of the circulation are often qualitative since speeds are not well known, and in certain cases a geostrophically controlled circulation could be directed along isopleths rather than across them. Quantitative tracer-based approaches include the “box-model”-based Knudsen relations

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FIG. 1. A bathymetric map and circulation schematic of the Salish Sea. Dashed lines and bold font approximately separate and identify the following Strait of Georgia regions: Haro Strait (H.S., also shaded red to identify the IW formation region), the southern basin (S.B.), and the northern basin (N.B.). Johnstone Strait (J.S.), seen in the inset map, connects Discovery Passage to the Pacific in the northwest of this region. The island marked L.I. is Lasqueti Island. The black box in the inset map depicts the spatial extent of SalishSeaCast numerical modeling, the red box depicts the limits of the main map. Schematic shows a simplified side-on view of regional circulation. Red region responds to the Haro Strait region of the main map, bold black arrows represent IW circulation, and gray arrows represent other aspects of IW circulation.

for estuarine flow (Hansen and Rattray 1965; MacCready et al. 2018) or more complex variations thereof, applied locally by Pawlowicz (2001), Riche and Pawlowicz (2014), and MacCready et al. (2021); or optimal water mass analyses (Tomczak 1981)—this latter approach was previously used along the thalweg of the Salish Sea as a whole by Masson (2006).

A second approach is to consider the Eulerian mean velocity field, obtained by taking a long-term average of direct current

measurements at fixed locations. This method could include means derived from mooring programs, but also those inferred from an observed (mean) three-dimensional density field using mathematical theories (e.g., geostrophic/ β spiral and inverse methods; Wunsch 1996) or the mean velocity fields in numerical models. Eulerian mean velocity fields can be problematic for predicting tracer transport, since they ignore the turbulent nature of oceanic flow. The turbulent flux of particular

properties—which can be separated from the mean flux via a Reynolds decomposition and termed “eddy flux” (Davis 1991)—depends on details of the time-varying current field, and is important to the transport of tracers in energetic regions of the ocean (Wunsch 1999). Numerical investigations have been attempted for the Salish Sea circulation on a number of occasions, but have generally concentrated on circulation aspects closer to the Pacific entrance (Marinone and Pond 1996; Masson and Cummins 2004; MacCready et al. 2021).

A third approach is Lagrangian, in which water parcels are tracked either directly by instruments (floats or drifters; Swallow and Worthington 1961; Rossby et al. 1986) or indirectly by numerical integration through time-varying velocity fields of numerical models (Daher et al. 2020; Onink et al. 2019). These approaches are perhaps the most realistic in terms of predicting where objects within the ocean will go, but are for various reasons the most difficult to analyze, involving orders of magnitude more information (Davis 1994). While producing results from the first two methods generally requires few analytical steps, the analyses of Lagrangian summaries are often more complicated (LaCasce 2008; van Sebille et al. 2018). These approaches have not yet been applied to the Strait of Georgia’s IW layer, although they have been used to study regional surface circulation (Pawlowicz et al. 2019; Pawlowicz 2021) and Boundary Pass/Haro Strait circulation (Soontiens and Allen 2017).

Usually a particular investigation relies on only one of these approaches. Here, we take advantage of recently available resources to implement all three approaches to better understand the spatially varying circulation within the intermediate layer of the Strait of Georgia. The newly available resources that allow for this intercomparison include observations from a citizen science program that has provided many thousands of hydrographic profiles between the years of 2015–18, relatively well spread out in both time and space (Pawlowicz et al. 2020). These measurements are used to derive, model, and track a temperature-based “seasonality” tracer, providing direct estimates of water mass age within the intermediate layer, and to evaluate the accuracy of a newly available high-resolution regional model, which provides time-varying hindcast estimates of hydrographic and velocity fields (Soontiens et al. 2016; Soontiens and Allen 2017; Moore-Maley et al. 2016; Olson et al. 2020). Interpretation of the seasonality tracer is assisted by comparison with a 1D advective/diffusive “pipe” model, which provides a simplified but intuitive representation of the system. We then analyze the model velocity fields to estimate Eulerian currents and to seed numerical Lagrangian pseudo-drifters to directly predict the transport of IW water parcels. Using information from these analyses, we identify gyres and regions of strong flow to create a comprehensive overview of the Strait of Georgia intermediate circulation.

2. Materials and methods

a. Regional overview

The Strait of Georgia (SoG), Canada, is part of the Salish Sea, a region within both the United States and Canada lying on the inshore coast of Vancouver Island (Fig. 1). The SoG is

about 30 km wide and 200 km long, with a maximum depth of about 430 m and a mean depth of 150 m. Its water column can be divided into three layers: a seaward flowing surface layer (0–50 m; Pawlowicz et al. 2019), an inflowing (on average) Intermediate Water (50–200 m; LeBlond et al. 1991; Pawlowicz et al. 2007), and a deep water layer (200 m–bottom), which is stagnant most of the year, but is ventilated by (approximately) monthly pulses of oceanic water during the summer (LeBlond et al. 1991; Masson 2002). These water masses represent 27%, 53%, and 20% of the overall volume of the SoG, respectively (Pawlowicz et al. 2007). By volume alone, the IW may therefore be the most important water mass in the system—especially considering it is the layer into which most contaminants enter via effluent outfalls (Sun et al. 2018)—but it is least well understood. Various other wide, layered coastal seas with subsurface intrusions to which the results presented here may broadly apply include the Baltic Sea (Lehmann et al. 2002); Prince William Sound, Alaska (Niebauer et al. 1994); and the Gulf of St. Lawrence (Saucier et al. 2003).

The Salish Sea estuarine circulation is driven by significant freshwater input, dominated by that of the Fraser River (Griffin and LeBlond 1990; Li et al. 1999; Riche and Pawlowicz 2014). The IW is formed in the Haro Strait area by tidal mixing between outflowing surface water, with a large seasonal temperature cycle, and inflowing deeper Pacific water. Thus, the IW source water has seasonally varying characteristics. Previously, the phase lag in seasonal temperature cycles from a handful of stations in the southern basin has been used to drive a simple box model to estimate an IW residence time of 160 days (with an uncertainty range of 100–330 days) before being entrained into the surface layer (Pawlowicz et al. 2007).

