

From Ice Shells to Subsurface Oceans: Geophysical Fluid Dynamics of Icy Moons

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Key Points:

- Convection, tides, and baroclinic eddies interact to control heat and momentum transport in subsurface oceans of icy moons.
- Ice thickness variations (ITV) constrain heat partitioning: sustaining large ITV requires most heat production to occur within the ice shell.
- The direction of non-synchronous rotation constrains whether the subsurface ocean is in the convective or tidally-driven stratified regime.

Abstract

Several icy satellites are believed to host subsurface oceans beneath their outer ice shells. These oceans are promising targets in the search for extraterrestrial life, but their internal dynamics remain largely hidden from direct observation. The overlying ice shell, however, is relatively accessible to remote sensing and may serve as a window into the ocean below. Because the subsurface ocean transports heat and momentum, it influences the ice shell in observable ways: heat transport drives spatially variable freezing and melting, modifying the ice thickness variations (ITV), while momentum transport induces non-synchronous rotation (NSR) of the ice shell relative to the rocky interior. In this paper, we review how convection, tides, and baroclinic eddies together control heat and momentum transport in icy moon oceans. Using Enceladus as a case study, we demonstrate how the observed ITV and NSR can be used to infer the ocean's equation of state and the partitioning of heat production among the ice shell, ocean, and silicate core. We conclude that the majority of heat production on Enceladus, of order 10 GW, is likely to occur within the ice shell. We also outline strategies for cross-validating these inferences with independent observations from upcoming missions.

Plain Language Summary

Oceans beneath the ice shells of icy moons such as Enceladus and Europa are among the most promising places to search for life beyond Earth. However, these oceans are hidden beneath thick ice, making them difficult to study. Fortunately, the ice shell itself can offer important clues. Because fluid flow in the liquid layer can transport heat and momentum, changes in the ocean can affect the shape and movement of the ice shell above it. For example, heat transport by ocean currents can change the ice thickness pattern by inducing freezing and melting, and momentum transport by ocean currents can cause the ice shell to slowly rotate relative to the moon's rocky core. In this work, we explain how ocean processes – such as convection, tides, and large swirling currents called baroclinic eddies – transport heat and momentum around the ocean. Using the moon Enceladus as an example, we describe how studying the ice shell can help us learn about the properties of the underlying ocean and how the generation of heat by tidal energy dissipation is distributed between the ocean, the ice shell, and the core. Finally, we discuss approaches that future missions could adopt to verify these inferences.

1 Introduction

Several icy satellites in the Solar System, including Enceladus, Europa, and Ganymede, are believed to host global subsurface oceans that are potentially habitable (K. Khurana et al., 1998; Zimmer et al., 2000; Thomas et al., 2016; Nimmo & Pappalardo, 2016). These oceans do not receive direct sunlight, ruling out photosynthesis as a mechanism to produce the chemical energy needed to support life. Nevertheless, life may be sustained by long-lived chemical gradients (Peng et al., 2025; Cockell et al., 2016; Vance et al., 2018; Glein & Zolotov, 2020; Vance et al., 2023; Cockell et al., 2024; Fisher et al., 2024). The reduced chemical species produced by water-rock reactions on the ocean floor (Deamer & Damer, 2017; Sekine et al., 2015; McCollom & Bach, 2009) can react with oxidants delivered from the ice shell above (Carlson et al., 1999; Chyba, 2000; Johnson et al., 2003; K. P. Hand et al., 2007), providing a potential energy source for life. Ocean circulation is therefore essential to connect these spatially separated redox reservoirs (McGillicuddy Jr et al., 1998; Falkowski et al., 1998; Ducklow et al., 2001; M. Freilich et al., 2021; M. A. Freilich et al., 2022). Furthermore, if such subsurface oceans are inhabited, active circulation would also be required to transport potential biosignatures upward to the ice shell, where they might subsequently be detected (e.g., German et al., 2022).

One of the most important yet poorly constrained controls on tracer transport is the partitioning of heat production among the ice shell, ocean, and silicate core. When

heat is generated predominantly in the core or ocean, the ocean tends to be convective, allowing efficient vertical transport of tracers from the seafloor to the ice shell. By contrast, when heat is produced mainly within the ice, especially with significant spatial variations, the resulting lateral temperature and salinity gradients at the ice-ocean interface may stratify the ocean, in much the same way that poleward-weakening solar heating stratifies Earth’s atmosphere and oceans (Schneider, 2006). Stratification suppresses vertical exchange, forcing tracers to follow longer, more convoluted pathways.

The circulation in the subsurface ocean and the heat partitioning, however, are hidden beneath ice shells tens of kilometers thick, making direct quantification extremely challenging. Fortunately, the ice shell itself can potentially serve as a diagnostic window into the hidden ocean. Because the ocean is the only liquid layer between the silicate core and the surface, it can redistribute heat and momentum in ways neither solid layer can. This redistribution leaves observable fingerprints: oceanic heat transport produces localized freezing and melting that shape ice thickness variations (ITV) (Kang & Jansen, 2022; Kang, 2022; Kang, Mittal, et al., 2022); oceanic momentum transport drives non-synchronous rotation (NSR) of the ice shell (Ashkenazy et al., 2023; Hay et al., 2023; Kang, 2024; Hay et al., 2024). Both ITV and NSR are potentially measurable (Iess et al., 2014; Hemingway & Mittal, 2019; Hoppa et al., 1999; Schenk & McKinnon, 2024).

Over the next decade, NASA’s Europa Clipper and ESA’s JUICE spacecraft will arrive at Jupiter’s icy moons Europa and Ganymede (Pappalardo et al., 2024; Grasset et al., 2013), returning unprecedented measurements of gravity, magnetic fields, ice-shell structure, and rotational states (Roberts et al., 2023a; Van Hoolst et al., 2024). With observational constraints on ITV and NSR in hand, ocean heat and momentum transports can be estimated, which can then be inverted to constrain ocean properties and forcing conditions. Although such retrievals are inherently non-unique, cross-validation with independent observations can substantially narrow the uncertainties.

This paper is structured as follows. In §2, we review the three key dynamical processes present in icy moon oceans: convection, tides, and baroclinic eddies. In §3, we discuss the interactions between these processes, with a focus on ocean heat transport (OHT) and ocean momentum transport (OMT), and §4 provides an energy perspective that constrains the intensity of ocean circulation on icy satellites. In §5 and §6, we demonstrate how observable ITV and NSR can be used to infer ocean properties and forcing conditions. Finally, in §7, we discuss means to cross-validate these inferences through independent observations. Readers less interested in geophysical fluid dynamics may skip the detailed discussions of individual dynamical processes in §2.

This review identifies two pathways for inferring subsurface ocean conditions.

Constraining heat partitioning between core, ocean and ice from ITV.

Large-amplitude planetary-scale ITV requires that the majority of heat production occurs within the ice shell rather than in the ocean or core. This conclusion follows because spatially inhomogeneous heating in the ice directly drives local freezing and melting that maintains ITV, whereas inhomogeneities in oceanic or core heating are smoothed by ocean circulation under the thermal/salinity forcing from the ice shell.

Diagnosing the dominant ocean dynamics from NSR direction.

The direction of NSR can distinguish between a convection-dominated, well-mixed ocean and a stratified ocean dominated by baroclinic eddies. Convection and tides drive NSR in opposite directions to that driven by baroclinic eddies, providing a diagnostic of the dominant ocean regime. The potential westward NSR inferred for Enceladus from tidal-bulge offset (Schenk & McKinnon, 2024) is consistent with the ice-dominant heat partitioning derived from the observed ITV.

2 The Three Drivers of Subsurface Ocean Dynamics: Convection, Tides, and Baroclinic Eddies

As sketched in Fig. 1a, subsurface oceans on icy satellites are forced by three primary mechanisms: 1) **Bottom heating** from radiogenic and tidal dissipation in the silicate core, 2) **Tidal forcing** from gravitational interactions with the host planet and adjacent moons, and 3) **Surface buoyancy forcing** from heat and salinity exchange with the ice shell. These forcings drive three corresponding dynamical responses, **convection**, **tidal waves**, and **baroclinic eddies**, each with distinct signatures for heat and momentum transport.

In this section, we first review each forcing-response pair in isolation: convection driven by bottom heating (§2.1), tidal waves driven by mechanical tidal forcing (§2.2), and baroclinic eddies driven by surface buoyancy forcing (§2.3). Throughout, we emphasize how heat and momentum are transported in the ocean, providing a perspective complementary to previous reviews (Soderlund et al., 2024; Vance & Goodman, 2009; Soderlund et al., 2020). We do not consider ocean dynamics driven by magnetic induction pumping (Gissinger & Petitdemange, 2019), which has been reviewed by Soderlund et al. (2024).

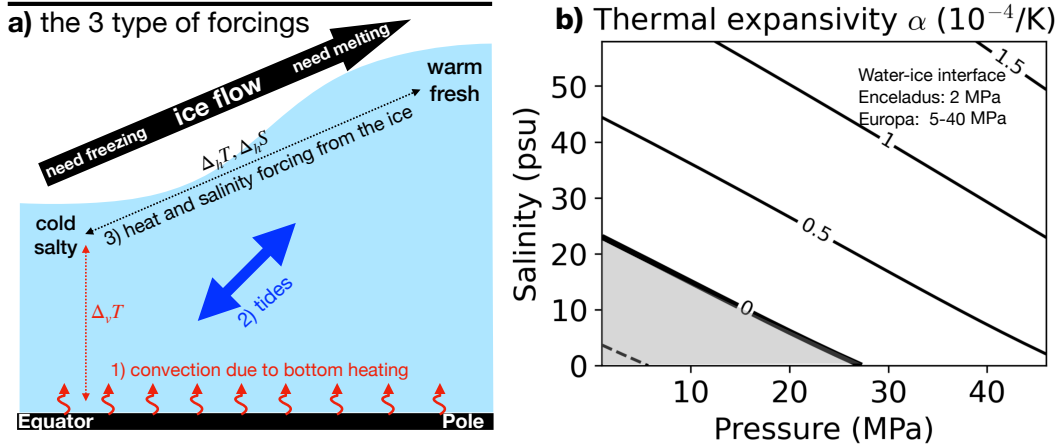


Figure 1: (a) Schematic illustration of the three types of forcings exerted on icy moon oceans. (b) Thermal expansivity α_T of water at freezing point as a function of pressure and salinity. Shading indicates anomalous expansion.

2.1 Convection due to bottom heating

The first forcing comes from heat released by the silicate core, generated by tidal dissipation and radiogenic decay (Běhounková et al., 2021). When the thermal expansivity is positive, this bottom heating drives *convection*: hot buoyant fluid rises, cold dense fluid sinks, and the resulting overturning mixes heat and tracers vertically from the seafloor toward the ice shell (Soderlund et al., 2014, 2020; Ashkenazy & Tziperman, 2021; Bire et al., 2022).

Nondimensional numbers: Ra, E v.s. Ro^* . Traditionally, rotating convection is characterized by two nondimensional numbers: the Rayleigh number $Ra \equiv \alpha_T g \Delta_v T D^3 / (\nu \kappa)$, which measures the degree of supercriticality, and the Ekman number $E \equiv \nu / (2\Omega D^2)$, which measures the relative importance of viscosity and rotation. Here D is ocean depth, α_T thermal expansivity, $\Delta_v T$ the vertical temperature contrast, Ω the rotation rate, and ν and κ are viscosity and thermal diffusivity. In simulations that do not resolve tidal wave

breaking, ν and κ are usually set to represent turbulent viscosity and diffusivity, which are orders of magnitude greater than molecular values. Using Ra to describe convection requires $\Delta_v T$ to be known a priori. For icy-moon oceans, $\Delta_v T$ is typically unknown, whereas the heat flux Q is better constrained because it should not exceed the conductive heat loss through the ice shell in an equilibrium state. Replacing $\Delta_v T$ with Q leads to the flux Rayleigh number $Ra_F \equiv \alpha_T g Q D^4 / (\rho C_p \nu \kappa^2)$.

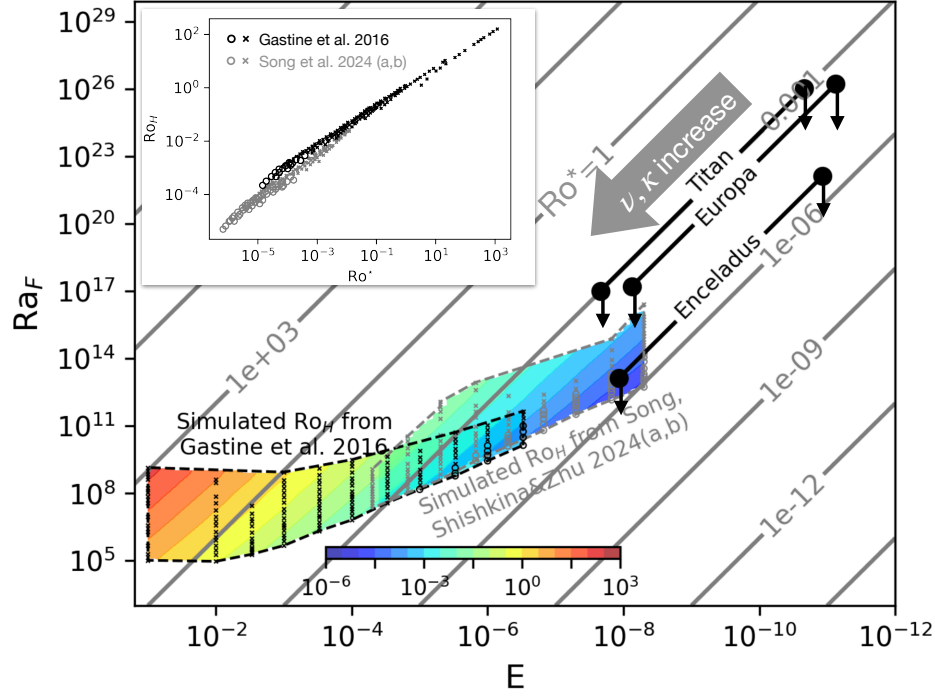


Figure 2: The parameter regimes relevant to Enceladus, Europa, and Titan are marked in a parameter space spanned by the flux Rayleigh number Ra_F and the Ekman number E , assuming a bottom heat flux on the high-end (40, 40 and 10 mW/m^2 respectively). The lower bound of viscosity and diffusivity is set to $10^{-6} m^2/s$, representing the molecular value, and the upper bound is set to $10^{-3} m^2/s$, representing potentially enhanced mixing by tides. Isolines of Ro^* are shown in gray contours. Since Ro^* (Eq.1) is independent of diffusivity by definition, increasing/decreasing ν and κ corresponds to a shift toward the bottom-left/top-right corner. The shading shows the simulated Rossby number $Ro_H \equiv U_{rms}/(\Omega D)$ from the DNS simulations in Gastine et al. (2016) and in Song et al. (2024b, 2024a), where U_{rms} denotes rms velocity. The locations of the original data points are marked by circles (experiments within the diffusiveless regime $Ra E^{8/5} < 0.4$) and crosses (all others). Since the shading isolines are nearly aligned with those of Ro^* , the diffusiveless output Ro can almost be expressed as a function of the diffusiveless predictor Ro^* alone and has weak sensitivity to ν, κ . The inset shows Ro_H scattered against Ro^* .

Because the molecular viscosity and diffusivity of water are only $\mathcal{O}(10^{-6} m^2 s^{-1})$ or smaller, a bottom heat flux of $\mathcal{O}(10 mW m^{-2})$ produces $Ra_F \sim \mathcal{O}(10^{22})$ for Enceladus and $\mathcal{O}(10^{27})$ for Europa, implying heat transport being dominated by advection and buoyancy force being dominant over viscous force. The Ekman numbers, $E \sim \mathcal{O}(10^{-11})$ and $\mathcal{O}(10^{-12})$, moreover indicate viscous forces being overwhelmed by the Coriolis force due to planetary rotation (Soderlund, 2019). In this high- Ra_F , low- E regime, viscosity and diffusivity affect the dynamics only at the smallest scales (especially the millimeter-scale boundary layers), while large-scale interior dynamics are governed by the prescribed heat

flux and rotation rate and are relatively insensitive to diffusivity and viscosity, when Q instead of ΔT is prescribed. In fact, a diffusiveless scaling emerges if the boundary-layer bottleneck is bypassed by imposing sufficiently large diffusivity, achieved when $\text{Ra} \cdot \text{E}^{8/5} < 0.4$ (Gastine et al., 2016). One may argue that boundary layer turbulence generated as tidal/librational flow rubs against the rough boundaries can efficiently mix out the millimeter-thin boundary layer.