Historically, scientific efforts have targeted the more dynamic southern basin and thus a comprehensive understanding of the lateral IW circulation and transport pathways remains obscure. However, a relatively rapid inflowing eastern boundary current in the southern Strait was deduced from hydrographic observations (Pawlowicz et al. 2007), and a variety of numerical models and observational programs have (inconsistently) depicted gyre-like horizontal circulations within the deeper parts of the Strait (Waldichuk 1957; Stacey et al. 1987; LeBlond et al. 1991; Marinone and Pond 1996; Masson and Cummins 2004). An internal deformation radius of ~8 km suggests that the IW layer is strongly rotational (Stacey et al. 1991).

b. Data sources

Hydrographic observations in the SoG are available from multiple sources, consisting primarily of conductivity–temperature–depth (CTD) profiles and bottle measurements. Several regional monitoring programs include quarterly or triannual shipboard surveys of the area carried out by government scientists, and a long-term (from 1968 to date) set of weekly CTD profiles at a central location in the Strait obtained as a by-product of sound-speed measurements required by the Canadian Forces Maritime Experimental and Test Ranges (Masson and Cummins 2007). In the southern Strait, a set of 48 hydrographic surveys were carried out over three years in 2003–05 (Pawlowicz et al. 2007), and in the northern Strait a regular set of observations at one location has been carried out

weekly since 2015 by the Hakai Foundation. However, the most comprehensive dataset for this paper is obtained from the Pacific Salmon Foundation's Citizen Science Oceanography Program (Pawlowicz et al. 2020). Some 80 stations over the entire Strait are occupied at nearly the same day by 7–10 different patrols, about 20 times a year from 2015 to present (more than 1000 CTD profiles a year). These profiles extend only to 150 m, so that the deep waters of the Strait are not measured, but this depth range suffices for our investigation of the IW.

We can derive Eulerian current velocity estimates at locations in the southern basin where long-term observations are available. As part of a cabled observatory maintained by Ocean Networks Canada, three upward-facing acoustic Doppler current profilers (ADCPs) measure currents every 2–5 m throughout the water column. The ADCP records span a 7–12-yr period between September 2008 to May 2020, albeit with some data gaps (Ayranci and Dashtgard 2020). Additionally, cyclesonde and current meter measurements from Stacey et al. (1987) between June 1984 and January 1985 at 100 m are available from nine stations in the central Strait.

Estimates of tracer fields and velocities in the Strait are also available from SalishSeaCast, a Nucleus for European Modeling of the Ocean (NEMO) 3.6 model configured for the Salish Sea (Olson et al. 2020; Soontiens and Allen 2017). SalishSeaCast is a baroclinic, three-dimensional, primitive equation model with 40 vertical depth levels (1 m spacing near the surface, 27 m at the bottom) and an approximately 440 m $\times$ 500 m horizontal grid. The model domain encompasses the entire Salish Sea (black box on Fig. 1 inset map), with open boundaries at the mouth of Juan de Fuca Strait at the west and Johnstone Strait at the north. It is forced by Environment and Climate Change Canada's High Resolution Deterministic Prediction System (HRDPS) atmospheric fields (Milbrandt et al. 2016), a mix of measured and climatological rivers, modeled tracer concentrations at the western boundary (Live Ocean Model; Davis et al. 2014; Siedlecki et al. 2015), observed climatologies at the northern boundary (provided by the Hakai Institute), and modeled tides (Webtide; Foreman et al. 2000) tuned with tidal measurements and sea surface height observations. Vertical turbulence is handled using turbulent kinetic energy closure and generalized length scale schemes.

From model hindcasts (version v201905), we extract daily velocity and temperature fields spanning November 2014 to January 2020 (inclusive). Pseudodrifter trajectories within SalishSeaCast hourly velocity fields are calculated over a 2-yr period (January 2016–December 2017, inclusive) using the community-based Lagrangian numerical analysis tool, Ariane (Blanke and Raynaud 1997; van Sebille et al. 2018).

3. Results

a. Seasonality as an age tracer

Our first circulation estimate is based on the assumption that water is generally transported in the direction of greater age, where age represents the time since the water was last in contact with the atmosphere. A typical age tracer is based on dissolved oxygen concentration, but this technique is not

directly useful here because the temperature of the source water changes seasonally, so that the oxygen content also changes seasonally. Instead, we derive a new tracer based on seasonality using the conservative property of temperature. The seasonality tracer is analogous to anomaly tracers in open-ocean studies (e.g., Sasaki et al. 2010), although the regularity of the seasonal cycle allows us to more quantitatively and accurately estimate age. In addition, as we shall explain below, the seasonality tracer can also be used to model the relative importance of both advective and (eddy) diffusive processes in transport.

Previous studies have noted that the IW temperature varies seasonally in an approximate sinusoidal pattern (LeBlond et al. 1991; Pawlowicz et al. 2007). In the surface layer, the seasonal temperature cycle is in phase with the atmospheric seasonal cycle—coolest in early February and warmest in late August. However, the coldest and warmest times in the IW seasonal temperature cycle occur later in the year. This delay increases with northward distance from Haro Strait, where these waters were last in contact with the atmosphere. Thus, water with a specific temperature, set by air/sea heat processes in the mixing regime of Haro Strait, subducts upon entering the SoG and then tends to keep a conserved temperature “identity” as water parcels move northward in the intermediate layer.

Rather than trace any particular temperature, we then consider the phase of the seasonal cycle as a measure of water age. For example, if the coldest point of the seasonal cycle at a given station occurs in April, then we infer that it has taken about 2 months for the water to get there from the source region. In addition to a delay in the timing of the temperature seasonal cycle, we will also find that the seasonal cycle amplitude will also decrease with age due to mixing with surrounding waters on the northward path; the degree of decrease will allow us to estimate the importance of mixing.

The basic concept of the seasonality tracer can be demonstrated with a simple toy model. Consider the IW layer as a one-dimensional pipe, in which temperature $T(x, t)$ for $x, t \geq 0$ is governed by both (constant) advection U and diffusion with a diffusivity K :

$$T_t + UT_x = KT_{xx}, \quad (1)$$

where subscripts indicate partial derivatives. If we set a seasonally varying boundary condition at $x = 0$, $T(0, t) = \Delta T_0 \cos(\omega t)$, then we have

$$T(x, t) = \Delta T_0 \exp(-k_r x) \cos(k_i x - \omega t), \quad (2)$$

with $k_r = \text{Re}(k)$ and $k_i = \text{Im}(k)$, and

$$k = \left(\frac{1}{\gamma}\right) \left\{ \sqrt{1 + 2i \frac{\omega}{U} \gamma} - 1 \right\}, \quad (3)$$

with $\gamma = 2K/U$ an advection/diffusion balance length scale. It is important to note that k_r sets the decay length scale with x of the input oscillation amplitude; ω/k_i determines the propagation speed of oscillation phase; U represents the mean propagation of our tracer through the pipe; and K represents all higher-order processes that are not encompassed in this mean

FIG. 2. Solutions of the 1D advection–diffusion pipe model with a seasonal oscillation at $x = 0$ for varying γ . The form of the solution is set by $\gamma = 2K/U$ with (a),(d) weak; (b),(e) intermediate; and (c),(f) strong diffusion cases shown. Shown in (a)–(c) are time/range contour plots, where the solid white line shows the speed of advection U (1.5 cm s^{-1}) and the dashed white line shows apparent propagation speed. In (d)–(f), the spatial solution at $t = 0$ is shown in black lines and the envelopes of the oscillation in red. The colored contours represent the modeled oscillation of temperature. (g) The real and imaginary components of k as a function of γ [Eq. (3)]. The black vertical lines represent $\gamma = 20, 100$, and $10\,000$ km for the three above cases. The blue shaded region represents the 56 – 75 -km γ range calculated from the tracer advection speed and the Lagrangian diffusivity in section 4.

advection term, which include spatially varying mean flow (e.g., lateral inhomogeneities) as well as time-varying influences associated with mixing and stirring.