The possible insensitivity of ocean interior dynamics to viscous and diffusive processes motivates the introduction of nondimensional numbers that do not explicitly involve ν or κ . Assuming Prandtl number $\text{Pr} \equiv \nu/\kappa = 1$, which holds if mixing is dominated by tidal wave breaking, we can eliminate ν and κ between Ra_F and E to obtain the natural Rossby number:

$$\text{Ro}^* \equiv \text{Ra}_F^{1/2} \text{E}^{3/2} = \left(\frac{g \alpha_T Q}{8 \rho C_p \Omega^3 D^2} \right)^{1/2}. \quad (1)$$

Ro^* was introduced in the 1990s in the Earth oceanography community and has been widely used since (Boubnov & Golitsyn, 1990; Jones & Marshall, 1993; Maxworthy & Narimousa, 1994). It quantifies the influence of rotation on the dynamics, and is related to the modified flux Rayleigh number Ra_F^* through $\text{Ra}_F^* \sim (\text{Ro}^*)^2$. In the diffusiveless regime, it has been shown to scale with the posterior Rossby number $\text{Ro}_H \equiv U/(D\Omega)$ which characterizes the number of rotation periods $2\pi/\Omega$ needed for convection to penetrate the ocean depth D (Christensen & Aubert, 2006; J. M. Aurnou et al., 2020). Readers interested in the conversion among the various nondimensional numbers in the Q -based and $\Delta_v T$ -based systems may refer to J. M. Aurnou et al. (2020).

In Fig. 2, the relevant parameter regimes for Enceladus, Europa, and Titan are marked in the parameter space spanned by Ra_F and E , assuming a heat flux at the higher end, with ν and κ ranging from 10^{-6} m²/s, representing the molecular value, to 10^{-3} m²/s, representing possible mixing enhancement by tidal wave breaking. Isolines of Ro^* are overlaid for reference. To illustrate the system's weak dependence on ν and κ , we show the Rossby number $\text{Ro}_H \equiv U_{\text{rms}}/(\Omega D)$ from all DNS simulations with $\text{Nu} > 2$ reported in Gastine et al. (2016) and Song et al. (2024b, 2024a). This Ro characterizes the interior velocity intensity relative to planetary rotation Ω without invoking ν or κ as a reference scale, as the Reynolds number does. Across all cases shown, Ro exhibits only weak dependence on ν and κ , both within the diffusiveless regime ($\text{Ra} \cdot \text{E}^{8/5} < 0.4$, denoted by circles) and outside it (denoted by crosses).

Estimates of Ro^* for a range of icy moons are summarized in Table. 1. Given that icy moon oceans usually have $\text{Ro}^*, \text{Ro}_H \ll 1$ (see Table. 1), the convective dynamics on icy satellites should be strongly influenced by planetary rotation, at least in the interior.

The diffusiveless scaling (Gastine et al., 2016) also provides an estimate for the interior temperature difference $\Delta_v T$, defined as the temperature contrast from the top of the bottom boundary layer to the bottom of the top boundary layer.

$$\Delta_v T \sim \left(\frac{Q \Omega^2}{0.15 \rho C_p} \right)^{2/5} \left(\frac{D}{(\alpha_T g)^3} \right)^{1/5}. \quad (2)$$

This relationship is commonly used to convert the heat flux-based Ra_F into the temperature-based Ra (Soderlund et al., 2014; Soderlund, 2019; Schoenfeld et al., 2023; Kverka & Čadek, 2024) to set up experiments with fixed temperature boundary conditions. Table. 1 summarizes the estimates of $\Delta_v T$ for a set of icy moons, which is usually less than 10 mK.

Heat and momentum transport. Under the influence of planetary rotation, even if the bottom heat flux is uniformly distributed, the convective dynamics are different between the polar regions and the equatorial regions, demarcated by the tangent cylinder (denoted in Fig. 3a). Over the poles, where the axis of rotation is generally aligned

	Enceladus	Europa	Titan
a (km)	250	1560	2575
H_0 (km)	~ 20	~ 20	O(100)
ΔH (km)	20	< 2	5
D (km)	40	80	O(100)
g (m/s ²)	0.113	1.3	1.35
Ω (1/s)	5.3e-5	2.1e-5	4.5e-6
α_T (1/K)	$< 1e-4$	3e-4	O(1e-4)
Q_b (mW/m ²)	< 40	< 40	$< O(10)$
Ro^*	$< 7e-6$	$< 9e-5$	$< 2e-4$
$\Delta_v T$ (mK)	$\lesssim 7$	< 1	< 0.1
$\Delta_h T$ (mK)	150	< 180	470
U_Q (cm/s)	< 1	< 6	< 2
U_T (cm/s)	27	27	17
Q_c (mW/m ²)	1e3	$< 3e4$	2e6
ΔH_c (km)	4	0.06	0.04

Table 1: Icy moon parameters. a : satellite radius; H_0 : assumed mean ice thickness (Thomas et al., 2016; K. Hand & Chyba, 2007); ΔH : ice thickness variation (Hemingway & Mittal, 2019; Nimmo et al., 2007); D : assumed mean ocean depth; g : gravity; Ω : rotation frequency; α_T : assumed thermal expansivity; Q_b : bottom heat flux which should not exceed the conductive heat loss through the ice shell; Ro^* : natural Rossby number following Eq. (1) ($Ra_F^* = (Ro^*)^2$); $\Delta_v T$: vertical temperature difference sustained by Q_b following Eq. (2); $\Delta_h T$: temperature difference along water-ice interface due to ITV; U_Q : maximum ocean characteristic speed assuming all heat is produced in the silicate core (Eq. (20)); U_T : maximum ocean characteristic speed assuming all heat is produced in the ocean; Q_c : critical bottom heat flux to overcome stratification enforced by ice forcings (Eq. (8)) assuming $\kappa_v = 0$. ΔH_c : minimum ITV needed to stratify the upper ocean in presence of a bottom heat flux Q_b . We caution that Titan may not possess a subsurface ocean (Petricca et al., 2025), and that the numbers quoted above should be interpreted as illustrative rather than definitive. Similar calculation may be done for Uranus moons once we confirm the existence of subsurface ocean(s).

with gravity, convection mostly manifests itself as spiral plumes along the rotational axis, whereas in low latitudes, where the axis of rotation is perpendicular to the direction of gravity, convection mostly appears as overturning rolls along the equatorial plane. This feature was first suggested by Busse and Cuong (1977) and confirmed by numerous numerical studies (J. Aurnou et al., 2007; Soderlund et al., 2014, 2020; Amit et al., 2020; Gastine & Aurnou, 2023; Bire et al., 2022). Shown in Fig. 3b-d are snapshots of simulations taken from Bire et al. (2022), clearly demonstrating the different nature of fluid dynamics inside and outside the tangent cylinder. The equatorial mode has a lower instability threshold and dominates the polar mode near the onset of convection (Busse, 1970; Dormy et al., 2004). As the Rayleigh number increases far beyond the marginally unstable Rayleigh number, the polar modes take over (Amit et al., 2020; Bire et al., 2022; Kvorka & Čadež, 2022; Gastine & Aurnou, 2023; Hartmann et al., 2024; Gastine & Favier, 2025). Depending on which mode is more efficient in heat transport, bottom heating may be redistributed from the polar regions to the equatorial regions (if equatorial mode dominates), or the other way around (if polar mode dominates) by up to $\sim 20\%$. Several criteria have been proposed to distinguish the equatorward and poleward heat-convergence regimes, but no consensus has emerged. Regardless, the high supercriticality characteristic of icy moon oceans suggests that they probably reside in the poleward heat-convergence

regime (Kvorka & Čadež, 2022; Hartmann et al., 2024; Gastine & Favier, 2025). Furthermore, if core heating is spatially nonuniform, the inhomogeneity of the heat flux delivered to the ice shell can be even more pronounced (Lemasquerier et al., 2023; Terra-Nova et al., 2023).