For a realistic simulation of the SoG, we set the length of the pipe $X = 200$ km, the frequency of the oscillating input as $\omega = 2\pi/365$ days (representing an annual cycle), and an advection speed $U = 1.5 \text{ cm s}^{-1}$ based on an approximate through-Strait transit time of 160 days. As γ varies, different behaviors are possible: in the case that γ is small, advection dominates and the oscillating signal propagates down-pipe at the advection speed U with a constant amplitude (Figs. 2a,d). When γ is large and diffusion dominates, the oscillating input at $x = 0$ is mixed almost instantly throughout the pipe and the apparent advection speed is greater than U with very little change in amplitude (Figs. 2c,f). For an intermediate γ where diffusion and advection are balanced, the signal also propagates with an apparent advection speed that is greater than U ,

but in addition the amplitude of the oscillation decreases with x (Figs. 2b,e). In this case, the speed of phase propagation is greater than the advective speed and, at the furthest extent of the pipe, the amplitude of the inflow signal has decreased by $\sim 55\%$ (Fig. 2b).

Thus, the seasonality tracer has an advantage over simpler tracers since it can also indicate the relative strengths of mixing and advection. The amplitude decay scale is set by k_r , which peaks between $\gamma = 50$ and 500 km (Fig. 2g). This result suggests that, for amplitude damping to occur within this system, the measured advection/diffusion length scale γ should fall within this range. This behavior is relatively robust to some realistic variations. For example, Pawlowicz et al. (2007) and Riche and Pawlowicz (2014) both imply that U might vary seasonally by up to a factor of two. Some numerical experimentation with Eq. (1) in which U is varied seasonally above and below a mean value shows that the sinusoidal shape of seasonal cycles

FIG. 3. Seasonal cycle characteristics. (a) Representative temperature climatologies and fitted seasonal sinusoids from two stations in the southern and northern basin. Scatter points represent the measurements from the two stations, solid lines represent the sinusoidal cycles fitted to these climatologies, and dotted lines indicate the coldest Julian day of the year (ϕ_{CD}) derived from these seasonal cycles. The colors of the scatter points and lines correspond to the colored stations in inset map. Observations from the southern station span March 2016–October 2018 and from the northern station span February 2015–January 2019. (b),(c) The variation of seasonal cycle characteristics as a function of depth. The sinusoid fitting algorithm (section 3b) was performed on temperature climatologies from each depth level at all stations, so that each individual station has a ϕ_{CD} depth profile and an amplitude (A_t) depth profile. Each depth level of these profiles was normalized by subtracting the overall mean (for that depth level) and dividing by the standard deviation (for that depth level) to produce normalized property profiles (plotted in gray). The black line represents the mean of these normalized profiles, intersected with colored lines to denote upper (blue; 60–80 m), core (red; 90–110 m), and lower (yellow; 120–140 m) IW.

within the Strait becomes distorted by the presence of higher-frequency harmonics. However, the phase of an annual sinusoid fitted to these distorted shapes remains almost unchanged from that predicted using the time-invariant U , while the amplitude decay of the fitted sinusoids increases only slightly by an additional 10%.

b. Seasonality tracer: Renewal signal

Now, we apply these ideas to our observations by fitting a sinusoidal cycle to a yearly temperature climatology at each location and depth in the Strait for which a long (>11 months) time series of hydrographic data is available. A least squares method is used to find coefficients of the function:

$$T_{\text{obs}}(t) = A_t \cos(2\pi t/L + \phi) + b + \varepsilon, \quad (4)$$

where T_{obs} are the observed temperatures, t represents the Julian yearday upon which the data were measured, L represents the length of the annual cycle (i.e., 365.25 days), A_t the amplitude and ϕ the phase of the seasonal cycle, b represents the climatological mean, and ε is model misfit. It is more intuitive to represent ϕ by an equivalent variable, the yearday of coldest water (ϕ_{CD}): typically, $\phi_{CD} \sim 40$ (i.e., in early February) for surface water at all locations. A_t is $O(1^\circ\text{C})$.

Since the dataset contains both seasonal and interannual variability, there is some deviation around the sinusoidal fits

(Fig. 3a). An error analysis for the sinusoid fitting is carried out using a balanced bootstrap resampling technique (Efron and Tibshirani 1994). The temperature climatologies for each station are randomly resampled 500 times while maintaining the same overall occurrence of each sample, and the fitting process carried out for each permutation. Uncertainties are then taken as the standard error of the bootstrap replicates. This uncertainty, plotted as error bars in Fig. 4, is typically about 13 days in ϕ_{CD} and 0.1°C in A_t .

To assess whether the IW behaves uniformly at all depths, we can examine how the seasonality tracer varies as a function of depth (Fig. 3). To focus on the vertical structure of the water mass, we normalize each profile by subtracting its mean and dividing by its standard deviation, centering the profiles around a zero value. Examining these normalized A_t profiles from all of the stations (Fig. 3c), we find large amplitudes at the surface where the water column is influenced by the atmospheric seasonal cycle of warming and cooling. This influence decreases with increasing depth until ~ 60 m, and then increases to a weak maximum at ~ 100 -m depth. This broad maxima in A_t , coinciding with a depth range over which the phase is roughly constant (Fig. 3b), represents the advected signal of IW present at this depth over a large portion of the hydrographic stations. Later we find there is some spatial variability in the depth of the maximum, thus for a thorough representation of the IW layer, we divide the water mass into

FIG. 4. Seasonal cycle spatial variability for the Strait of Georgia. Hydrographic measurements span 1969–2019, though are primarily from the 2000–19 time period. (a) A map detailing station locations denoted by the colored markers. Marker colors in (a) correspond with the colored y axis in (b), (e), and (h) to denote seasonal cycle properties as a function of along-strait distance from Haro Strait. Shown are data from (b)–(d) the upper IW (60–80 m), (e)–(g) the core IW (90–110 m), and (h)–(j) the deep IW (120–140 m). In (c), (d), (f), (g), (i), and (j) gray scatter points show the variability of the seasonal cycle phase shifts ϕ_{CD} and amplitudes A_t as a function of along-strait distance from Haro Strait; the error bars show the mean values $\pm$ the standard deviation, representing the uncertainty in the sinusoidal fitting algorithm [Eq. (4)] calculated using a balanced bootstrap resampling technique. For reference, the quantity age_t is calculated $age_t = \phi_{CD} - \phi_{HS}$, where ϕ_{HS} is the earliest ϕ_{CD} measurement from Haro Strait. The solid red lines detail the linear trends in these properties. The yellow and blue lines in (f) detail the same trends for the Lagrangian transit time and SalishSeaCast model seasonality analyses, respectively. Trends are derived using a Theil–Sen method.

upper IW (60–80 m), core IW (90–110 m), and lower IW (120–140 m) and analyze each division separately.