This inhomogeneous heat flux can potentially induce ice thickness variations (Amit et al., 2020; Kvorka & Čadež, 2022; Lemasquerier et al., 2023; Hartmann et al., 2024; Gastine & Favier, 2025). But one should note that, once the ice thickness variations emerge, the pressure dependence of the freezing point induces a substantial temperature contrast at the water-ice interface (Fig. 1a). This temperature gradient can drive heat from thin-ice regions (warmer interface) toward thick-ice regions (colder interface), eroding the developing ice thickness gradients. We elaborate on the mechanisms when talking about the interaction between baroclinic eddies and convection in §3.

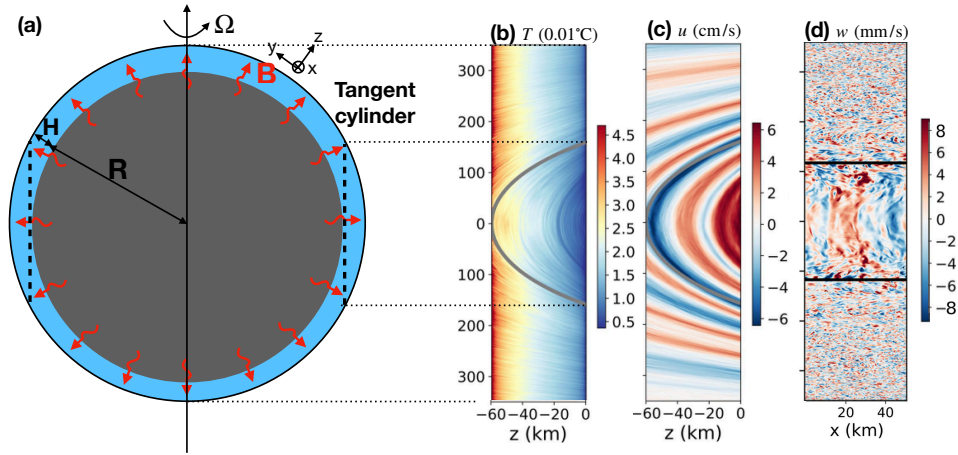


Figure 3: Ocean convection driven by bottom heating. Panel (a) sketches the geometry, setup and the definition of tangent cylinder. Panels (b,c) are temperature and zonal velocity (along x -axis) snapshots across a meridian and panel (d) shows the vertical velocity in the middle level of the ocean. Figures taken from Bire et al. 2022.

Besides heat, momentum can also be redistributed by convection. At the equator, the convective flow is organized into rolls (Fig. 3d), known as thermal Rossby waves, which exhibit a prograde tilt with height. This geometry enables the upward flux of prograde momentum from the bottom of the ocean toward the ice shell (Busse, 1970; Dormy et al., 2004). When this thermal Rossby mode is present without being too disturbed by turbulence, a prograde jet would form near the equator (Fig. 3c). This nature of thermal Rossby waves has motivated the hypothesis that the ice shell of icy satellites may undergo non-synchronous rotation (NSR) (Ashkenazy et al., 2023; Hay et al., 2023). We note that the analyzes by Busse (1970); Dormy et al. (2004); Busse and Hood (1982); Cardin and Olson (1994) were conducted near the onset of convection. However, as supercriticality increases, dynamics can eventually escape the constraint imposed by planetary rotation ($\text{Ro}^* > 1$), and the momentum transport and the direction of the resultant NSR can reverse from prograde to retrograde (J. Aurnou et al., 2007; Soderlund et al., 2014; Bire et al., 2022; Hay et al., 2023).

In addition to the equatorial jet, Fig. 3c reveals numerous jets in each hemisphere aligned with the rotation axis. The formation of jets perpendicular to the rotation axis is common in rapidly-rotating fluids such as the atmospheres of Jupiter and Saturn (characterized by small Rossby numbers) and is insensitive to how dynamics are sustained energetically (Galperin et al., 2008; Srinivasan & Young, 2012; Farrell & Ioannou, 2008).

Although these alternative jets can also exert torque onto the ice shell and the silicate core, the net torque is much weaker than the torque exerted by each of the individual jets, as they tend to cancel each other out. In fast-rotating systems ($Ro^* \ll 0$), the thermal wind balance should be well satisfied due to the dominance of planetary rotation in icy moon oceans (Kang, 2024). This relationship may be used to estimate the total momentum exchange between the ocean and ice, as will be elaborated in §6.

Convective plume driven by localized heat sources. Motivated by the potential role of seafloor hydrothermal activity, a broad literature has examined convection driven by localized point or line heat sources in rotating fluids (Goodman et al., 2004; Goodman & Lenferink, 2012; Kang, Marshall, et al., 2022; Bire et al., 2023; Zhang et al., 2025; Wang et al., 2025). Building on foundational laboratory and theoretical studies from the 1970s and 1990s (Saunders, 1973; Jones & Marshall, 1993; Maxworthy & Narimousa, 1994; Visbeck et al., 1996; Legg et al., 1996; Jacobs & Ivey, 1998, 1999; Bush & Woods, 1999; Okada et al., 2004), recent studies extend the analysis to parameter regimes relevant for icy-moon oceans. These studies confirm earlier findings that line plumes remain concentrated only over a finite vertical distance (Bush & Woods, 1999) and further show that point plumes exhibit similar behaviors given enough space in the vertical direction.

By conducting large-domain numerical simulations and synthesizing previous laboratory and numerical results, Wang et al. (2025) demonstrated that the evolution of a point plume under polar configuration (i.e. planetary rotation aligned with gravity) is determined by the point-source natural Rossby number

$$Ro^{*p} \equiv \left(\frac{F_0}{8\Omega^3 D^4} \right)^{\frac{1}{4}}, \quad (3)$$

where $F_0 = \iint \frac{\alpha_T g Q}{\rho_0 C_p} ds$ is the area-integrated buoyancy flux, with Q denoting heat flux in W/m^2 . Similar to Ro^* , Ro^{*p} quantifies the degree to which the dynamics are controlled by rotation, with smaller values indicating a stronger rotational influence. The plume morphology is found to vary with Ro^{*p} , as demonstrated by Fig. 4. When $Ro^{*p} > 0.17$, rotation has no appreciable effect on the plume evolution prior to reaching the ice shell, and the plume exhibits a typical non-rotating conical shape (panel c). A slightly higher Ro^{*p} value would suppress lateral plume expansion and constrain the plume to a cylindrical morphology (panel b). When Ro^{*p} drops below 0.018, the plume undergoes strong lateral leakage of buoyancy by baroclinic eddies and ceases to rise before reaching the ice shell, ultimately dispersing uniformly into the surrounding ocean (panel a). On Europa, a hydrothermal vent with a heat flux of 5 GW yields $Ro^{*p} \approx 0.092$, lying between 0.018 and 0.17 and suggesting that the plume is able to reach the ice shell (Goodman et al., 2004; Goodman & Lenferink, 2012). In contrast, the same vent on Enceladus yields only $Ro^{*p} \approx 0.0047$, substantially smaller than 0.018. This suggests that a buoyant plume originating from a localized seafloor vent cannot rise to the ice shell while maintaining its concentrated structure (Kang, Marshall, et al., 2022; Bire et al., 2023). The main reasons for these differences are the weaker gravity and faster rotation at Enceladus, as compared to Europa. We thus need another mechanism to generate regional topographic variations on Enceladus.