On a regional scale (Fig. 4) the seasonal cycles show a consistent spatial variation, with a general delay in ϕ_{CD} with along-strait (i.e., northward) distance from the formation zone. To calculate the age of the water using the seasonality tracer (age_t , hereafter), the earliest ϕ_{CD} value from stations in Haro Strait (between 40 and 65 days, depending on the depth level) was subtracted from ϕ_{CD} at each station. The relationship between along-strait distance and age_t is strongest in the deep IW (correlation coefficient $R = 0.87$) and weakest in the surface

IW ($R = 0.37$). The gradient of this relationship [a linear fit, calculated using a Theil–Sen approach (Sen 1968); error values denote 95% confidence interval], yields an approximate northward transport velocity of $1.5 \pm 0.07 \text{ cm s}^{-1}$ in the upper IW, $2.0 \pm 0.10 \text{ cm s}^{-1}$ in the core IW, and $1.9 \pm 0.10 \text{ cm s}^{-1}$ in the lower IW. When considering the constant advection term U in the 1D model [Eq. (1)], the 25% difference in transport velocity between sublayers equates to small changes in the modeled propagation speed and amplitude damping signal (on the order of $\sim 10\%$), but does not significantly change the position of the k_r maxima (Fig. 2g), thus it does not impact the

advection/diffusion properties of 1D system. An accompanying reduction in seasonal cycle A_t also occurs at all depths (Fig. 4): a reduction of $0.45^\circ \pm 0.06^\circ\text{C} (100\text{ km})^{-1}$ ($R = 0.79$), $0.44^\circ \pm 0.05^\circ\text{C} (100\text{ km})^{-1}$ ($R = 0.82$), and $0.37^\circ \pm 0.05^\circ\text{C} (100\text{ km})^{-1}$ ($R = 0.83$) for the upper, core, and lower IW, respectively, representing a 78%–85% attenuation of A_t over the length of the strait. This measured attenuation is larger than that derived from the intermediate γ case of the 1D model ($\sim 55\%$; Fig. 2e), indicating that, as might be expected, the 1D toy model, although showing the essential behavior, is not completely satisfactory as a quantitative model of the Strait. Other issues that may be important include the effective length of the Strait used—the curved cross-isobath paths (coral arrows) indicated in Fig. 5a may be longer than the straight-line 200-km distance we have estimated. More importantly, the “leakage” of water out of the IW layer along the length of the Strait (which almost by definition must approach 100% in an estuarine system) suggests that in a better approximation the advection speed U should decrease northward (to zero at the northern end of the Strait). Numerical experimentation shows that such a decrease does significantly reduce the cycle amplitude in the northern Strait relative to that calculated in our analytically convenient constant U approach; as long as diffusion is also important the phase delays are not greatly affected.

We can also employ this technique using other seasonally varying properties; an identical analysis using dissolved oxygen produces similar results to our temperature-derived analysis (Figs. S1 and S2 in the online supplemental material). However, the mean temperature b from the sinusoidal fitting is almost constant throughout the Strait (as one might expect for a conservative property), whereas the mean dissolved oxygen level decreases with age. We then calculate a decrease of $0.69 \pm 0.18 \mu\text{mol kg}^{-1} \text{ day}^{-1}$ in the IW, representing oxygen utilization by remineralization processes (see supplemental material for further details on dissolved oxygen analysis).

c. Seasonality tracer: Lateral circulation

The strong relationship between along-strait distance and the two seasonality tracers— ϕ_{CD} and A_t (Fig. 4)—is the first-order description of IW renewal that is directly related to the features of our 1D pipe model (Fig. 2). However, the scatter in ϕ_{CD} and A_t around the mean northward trend suggests that unresolved lateral variability may be present in the mean fields. Although the conceptual link to the 1D pipe model is now less direct, we will proceed by mapping these mean fields, and then assume that, as in the “core” method, a spatially variable mean advection occurs mostly across isopleths.

To analyze data from the irregularly spaced hydrographic stations, we used a kriging interpolation method (Davis 2002) to estimate the age, and A_t fields over a uniform horizontal grid of data points with a spacing of $0.1\text{ km} \times 0.2\text{ km}$ over the deepwater areas of the Strait. Ordinary kriging is used with a spherical model for the semivariogram of the derived dataset and a best-fit algorithm was used to derive the coefficients of the covariance function. The range coefficient (i.e., the decorrelation length scale) for all interpolations was $\sim 35\text{ km}$; no nugget effect was applied. We then map the interpolated tracer

fields to evaluate the 2D spatial expression of the circulation (Figs. 5 and 6).

Just north of Haro Strait, the younger water of the southern inflow is transported primarily along the eastern boundary of the southern basin; IW ages in this boundary flow range from 10 to 40 days (Figs. 5a–c). This tongue appears at all three depth levels, extending as far north as Texada Island. The water west of this inflow on the shores of Vancouver Island is 20–40 days older.

At the northernmost extent of the southern basin, just south of Texada Island, an age, of ~ 70 days is fairly uniform across the Strait, identifying the northern limit of the boundary current. Note that a gyre-like pattern appears in the southern basin, in which regions of older water are surrounded (or nearly surrounded) by younger water, depicting a system with two interior cores (coral-colored stars in Figs. 5a,d). The oldest waters are generally found at the northeast of the region (Fig. 5), and are 10–20 days older in the deep sublayer compared to the surface sublayer.

The spatial maps also reveal a new intermediate feature. Originating in Discovery Pass, a second, weaker tongue of younger water extends southward along the coast of Vancouver Island in the shallowest intermediate layer (Figs. 5a,d). Similar to Haro Strait, Discovery Pass is a region where strong tidal mixing homogenizes the water column (Thomson 1976). This feature suggests that these Discovery Passage transformation processes form intermediate waters that then enter the northern basin as an inflowing tongue of slightly shallower IW. The presence of this southward tongue is responsible for the $\sim 25\%$ decrease in the mean northward velocity of upper IW relative to the values for core and deep IW (Fig. 4).

Comparing spatial maps of A_t (Figs. 6a–c) with those of age, (Figs. 5a–c), we find that A_t (Fig. 6) decreases with age. At the southern entrance, amplitudes are as large as 1.5°C , but they decrease to $<0.5^\circ\text{C}$ in the northern Strait. This signal is again found entering the southern basin preferentially along the eastern boundary, which has A_t of 0.75°C – 1.25°C .

The 1D pipe model does not incorporate the influence of the Discovery Pass inflow or any other signal. We can, however, apply the advection/diffusion characteristics of the 1D model to help understand the dynamics of the northern inflow. Based on our toy model (Fig. 2), we have intuited: 1) the amplitude of the seasonal signal is damped when advection and diffusion are balanced but stays roughly constant if one dominates; 2) if advection dominates, then the age progression (i.e., the aging of water as a function of along-strait distance) provides a good estimate of mean flow; and 3) if diffusion dominates, then the age progression will be far faster than the mean flow. Applying this intuition to the seasonality observations, the overall decrease in A_t as age, progresses suggests that advection and diffusion are both important in understanding the overall northward transport. Furthermore, although both the southern boundary current and the overall northward transport appear similar in both age, and A_t plots, the northern tongue is only apparent in the age, mapping (Figs. 5a,d); there appears to be limited amplitude damping along the coast of Vancouver Island (Figs. 6a,d). This finding suggests that either advective

FIG. 5. Maps of seasonality tracer-based age (age_t) from (a),(d) the upper IW (60–80 m); (b),(e) the core IW (90–110 m); and (c),(f) the deep IW (120–140 m). Shown is age_t derived from (top) observational temperature climatologies and (bottom) SalishSeaCast hindcast temperature climatologies. Coral-colored arrows drawn in in (a) and (d) represent circulation estimated from a core method perspective, coral-colored stars represent the cores of the gyre-like circulation. Hydrographic station locations are plotted as black scatter points in (a)–(c).

FIG. 6. Maps of IW seasonal cycle amplitude A_t from (a),(d) the upper IW (60–80 m); (b),(e) the core IW (90–110 m); and (c),(f) the deep IW (120–140 m). Shown is A_t derived from (top) observational temperature climatologies and (bottom) SalishSeaCast hindcast temperature climatologies. Hydrographic station locations are plotted as black scatter points in (a)–(c).

or diffusive processes dominate southward transport along Vancouver Island, and that this inflow behaves differently from the advection/diffusion regime of the southern inflow.

d. Evaluation of SalishSeaCast model

For the Eulerian and Lagrangian analyses, we now consider the numerical SalishSeaCast model simulations. We first confirm that these simulations accurately portray the Strait by applying the seasonality tracer methodology to calculate age, and A_t from modeled temperature climatologies, which are compiled from daily hindcasts spanning November 2014–January 2020, and comparing these fields to the observations (Figs. 5 and 6). The model grid is highly resolved and nearly uniformly distributed in space; no interpolation was performed on the model fields prior to mapping.

To test the agreement between the observed seasonality fields and the modeled, full-resolution seasonality fields, we first use a nearest neighbor scheme to map the (interpolated) observational data grid to the model data grid. We then calculate skill scores (SS) per Murphy (1988), where we compare the root mean squared error between observed (X) and modeled (Y) values to the variance of the observations:

$$\text{RMSE} = \left[\frac{1}{n} \frac{1}{m} \sum_{i=1}^n \sum_{j=1}^m (Y_{ij} - X_{ij})^2 \right]^{1/2}, \quad (5)$$

$$\text{SS} = 1 - \frac{\text{RMSE}^2}{\sigma_X^2}, \quad (6)$$

where indices i and j represents the grid coordinates and σ represents the standard deviation. An $\text{SS} = 1$ indicates perfect agreement between the X and Y fields, while an $\text{SS} > 0$ is generally considered a minimum requirement for acceptable performance (Murphy and Epstein 1989). For age_t fields, $\text{SS} = 0.65$ (upper), 0.72 (core), 0.48 (lower) and for A_t fields, $\text{SS} = 0.75$ (upper), 0.42 (core), 0.11 (lower), suggesting that the model replicates the observations to an acceptable degree, performing better in the upper and core sublayers than in the deep sublayer.

Qualitatively, the spatial expression of seasonality in the model depicts the same features as those identified in the observational fields (Figs. 5 and 6). There is a general trend of increasing age, and decreasing A_t with along-strait distance and the inflowing tongues of younger water are present in both the basins. The northern tongue is most clearly expressed in the age_t of the upper IW waters, as in our observational analysis, and is less obvious at deeper depths. The southern basin inflowing tongue correlates with a corresponding feature in the A_t maps; however, this correlation is not as obvious in the northern basin tongue. The oldest water is amassed in the northeast of the region and regions of older/lower-amplitude water are again surrounded by younger/higher-amplitude waters in the southern basin.

However, there are also some notable discrepancies between the model and observational fields. For example, the age_t of northeastern Strait waters are ~20 days older in the model analyses compared to the observations (Figs. 5b,e), the A_t of the southern basin is larger in the model analyses compared to the observations (Figs. 6b,e), and the interior cores (coral-colored

stars in Figs. 5a,d) are more well defined in the model analyses compared to the observations. These disparities could be explained by a lack of observations in those regions rather than a model failing. Generally, the agreement between the model seasonality and that derived from observations is good, giving us confidence in the ability of SalishSeaCast to accurately reflect the Strait's circulation.

e. Eulerian mean

To assess the subsurface circulation from a Eulerian perspective, we can map IW currents from fixed locations. Here, this mapping is achieved using estimates from SalishSeaCast daily-averaged velocity fields (Fig. 7a), in addition to measurements from three ADCP moorings and nine cyclesondes/current meters in the southern basin (Fig. 7b). By projecting these velocities along a vector in the along-channel direction we can classify the velocities as up-strait (from Haro Strait toward Discovery Passage) and down-strait (the opposite).

We separate the velocity fields U into a mean $\bar{U}$ and subtidal time-varying component U' :

$$U = \bar{U} + U', \quad (7)$$

where the mean is calculated over 5 years of the hindcast dataset.

The mean Eulerian current field $\bar{U}$ (Fig. 7a) has similar features to those depicted by the seasonality tracer. Ventilation from the south occurs via a strong eastern boundary current that transports water up-strait at mean speeds as large as 16 cm s^{-1} (found within black rectangle in Fig. 7a). Upon reaching Texada Island this inflow recirculates, producing a cyclonic system within the southern basin. Two cyclonic cells are present in the southern basin, one located near the center of the basin and another located toward the north of the basin. A strong down-strait jet emanates from Discovery Passage with mean velocities up to 19 cm s^{-1} . The mean along-strait velocity field becomes intermittent at, and north of, Texada Island.

Model velocity estimates (in parentheses hereafter) in the southern basin largely agree with sparse observations from the ONC ADCP moorings, albeit with a somewhat smaller magnitude (Fig. 7b). Mean IW velocities from the East and DDL nodes are 14 (11) and 10 (5) cm s^{-1} , respectively, in an up-strait direction. Subsurface currents measured at the Central node are slower, with a mean velocity of 3 (1.5) cm s^{-1} in an approximately down-strait direction. Model currents are also consistent with the more spatially comprehensive measurements of Stacey et al. (1987; Fig. 7b) and depict a gyre-like circulation with velocities ranging from 2.8 to 8.4 cm s^{-1} (1.5–7.9 cm s^{-1}).