On small icy moons, when the ocean salinity is sufficiently low, the thermal expansivity α of seawater can become negative (Melosh et al., 2004). This so-called anomalous expansion causes the water to contract with warming, preventing convection from being triggered by the heating generated in the silicate core (Zeng & Jansen, 2021). As indicated by the shaded region in Fig. 1b, anomalous expansion occurs only for salinities below ~ 23 psu and pressures below ~ 20 MPa. Such conditions are expected only on small icy satellites, including Enceladus, Mimas, and the Uranian moons. In contrast, Europa's ocean experiences bottom pressures of order 100 MPa, which precludes anomalous expansion regardless of ocean salinity and ice thickness. Even at the water-ice in-

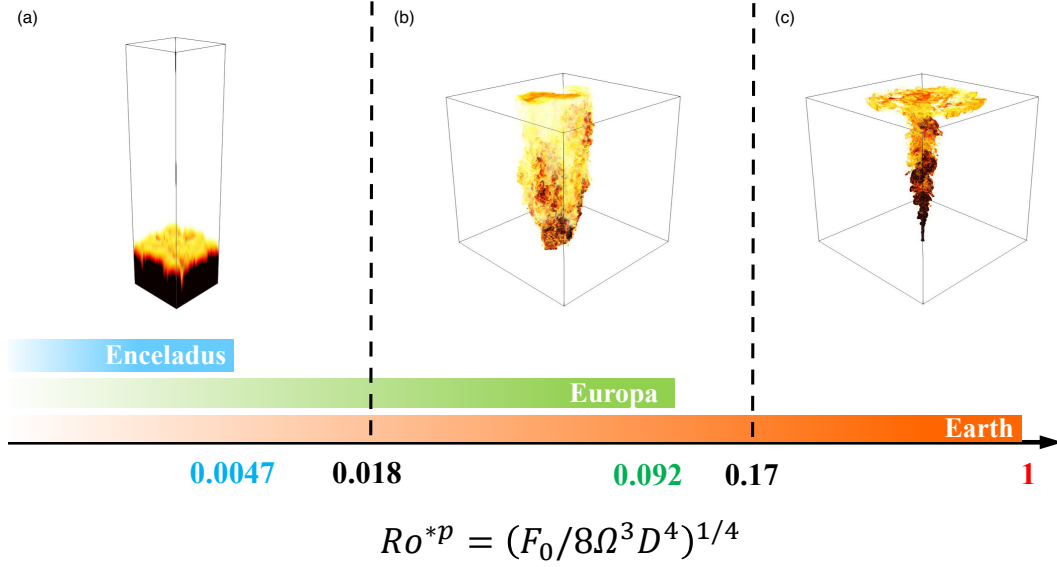


Figure 4: Simulated morphologies of a rotating point plume, adapted from Wang et al. (2025). As the natural Rossby number Ro^{*p} decreases, the plume transitions from a cone-shaped, non-rotating plume (panel c), to a rotationally constrained cylindrical plume (panel b), and finally to a state in which the plume destabilizes and scatters horizontally before reaching the upper surface (panel a). The transitions occur at $Ro^{*p} = 0.17$ and $Ro^{*p} = 0.018$. Red, green, and blue shading highlight the parameter spaces for Earth, Europa, and Enceladus, respectively. The upper bound is estimated by assuming that all hydrothermal heat on the planet is concentrated at a single point.

terface of Europa, anomalous expansion is unlikely, if its ice thickness exceeds 23 km (Melosh et al., 2004).

When convection is suppressed, heat accumulates in the bottom waters until the temperature there exceeds a critical temperature T_c , at which the maximum density is attained ($T_c \approx 4^\circ\text{C}$ for freshwater at one standard atmosphere). Although T_c decreases with increasing salinity and pressure, it is typically $\mathcal{O}(1^\circ\text{C})$ unless the anomalous expansion is nearly completely suppressed. In equilibrium, the bottom ocean, where temperatures exceed T_c , becomes convective, while the upper ocean remains strongly stratified (Vance & Goodman, 2009; Zeng & Jansen, 2021; Kang, Marshall, et al., 2022). Since convection near the ice shell is strongly suppressed, nonuniform heating from the seafloor may not project onto the ice shell as readily as it would without anomalous expansion. In cases where ice thickness variations exceed the thickness of the stratified layer, the thicker portions of the ice shell may submerge into the warm convective pool, inducing rapid melting and a reduction in ice thickness variations.

2.2 Surface and internal inertia-gravity waves due to tidal forcing

Tidal forces also act on icy satellite oceans. This drives periodic *tidal motions* that induce momentum transport, thus potentially driving NSR (Hay et al., 2024), and ocean dissipation, releasing heat and generating small-scale turbulence that can further affect ocean circulation and tracer transport (Tyler, 2011b; Chen et al., 2014; Rovira-Navarro et al., 2023; Hay & Matsuyama, 2019; Rekier et al., 2019; Zhang et al., 2024).

Barotropic and baroclinic tidal responses. All icy moons pass through the time-varying, tide-raising gravitational fields of the central planet. In response, the en-

tire icy body deforms into a triaxial figure, creating a bulge that nearly aligns with the planet. From the perspective of an observer on the surface, when the moon has an eccentric orbit, the central planet appears to librate back and forth in longitude, as well as change apparent size, due to differences between the rotation and orbital velocity of the satellite. This relative motion causes tidal forces to oscillate in longitude and magnitude. In this general case, the corresponding deformation, and associated flow in the ocean, is known as the *eccentricity tide*. The forcing and response consist mainly of zonal wavenumbers 0 and 2. If the satellite has an oblique (i.e., tilted, relative to the orbital plane) rotation axis, however, then the same observer instead sees the tide-raiser periodically librate north and south over the course of an orbit. Thus, these tide-raising forces oscillate in latitude and the corresponding additional deformation is known as the *obliquity tide*. In this case, the forcing and response are primarily limited to zonal wavenumber 1. Both eccentricity and obliquity forcing can drive a range of different tidal waves, depending on the thickness and stratified state of an ocean. Surface tidal waves are related to the barotropic (i.e. depth-averaged) response of an ocean to tidal forces, while baroclinic internal tidal waves vary through the ocean column and thus can depend on the ocean's vertical structure. In the following, we first review barotropic tidal waves, followed by baroclinic waves.

Barotropic inertial-gravity waves are driven by both obliquity and eccentricity forcing, and are the primary tidal waves associated with the tidal deformation at the surface of an icy moon (Tyler, 2008), shown in Fig. 5a. Coriolis force and gravity are the main restoring forces that act on these waves (Longuet-Higgins, 1968), although the elasticity of the overlying ice provides an additional restoring force that resists surface deformation (Beuthe, 2016; Matsuyama et al., 2018; Hay & Matsuyama, 2019; Rovira-Navarro et al., 2023). Hence, barotropic inertial-gravity waves propagate faster under an ice shell than in an ice-free ocean. The natural frequencies of these waves for typical satellites are much higher than the frequency of the dominant degree-2 tide-raising potential ($\omega \gg \Omega$), particularly for the thick oceans expected in bodies such as Europa and Ganymede. Consequently, surface inertial-gravity waves are usually far from resonance. Higher-frequency tidal forcing due to adjacent moons, although smaller than the daily tidal forcing, can potentially excite resonant, equatorially trapped Kelvin waves (Hay et al., 2020, 2022).

Obliquity forcing can additionally excite a single barotropic Rossby (or vorticity) wave of planetary scale (Longuet-Higgins, 1968; Tyler, 2008). This wave consists of a single vortex that spans from pole to pole; the centers of the vortex propagate westward, while the fastest flow occurs at the poles. However, interaction with other flow fields can potentially modify the position of the vortex; see Vincent et al. (2022), shown in Fig. 5b. The variation in Coriolis force is the restoring force acting on this planetary Rossby wave, which produces minimal surface displacement as it flows. The elastic resistance of the ice shell hence has little impact on the dynamics of this wave. For synchronously rotating satellites, these waves are close to resonance, although long-wavelength meridional variations in ocean thickness can modify the wave's natural frequency away from that of the forcing (Rovira-Navarro et al., 2020; Rovira-Navarro et al., 2023). These waves persist in thick oceans, and hence the general anticipation is that Rossby waves driven by obliquity forcing produce the strongest barotropic tidal currents in icy satellite oceans (Chen et al., 2014; Matsuyama et al., 2018; Hay & Matsuyama, 2019), unless inertial-gravity waves are also resonant. These Rossby waves driven by the tide are also the primary candidate for ocean-driven NSR due to the tidal forcing (Hay et al., 2024), the mechanism for which we discuss later. However, the obliquity of all icy satellites, with the exception of Titan (Stiles et al., 2008; Meriggiola et al., 2016), has never been measured and is only constrained through modeling (Baland et al., 2012). The predicted values of obliquity based on the assumption of a damped Cassini state are roughly 0.05° and 0.0001° for Europa and Enceladus, respectively (Chen et al., 2014).