If we associate the mean velocity field $\bar{U}$ with advective transport and the fluctuating residual velocity field U' with a (subtidal) diffusive contribution to transport, we can assess the relative contribution of each by comparing the mean kinetic energy (MKE) and eddy kinetic energy (EKE) of the SalishSeaCast velocity fields:

$$\text{MKE} = \frac{\bar{u}^2 + \bar{v}^2}{2}, \quad \text{EKE} = \frac{u'^2 + v'^2}{2}, \quad (8)$$

FIG. 7. Map of Eulerian circulation. (a) Arrows show mean velocity field ($\bar{U}$) and contours show along-strait component of velocity calculated from SalishSeaCast velocity hindcasts from November 2014 to June 2019. (b) Inset map arrows show long-term mean currents derived from cyclesonde measurements of Stacey et al. (1987) at 100 m from June 1984 to January 1985 and three VENUS ADCP records: Central node (September 2008–May 2020); East node (September 2007–May 2020); and DDL node (May 2013–May 2020), where black arrows denote corresponding SalishSeaCast mean currents from the nearest grid cell. Data from between 50 and 150 m were used to calculate ADCP mean currents and velocity fields apart from the DDL node, where the mean is calculated between 50 and 140 m (roughly the seafloor depth). (c) The ratio of eddy kinetic energy (EKE) to mean kinetic energy (MKE), KER.

where $\bar{U} = (\bar{u}, \bar{v})$ and $U' = (u', v')$ are the horizontal velocity vectors. We can then calculate the ratio of the components at each grid cell:

$$\text{KER}_{ij} = \frac{\text{EKE}_{ij}}{\text{MKE}_{ij}}, \quad (9)$$

where subscript i denotes across-strait distance, and subscript j denotes along-strait distance, and map KER for spatial comparison (Fig. 7c).

The largest MKE contributions to the kinetic energy field are observed in the boundary current regions (Fig. 7c), however the spatial extent of these features is limited; only 25% of

the KER field is <0.75 . It should be noted that the boundary currents are also regions of high EKE, but the contribution of EKE to the overall energy of these features is small. Outside of the boundary current regions, EKE dominates in 61% of the KER field ($KER > 1.25$) and 13% of the field has an intermediate KER (0.75–1.25).

To evaluate the overall influence of advection and diffusion within the IW, we can sum the total EKE and MKE in the system and compare the two:

$$KET = \frac{\sum_{i=1}^n \sum_{j=1}^m EKE_{ij}}{\sum_{i=1}^n \sum_{j=1}^m MKE_{ij}}, \quad (10)$$

yielding a ratio $KET = 0.79$. Note that this summation only includes points within the SoG; data from Haro Strait and surrounding inlets are discarded. This ratio suggests that while the influences of diffusion and advection are similar, mean flows (i.e., advective features with high MKE) are slightly more important in transporting water and material at subtidal scales, even though eddy-diffusive transport (i.e., regions with high subtidal EKE) dominates in a large majority of the Strait. This finding provides insight into the linkages between the 1D and 3D model presented here: while a KET ratio of around 1 suggests that the balanced diffusion/advection regime of the 1D model scales up well as an overall representation of the system, is clear from Fig. 7c that there is a large amount of spatial variability in KER, and that a balanced diffusion/advection regime is not applicable on smaller spatial scales.

It is important to note that tidal variability is ignored in the Eulerian analysis (though not in the Lagrangian analysis; section 3f); however, while tidal flows are large in the shallower Haro Strait and Discovery Passage regions, they are generally small within the Strait of Georgia itself, and the influence of tides on the lateral dispersion of particles in the deeper basins of the Strait is thought to be limited.

f. Lagrangian particle tracking

The analyses above suggest that the IW circulation is dominated by a large northward flowing renewal signal, consisting of a combination of advective and eddy-diffusive effects. We can further investigate the properties of this inflow from a Lagrangian perspective by tracking the dispersion of numerical particles from a Boundary Pass release point. Particles were introduced into the velocity fields every 60 min for an entire year at 50, 75, and 100 m. In total, 26 352 particles were released. The trajectories of all existing particles in the system were calculated every 20 min until they expired after 365 days. If a particle reached the oceanic boundaries of the model, it was removed from the system. Each trajectory was smoothed using a 24 h window average to minimize tidal oscillations and decimated for computational ease. The mean daily positions of all of the particles were then spatially binned into $0.01^\circ \times 0.01^\circ$ horizontal bins and 1-m depth bins.

To calculate IW transit time fields from the Lagrangian information, which are comparable but not identical to age estimates (Bolin and Rodhe 1973), we can form a histogram of

particle ages for any grid cell in the region; this is known as the transit time distribution (TTD; van Sebille et al. 2018; Deleersnijder et al. 2001; Haine and Hall 2002). A TTD is the probability distribution that describes the transit time of particles to reach any grid cell in the system by any pathway. The mean of the TTD for a grid cell defines the average transit time for particles to travel from their release point to that grid cell, and mapping the mean transit times allows us to approximate IW age and infer circulation dynamics (Fig. 8).

The mean transit time field (Fig. 8a) depicts rapid transport through the eastern boundary current, which advects particles in an along-strait fashion toward the northern basin. Particles tend to stay in an IW tongue centered at a depth of about 100 m (Fig. 8f), although they also appear in surface and deeper waters. Average IW ages within the boundary current range from 0 to 20 days at the southern entrance of the basin to 80 days at the basin boundary. There is also a clear but less rapid transport of particles into the interior, though it is not immediately obvious whether these particles enter the interior via the boundary current, directly from the south, or some combination of both. The transit time in the interior and western extents of the basin range from 60 to 120 days—generally ~ 60 days older than the adjacent boundary current waters. We can estimate the inflow propagation speed from a Lagrangian perspective by considering the slope of the transit time/along-strait distance relationship (yellow line in Fig. 4f), yielding a propagation speed of 1.5 cm s^{-1} for the intermediate layer.

The variability in these features is illustrated by examining snapshots of particle distribution at different times since release. After 20 days, the particles are vertically distributed around the release region by strong mixing (negating any dependence on initial release depth) and subduct to fill the IW layer in the southern Strait (Figs. 8b,g), although some remain near the source. A small number of particles disperse rapidly and enter the northern basin. In 40 days they fill the entire southern basin, and begin to fill the northern Strait through Malaspina Strait, Sabine Channel, and the main channel connecting the two basins west of Lasqueti Island (LI in Fig. 1). After 80 days both basins are mostly filled, and the particles are uniformly distributed around the Strait after 160 days.

A particle budget analysis shows that the flux of particles through the upper (50 m) and lower (150 m) IW boundaries is highest in the initial week following release, where 19% of the active particles leave the IW through these boundaries—this flux is a result of strong mixing in Boundary Pass that vertically redistributes the particles (Fig. 8e). The particles then disperse through the SoG and the vertical flux of active particles out of the IW falls to $0.2\% \text{ day}^{-1}$, on average, which occurs primarily at the upper interface. This slow leakage of particles from the IW represents a combined signal of vertical mixing and upwelling (and downwelling, to a lesser degree) that is not captured in the 1D model; however, this small flux of particles represents an amount of vertical mixing that is unlikely to play a significant role in the diffusion/advection scheme presented in the 1D model on time scales less than a number of months. We can calculate the “flushing time” as the amount of time for 63% of particles to leave the layer (Monsen et al. 2002). Here, the flushing time for IW originating in Boundary

FIG. 8. Transit time and time evolution of particle positions derived from the Haro Strait particle tracking experiment. (a),(f) The mean transit time horizontally and vertically, respectively, derived from transit time distributions (TTDs). Dashed white lines depict the basin boundaries. (b)–(e),(g)–(j) The time evolution of the released particles 20, 40, 80, and 160 days after release. The particle concentration is calculated as the number of particles contained in a grid cell normalized by the grid cell containing the largest number of particles, for each time step. Particle release points are denoted by a white marker and/or black cross. Gray seafloor in (f)–(j) is representative of depth along the thalweg. The 10-m isobath of the Fraser River slopes is highlighted in (b)–(e) with a gray contour.