In addition to surface-expressed tides, the ocean interior can also support tidal waves. These baroclinic tides are characterized by vertically-sheared motions that produce little signature in sea-surface height. On Earth, these motions are mostly excited when the barotropic tide flows over bathymetry, perturbing density interfaces in the ocean. This topographic conversion of barotropic to baroclinic tides can occur in icy-moon oceans (Blackman et al., 2022), for example, as the ellipsoidal ice shell librates (Rekier et al., 2019). In addition, icy moon oceans are substantially thicker than Earth’s ocean (compared to their radii), meaning that (i) non-traditional components of the Coriolis force become important, and (ii) tidal forces decrease with depth at a rate of $(r/R)^2$, driving a tidal force differential from ocean top to bottom. These effects can also drive the spontaneous generation of baroclinic waves (Rovira-Navarro et al., 2023; Idini & Nimmo, 2024).

Regardless of the excitation mechanism, oscillatory baroclinic tidal waves are supported by two restoring forces: Coriolis forces due to rotation and buoyancy forces induced by stratification. In the absence of stratification, only Coriolis forces can restore wave motion; these baroclinic waves are known as inertial waves or gyroscopic waves (LeBlond, 1977; Gerkema et al., 2008). An example of this is shown in Fig. 5c. With density stratification, gravity becomes an additional restoring force; these baroclinic waves are known as mixed gravito-inertial waves (see Fig. 5d). The dispersion relationship for gravito-inertial waves satisfies (LeBlond, 1977; Rovira-Navarro et al., 2019)

$$\omega^2 = N^2 \sin^2 \alpha + f^2 \cos^2 \lambda \quad (4)$$

where $f = 2\Omega$ is the Coriolis parameter, N denotes the Brunt-Vaisala frequency, α is the angle between gravity $\mathbf{g}$ and the wave vector $\mathbf{k}$, and λ is the angle between the rotation axis and $\mathbf{k}$. Note that, regardless of wavelengths $2\pi/|\mathbf{k}|$, all waves of a specific frequency always travel along the same direction, and this energy propagation direction is always aligned with fluid motions, both perpendicular to the wave’s phase speed (i.e., perpendicular to $\mathbf{k}$) (LeBlond, 1977). This indicates that all waves forced at the tidal frequency will coalesce into the concentrated wave beams, and the beam orientation depends on how ω compares to N and f .

Icy moons rotate synchronously such that $|\omega| = \Omega$, so in an unstratified ocean ($N = 0$), Eq. 4 indicates that tidally excited internal-inertial waves form a constant angle of 30° relative to the rotation axis. These waves propagate along conical shear surfaces within the ocean, shown in Figs. 5c–f (Rovira-Navarro et al., 2019; Aygün & Čadek, 2023). Similarly, in the absence of rotation ($f = 0$), the internal gravity waves of a given frequency propagate at a fixed angle to *local* gravity. If the stratification is particularly strong ($N^2 \gg \Omega^2$), then Eq. 4 indicates that internal gravity waves propagate vertically ($\alpha \rightarrow 0^\circ$), becoming oscillatory in depth. This situation is shown in Figs. 5g–i. For all internal gravito-inertial waves, certain configurations focus wave energy into narrow shear zones after circulating through the spherical shell, so-called wave-attractors (Rovira-Navarro et al., 2019). These shear zones, evident in the meridional cross section in Fig. 5d, are likely where most of these waves dissipate energy and are regions susceptible to instability, turbulence, and therefore vertical mixing.

All of the above tidal waves can behave resonantly when the tidal force frequency (Ω) is close to the natural frequency of the unforced wave. In such a state, the amplitude and dissipation rate of tidal waves can be significantly enhanced (Tyler, 2014; Matsuyama, 2014; Matsuyama et al., 2018; Rovira-Navarro et al., 2019; Hay et al., 2020, 2022; Aygün & Čadek, 2023; Idini & Nimmo, 2024). However, resonant configurations are limited and cover only a small fraction of the possible parameter space. Keeping a system close to resonance would require a feedback between the consequence of the resonance (enhanced dissipation, orbit alteration, modified ocean stratification, etc.) and the natural mechanism triggering it (strength of stratification, ocean depth, ice shell geometry, etc.). Such feedback, sometimes termed resonance locking or self-tuned resonance, has been proposed for tides on giant planets (Fuller et al., 2016) and icy satellites (Tyler, 2011b; Idini & Nimmo, 2024).

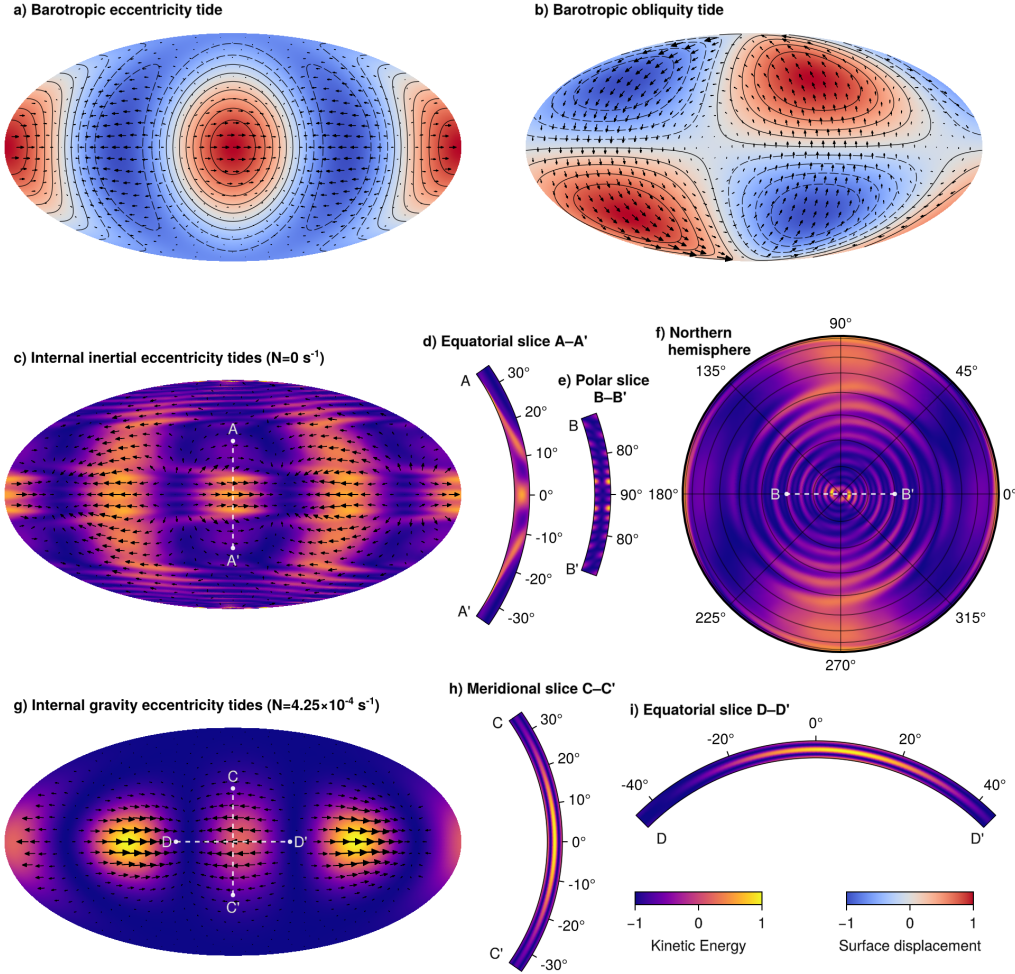


Figure 5: Snapshots of barotropic and baroclinic tidal waves from Europa-relevant simulations. The top row shows depth-averaged velocity and surface displacements for the barotropic (a) eccentricity and (b) obliquity tides. The colors in a) and b) represent the deformation of the ocean surface. The middle row shows internal tidal inertial waves in an unstratified ocean, with the tidal kinetic energy field at the near-surface c), and in equatorial d) and polar e) cross sections. The orthographic projection in f) is the northern hemisphere of panel c). The sharp features in c)–f) are internal inertial waves. The bottom row shows internal tidal gravity waves in a strongly stratified ($N = 4.5 \times 10^{-4} \text{ s}^{-1}$) ocean, with the tidal kinetic energy field at the near-surface g), and in meridional h) and equatorial i) cross sections. The internal flow due to tidally excited internal gravity waves is evident in the cross-sections in h) and i). Barotropic tides were computed using the shallow-water code ODIS (Hay & Matsuyama, 2019), while the baroclinic modes were computed using the MITgcm (Marshall et al., 1997; Abdulah et al., 2022).

Momentum transport by tides. All the tidal waves described above are driven by an oscillatory gravitational forcing and are to first order, periodic in time. However, these waves nonlinearly interact with themselves and each other. Such nonlinear interactions are small due to the low-amplitude nature of large-scale tidal waves, but they induce permanent momentum transport that sustains time-mean ocean currents. Such

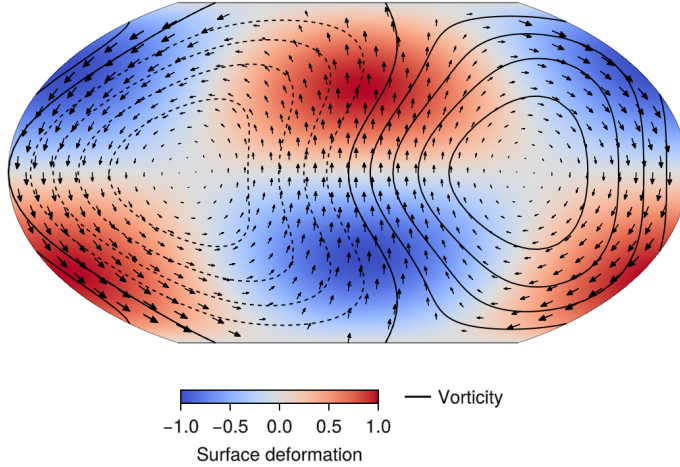


Figure 6: Mechanism for westward jet formation due to obliquity tides. Snapshot of the obliquity tide for Europa-relevant parameters, where black contours denote vorticity ($\nabla \times \mathbf{u}$), colors are ocean surface displacement, and the arrows are the barotropic (depth-averaged) tidal velocity field, similar to Fig. 5b. Tilting of the Rossby wave arises due to drag, resulting in correlations between regions of vorticity, tidal deformation, and meridional velocity. These correlations drive a north–south transport of angular momentum, and ultimately the formation of a weak westward-flowing jet. All units are normalized. Adapted from Hay et al. (2024).

currents have long been observed on Earth, typically resulting from the interaction of tidal flow with topography; they are known as rectified tidal currents (e.g., Wright & Loder, 1985). Hay et al. (2024) determined the time-mean currents driven by barotropic tidal wave–wave interactions. They found that, depending on the type of tidal waves and forcing involved, many different time-mean currents can be driven, consisting of one to multiple zonal jets, as well as permanent north–south currents. Unless the ocean was within a gravity-wave resonance, the most significant current was driven by the obliquity tide Rossby wave. Shown in Fig. 6, the Rossby-wave tilts due to drag, forming a chevron-like pattern. This tilt induces a time-average meridional transport of zonal momentum due to nonlinear wave–wave interactions, forming a single westward zonal jet centered on the equator and ultimately opposed by drag. Although the amplitude of this “jet” is small ($\leq 1 \text{ mm s}^{-1}$), it acts in the same sense everywhere. Thus, there is no meridional cancellation of zonal ice–ocean stresses, leading to net torques of $\mathcal{O}(10^{14}\text{--}10^{15}) \text{ N m}$ (Hay et al., 2024). Hence, these tidal time-mean currents exchange momentum with the ice shell and can contribute to offsets of permanent tidal bulges, as well NSR and True Polar Wander (Schenk & McKinnon, 2024). Baroclinic tides will also drive mean flows through wave–wave interactions, but this has not yet been investigated and represents a fruitful area for further study.

2.3 Baroclinic eddies arising from temperature and salinity gradients at the ice-ocean interface.

Thermal and salinity forcings from the ice shell. In addition to seafloor heating and tidal forcing, ocean circulation is also shaped by the temperature and salinity forcing from the ice shell. Because eccentricity-driven tidal dissipation within the ice shell is enhanced at high latitudes, it generally promotes a poleward-thinning ice shell, unless this tendency is offset by the poleward decrease in solar insolation (Ojakangas & Steven-

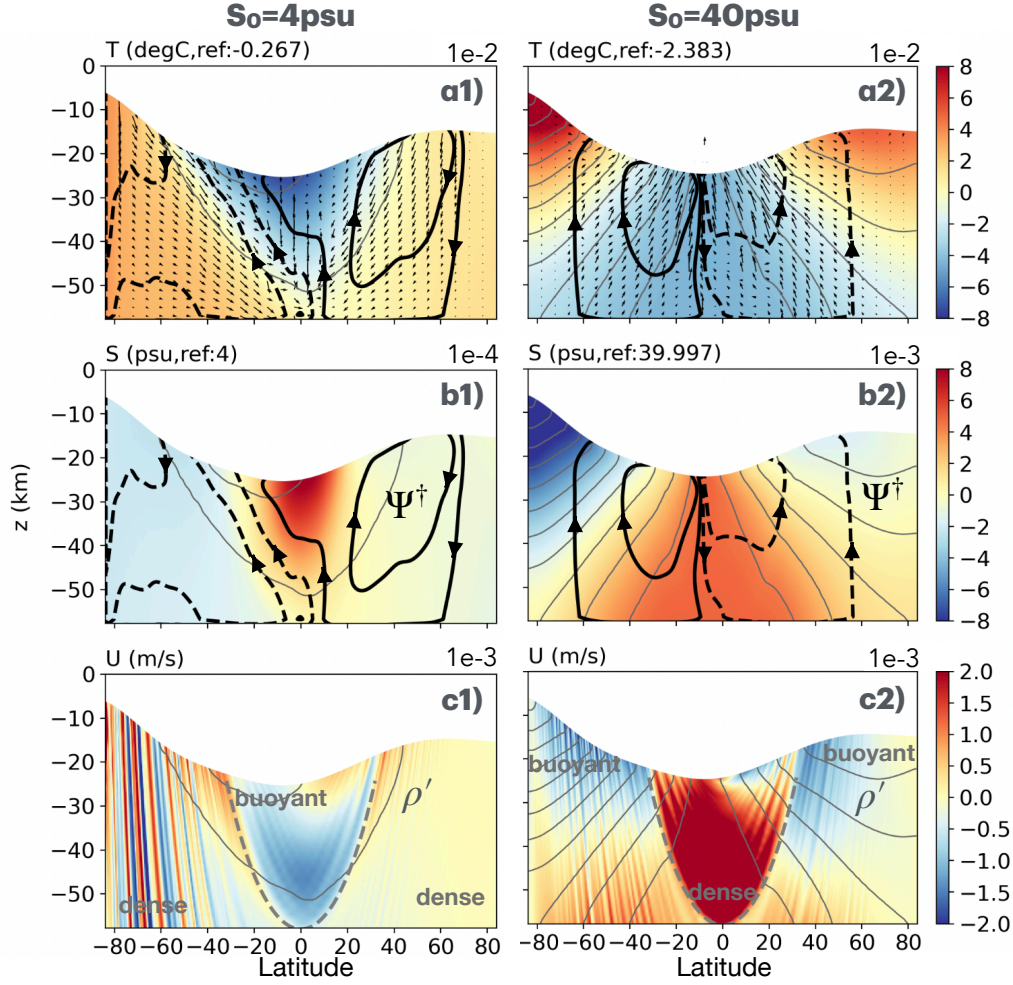


Figure 7: Ocean dynamics driven by heat and salinity forcings from the ice shell in a fresh ocean (left, 4 psu) and a salty ocean (right, 40 psu). Shadings in panels (a-c) present zonally averaged temperature T , salinity S and zonal flow speed U , respectively. To highlight the T and S variations, domain averaged values are subtracted as the reference value. Black contours in panels (a,b) present the streamfunction of the residual circulation $\Psi^\dagger$, whose direction is indicated by arrows. Arrows in panel (a) show the heat transport by baroclinic eddies. Gray contours present the isopycnals (constant density); dense and buoyant sides are marked in the plots. Tangent cylinders are denoted by a gray dashed arch in the bottom panels. Figures taken from Kang and Zhang (2026).