Pass is 128 days, in agreement with other recent SoG flushing time estimates of 125 days (MacCready et al. 2021).

g. Eddy diffusivity

We can gain quantitative insight into the dispersion characteristics of the IW via statistical analysis of the Lagrangian particle trajectories (Fig. 9). Here, we derive single particle statistics pertaining to the movement of particles in a modified coordinate system: an along-strait/across-strait axis (as indicated on Fig. 9a). For the purposes of this analysis, trajectory data from Juan de Fuca Strait, surrounding inlets, or the surface and deep layers were discarded.

Various statistical measures can be performed on Lagrangian data (see LaCasce (2008) for a comprehensive review of these techniques). The first-order moment of the particle displacements describes the mean drift of the cloud's center of mass in x direction:

$$M_x(t) = \frac{1}{N} \sum_{i=1}^N [x_i(t) - x_i(0)]. \quad (11)$$

The second-order moment of the displacements (the variance of the particles' displacement, or "dispersion") describes the spread of the cloud around its center of mass in x direction, and is calculated:

$$D_x(t) = \frac{1}{N-1} \sum_{i=1}^N [x_i(t) - x_i(0) - M_x(t)]^2. \quad (12)$$

The fourth-order moment, or kurtosis, describes the width of the displacement probability density function (PDF) in the x direction:

$$ku(x) = \frac{\sum_i (x_i - M_x)^4}{\left[\sum_i (x_i - M_x)^2 \right]^2}, \quad (13)$$

FIG. 9. Lagrangian statistics derived from particle trajectories. (a) The red arrows are a projection of the rotated along/across-strait coordinate axis with distances marked in either direction. The bathymetry of the Strait is shown, and an example trajectory is plotted in white; note that the axis is centered on the particle release point. (b),(c) The time evolution of particle drift in both directions; the colored regions show the drift of all particles binned into 500-m along-strait bins, and the white line is the mean drift of all the particles. The remaining plots show the time evolution of (d),(e) particle dispersion; (f),(g) kurtosis; and (h),(i) diffusivity. All time series are truncated to show only the first 200 days of the 365 day experiment as the system reaches quasi-steady state after 100–150 days.

where Gaussian distributed PDFs have $ku \approx 3$ and PDFs with higher “tailedness” have values $ku > 3$. Finally, we calculate the relative horizontal diffusivity of the system as the time derivative of the dispersion in x direction:

$$K = \frac{1}{2t} \frac{d}{dt} D_x. \quad (14)$$

There is an increased spreading of particles in the along-strait direction. This spreading is apparent in the mean drift and dispersion characteristics (Figs. 9b–e)—on average, from their release point in Haro Strait, particles drift 93 km along-strait and disperse over a 3000 km^2 area, compared to 15 km across-strait drift and 250 km^2 dispersion.

This spreading is also reflected in the kurtosis characteristics of the displacements (Figs. 9f,g): kurtosis values, large initially

where particles are grouped around the release point, fall to $ku \approx 3$ after two weeks as particles disperse from the center of mass. In the across-strait direction, the kurtosis then increases to $ku \approx 5$ —physically, this increase indicates that particles become grouped in the interior of the basins. In the along-strait direction, the kurtosis briefly increases to a $ku \approx 3$ maxima after 40 days, before decreasing to $ku < 3$ after 120 days; the final state of the system is a flat distribution of particles spanning the along-strait extent.

The relative diffusivity measured here reflects the system’s ability to laterally spread particles (Figs. 9h,i). Peak across-strait diffusivity values of $160 \text{ m}^2 \text{ s}^{-1}$ occur near the beginning of the experiment, falling to zero after ~ 20 days. In the along-strait direction, diffusivity values as large as $560 \text{ m}^2 \text{ s}^{-1}$ occur near the beginning of the experiment, falling to zero after

~100 days. Recent observational estimates of SoG surface layer diffusivity— $4000 \text{ m}^2 \text{ s}^{-1}$ at the surface (Pawlowicz et al. 2019) compared to $88\text{--}167 \text{ m}^2 \text{ s}^{-1}$ for the IW—indicate that substances disperse more rapidly at the surface, presumably due to the direct effects of wind and river inflow, than at depth.

The reduction in diffusivity to ~ 0 as a function of time (or “filling time scale”) is partially a product of the aspect ratio of the Strait, whereby particles take less time to spread across the Strait than along it due to the elongated shape of the basin. Thus, by assessing the diffusivities from $t < \text{filling time scales}$, we can approximate the diffusivity in either direction prior to the time when particles have maximally dispersed. The mean diffusivity values for across(along)-strait in the first 20 (100) days is 83 (181) $\text{m}^2 \text{ s}^{-1}$. Comparing the two measures, we see that the mean diffusivity values are around twice as large in the along-strait direction: the Eulerian mean velocity and KER fields (Fig. 7) indicate that the major advective (MKE) features are directed along-strait. This finding suggests that the along-strait diffusivity represents components of both advective and diffusive transport, whereas across-strait diffusivity reflects mainly diffusive transport.

4. Discussion

Previous research on the SoG intermediate circulation has mainly focused on patterns of circulation in specific regions of the Strait (Stacey et al. 1987), or on the general estuarine exchange nature of the layer (Pawlowicz et al. 2007; Riche and Pawlowicz 2014). The maps and transit times presented in Figs. 5–9 aim to provide a more comprehensive overview of the mean IW circulation and to highlight the major pathways through which IW is transported. These findings are particularly significant when considering the subsurface circulation of the northern basin, where very little information is currently available.

The general northward progression of seasonal cycle phase, first noted by LeBlond et al. (1991), has been carefully reevaluated for the whole Strait using a much more comprehensive dataset. A new feature of this progression is a corresponding decay in cycle amplitude. We can use the measured advection and diffusion properties to reconcile these findings within the context of the 1D model. To calculate $\gamma = 2K/U$, we use peak along-strait relative diffusivity of $k = 560 \text{ m}^2 \text{ s}^{-1}$ to represent the unrestricted spreading of particles in an along-pipe direction (Fig. 9h) and the observed and modeled propagation speeds (slopes of relationships in Fig. 4f) to estimate an advection velocity range of $U = 1.5\text{--}2 \text{ cm s}^{-1}$, producing γ range of $56\text{--}75 \text{ km}$ (blue shaded region in Fig. 2g). This length scale is less than the modeled γ of 100 km (Fig. 2b) but within the theoretical range necessary to produce a physically realistic simulation of the tracer behavior in the Strait (Fig. 2g). This observation, in addition to the striking qualitative similarities between measured seasonality in the Strait (Fig. 4) and the intermediate diffusion–advection case in the 1D pipe model (Fig. 2b), suggests that, in a time-averaged sense, the first-order along-strait renewal processes can be well represented by a balanced advection–diffusion relationship [Eq. (1)]. However, it is clear from the various spatial analyses performed here

(Figs. 5–8) that this scheme only serves as a very general description of IW flow, and that multiple inflows and circulation cells contribute to a higher-order variability in advection–diffusion characteristics that cannot be rendered by a simple 1D approximation.

If we assume that mean transport crosses isopleths of age and cycle amplitude, we find, overlaid on the general northward propagation within the Strait, an anticlockwise circulation pattern, with faster northward speeds on the eastern side of the Strait and slower (or even southward) velocities on the western side. The $\bar{U}$ field derived from SalishSeaCast (Fig. 7a) also outlines an anticlockwise gyre-like circulation in the southern basin in agreement with the long-term but sparse observations of the ONC nodes and earlier current meter deployments (Fig. 7b; Stacey et al. 1987), with up-strait currents as large as 16 cm s^{-1} compared to down-strait currents up to 9 cm s^{-1} . However, while the tracer and $\bar{U}$ fields indicate a gyre-like circulation is present, it should be noted that the KER field identifies only the eastern limb of this loop as a strongly advective feature. This finding suggests that the eastern boundary current is the most strongly expressed feature of the circular flow, and that the rest of the gyre-like system is only expressed when we consider mean transport on long time scales. Further analysis of Lagrangian trajectories might answer how prevalent the cyclonic circulation is on subinterannual time scales—techniques such as most-likely path algorithms (van Sebille et al. 2012; Maximenko et al. 2012) or Lagrangian Coherent Structures analyses (Haller 2015), for example.

No obvious transport pathway can be found between the southern and northern basins at intermediate depths. Circulation patterns in Malaspina Strait suggest some exchange between Malaspina Strait and the southern basin (Fig. 7), but there is no clear communication between Malaspina Strait and the interior of the northern basin, the boundary between which is silled at a depth of $\sim 100 \text{ m}$. A measure of interbasin exchange occurs between Lasqueti Island and Vancouver Island (Figs. 8b–e), though the passage of water through this channel is likely impeded by shallow sills that separate the two basins (Fig. 9a). Sabine Channel—a 240-m -deep but very narrow corridor between Texada Island and Lasqueti Island—may then represent an important pathway between the two basins, though closer analysis of the transport through this channel is necessary to confirm this notion. Interestingly, there is no clear evidence of this interbasin topographic barrier in the linear evolution of age_t ($\sim 150\text{-km}$ along-strait distance in Fig. 4), suggesting that, regardless of the specific transport pathways, the transport between basins is fairly efficient.

Within the seasonality tracer framework we assume a constant intermediate transport. This assumption is supported by the strongly sinusoidal nature of seasonal cycles themselves (Figs. 4b,e,h) and the robust linear relationship between northward excursion and seasonal cycle phase (Figs. 4c,d,f,g,i,j), suggesting constant, near-constant, or frequent IW renewal from Haro Strait. This premise is opposed to that of an episodic signal of renewal, which would produce temperature signals with steplike variability that is

not apparent in our dataset. However, this assumption conflicts with studies that have linked subsurface inflow strength in the southern basin to discrete deep water renewal events (LeBlond et al. 1991; Masson 2002). Therefore, we propose that a constant inflow assumption is valid when investigating the time-averaged circulation of the region, but the time-varying nature of the inflow should be considered when assessing circulation on interannual or shorter time scales. Evaluating the seasonal and annual inflow strength variability is an important next step in determining the dispersion of contaminants through the SoG on shorter time scales, and will be aided by the rapidly expanding regional monitoring and modeling efforts, though this analysis is beyond the scope of this study.

The linear relationship between along-strait distance and both ϕ_{CD} and transit time (Fig. 4f), rendered in the toy model as a constant advection term U [Eq. (1)], suggests that the southern inflow ventilates the system steadily northward. This steady up-strait transport may not be an intuitive conclusion when studying the bidirectional flow of the Eulerian mean current field (Fig. 7a); however, the KER field (Fig. 7c) identifies the boundary current jets as the only strongly advective features, suggesting that these features are the only components of the flow field in which the mean flow is significantly expressed. As these features are oriented in the along-strait direction, we can consider the along-strait trends in age and transit time (and our model term U) as the sum product of all the processes that contribute to the mean northward propagation of the renewal signal, including the mean up-strait influence of advective features and the mean up-strait contribution of mixing and stirring. We can then class the residual variability—the spatiotemporal variability of the flow field, seen as the scatter in age around the age, trend and parameterized as K in the 1D model—as the higher-order processes that impact IW transport.

Finally, regional pollutant transport may be particularly sensitive to the subsurface circulation, as the majority of wastewater from the densely populated southern Strait region is incorporated into the IW directly via wastewater outfalls or indirectly via subducting water in Haro Strait. As a result of elevated wastewater levels in the IW, various semiconservative chemical tracers (such as polybrominated diphenyl ethers, or PBDEs; Sun et al. 2018) are concentrated within the intermediate layer. Further study into the dispersion characteristics of these chemical tracers throughout the IW may provide an approach to link the physical circulation with biogeochemical variability observed in the region. The Lagrangian analysis of the southern inflow (Figs. 8 and 9) suggests that small amounts of long-lived pollutants will spread quickly through the entire Strait in approximately one month while bulk spreading occurs on the scale of 3–4 months, after which the system is in quasi-steady state.

5. Conclusions

Here, we have used three different approaches to produce an intermediate circulation scheme for the Strait of Georgia. We can use this three-pronged approach to compare the strengths and weaknesses of the different methodologies.

Tracer-based techniques incorporate the influence of both advection and diffusion into transport estimates, though separating the contributions of each to the overall transport requires additional numerical analysis [e.g., Eqs. (1)–(3)] which, in the case of some tracers, may not be possible. The analysis of Eulerian velocity fields can create simple, intuitive maps of mean currents—indeed, it is easy for one to look at the arrows designating current velocities (as in Fig. 7) and draw simple pathlines to join them, giving a general sense of how water is transported through the system—however, these maps can be misleading unless we weigh the influence of time-invariant features against the time variability of the flows (e.g., Fig. 7c). Lagrangian techniques perhaps represent the most complete three-dimensional representation of the regional circulation, but are complicated observationally, numerically, and analytically. Of the three approaches, the seasonality tracer is distinct; while the analysis of velocity fields and Lagrangian trajectories are commonplace in circulation studies, the seasonality tracer derived here has provided a novel technique for identifying intermediate pathways and mixing dynamics that could plausibly be used to map circulation in other subducting water masses with cyclical variability. While there is overlap in the findings of each method, no one single method provides a complete explanation and the resulting intercomparison between results has provided a more rigorous characterization of the intermediate circulation, emphasizing that a multifaceted approach to circulation problems is advantageous, where possible.

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Data availability statement. The Pacific Salmon Foundation hydrographic and Ocean Networks Canada ADCP datasets can be accessed at <https://www.oceannetworks.ca/data-tools>. SalishSeaCast outputs can be accessed via the SalishSeaCast ERRDAP server: <https://salishsea.eos.ubc.ca/erddap/index.html>. All custom code used in this study can be found at <https://doi.org/10.5281/zenodo.4603254>.

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