son, 1989b). Current observations suggest that both Enceladus and Titan possess ice shells that thin toward the poles (Hemingway & Mittal, 2019; Nimmo & Bills, 2010). Lateral variations in ice thickness can induce a temperature gradient along the water-ice interface, since the elevated pressure beneath thicker ice depresses the local freezing point, following the Clausius-Clapeyron relation. This temperature contrast, $\Delta_h T$, scales linearly with the ice thickness variation, ΔH , and the moon’s gravity, g :

$$\Delta_h T = f_b \Delta P = f_b \rho_i g \Delta H \quad (5)$$

where $f_b = -7.61 \times 10^{-8}$ K/Pa measures the sensitivity of the freezing point, T_f , on pressure, P , and $\rho_i = 920$ kg/m³ is the ice density. Currently available gravity and topography data suggest that Enceladus and Titan may have significant ΔH of 20 km and 6 km, respectively (Nimmo & Bills, 2010; Nimmo et al., 2011; Hemingway & Mittal, 2019; Schenk & McKinnon, 2024; Park et al., 2024), while Europa’s ice shell thickness variation seems to be less than 2 km (Nimmo et al., 2007). Compared with $\Delta_h T$ with $\Delta_v T$ in Table. 1, it is clear that $\Delta_h T$ can easily dominate $\Delta_v T$ by orders of magnitude, even on Enceladus, an icy moon with very weak g .

In addition to thermal forcing, a nonuniform ice shell also imposes salinity forcing on the ocean beneath. This forcing is required to sustain the ice thickness variations against down-gradient ice flow: thick ice regions need to form new ice and thin ice regions need to melt. Freezing and melting, respectively, release salty brine and fresh water into the ocean, creating a salinity gradient $\Delta_h S$ under the ice (Vance & Goodman, 2009; Lobo et al., 2021; Zhu et al., 2017; Kang, Mittal, et al., 2022; Ashkenazy & Tziperman, 2021; Ashkenazy et al., 2018). If we assume that freezing and melting of the ice are exactly balanced by ice flow (Ashkenazy et al., 2018), the salinity flux $\mathcal{H}_S$ can be estimated as

$$\mathcal{H}_S = S_0 q \propto S_0 \frac{g H_0^3 \Delta H}{\eta_i a^2} \quad (6)$$

where S_0 is the mean ocean salinity, η_i is the equivalent viscosity of the ice shell, which accounts for the temperature variations from the surface to the bottom of the ice, and q denotes the freezing/melting rate needed to balance the ice flow, which is related to the mean ice thickness H_0 , the variation of the ice thickness between the equator and the pole ΔH , gravity g , and the radius of the moon a , following the ice flow model by Nimmo (2004) and Ashkenazy et al. (2018).

Unlike temperature forcing (Eq. (5)), the equator-to-pole salinity contrast is not prescribed a priori, even under the assumption of a steady-state ice shell. If ocean circulation is efficient, the salinity is well mixed between the equator and the pole, resulting in a weak meridional salinity gradient. In contrast, in a dynamically sluggish ocean, significant salinity gradients can build up over geological timescales, leading to strong haline stratification. These conclusions rely on the assumption that the ice shell is in equilibrium. If we relax this assumption, the link between salinity flux at the ice-ocean interface and ice thickness variations breaks down. In that case, the salinity flux can only be determined from precise measurements of temporal changes in ice thickness, after subtracting the contributions from ice flow.

Stratification and baroclinic eddies. Just as the equator-to-pole temperature contrast stratifies Earth’s ocean and atmosphere, the density contrasts beneath the ice, resulting from the shift of the freezing point and local freezing/melting, stratify the subsurface oceans on icy moons. In a stratified ocean, tidal mixing injects energy into the system by transporting dense fluid upward against gravity. Bottom heating has a similar effect (see §4). By extracting the stored potential energy, baroclinic eddies grow, transporting dense fluid downward and buoyant fluid upward along surfaces of constant density (isopycnals; arrows in Fig. 7a indicate the associated heat flux), thereby lowering the ocean’s center of mass. Although strong nonlinear interactions between baroclinic eddies generate complex turbulent motions, their net effect on heat transport can be represented by a planetary-scale overturning circulation, called residual circulation by the

meteorologists (Vallis, 2006; Zhang et al., 2024). This residual circulation is shown in Fig. 7(a,b) by thick black contours.

One potential complexity of icy-moon ocean circulation, compared to the atmospheric dynamics of Earth, is that the thermal expansivity α_T can take either sign. When $\alpha_T > 0$, as expected on large icy moons or on small icy moons with salty oceans, the water expands upon warming, and the cold salty water beneath thicker ice is denser than the warm fresh water beneath thinner ice (gray contours in Fig. 7a2, b2, c2). As a result, the cold, salty water beneath the equatorial ice tends to sink and occupy the deep ocean, while the warm, fresh water near the poles remains close to the surface. In contrast, when $\alpha_T < 0$, the under-ice density gradient can reverse due to anomalous expansion (gray shading in Fig. 1b). This reversal leads to a corresponding reversal of the circulation (black contours in Fig. 7a1,a2, Zhu et al., 2017; Kang, Mittal, et al., 2022).

Heat and momentum transport Regardless of the direction of ocean circulation, heat is always transported down the gradient toward regions with thicker ice, reducing the ITV, as indicated by the arrows in Fig. 7a. The theories and scaling laws developed by meteorologists to understand heat transport and stratification in the atmosphere of the Earth (Held & Larichev, 1996; Jansen & Ferrari, 2013; Chang & Held, 2019; Gallet & Ferrari, 2020) can be adapted to the context of oceans of icy moons (Zhang et al., 2024; Kang, 2022), as will be discussed in §3. Ocean heat transport (OHT) would therefore induce ice freezing where the ice is thin and melting where the ice is thick, thus acting to reduce ice thickness variations (ITV).

In addition to OHT, the meridional density gradient in the ocean sustains a zonal shear of the mean flow along the axis of planetary rotation over mid-to-high latitudes inside the tangent cylinder (see Fig. 3). This shear arises from thermal-wind balance, a fundamental relation that holds robustly in rotation-dominated fluid systems.

$$2\vec{\Omega} \cdot \nabla u = -\frac{\partial b}{a\partial\theta}, \quad (7)$$

where u is the zonal-mean zonal flow, $b \equiv -g\rho'/\rho_0 = g(\alpha T' - \beta S')$ is the buoyancy, $\vec{\Omega}$ is the planetary rotation vector, and θ denotes latitude, α and β are the thermal expansion and haline contraction coefficient, respectively. When $\alpha > 0$, the equatorial water is denser than the polar water, leading to retrograde shear (Fig. 7c2). When $\alpha < 0$, the opposite happens (Fig. 7c1). This shear flow will exert a zonal momentum drag onto the ice shell, which can potentially cause the ice to drift relative to the solid interior (Kang, 2024).

3 Interaction between the three types of dynamics

Prior work has focused primarily on dynamics driven by individual forcing mechanisms in isolation; however, understanding how these three types of dynamics interact is equally important. Due to the nonlinear nature of their interactions, the combined dynamics is unlikely to resemble a simple superposition of the three. In this section, we highlight two interaction mechanisms: one between convection and baroclinic eddies, and the other between tides and baroclinic eddies, as sketched in Fig. 8, and discuss how the resultant OHT influences the sustainability of the ITV. The main take-away is that a sufficiently strong ITV drives heat convergence toward regions with thick ice, which in turn suppresses the ITV. This suppression is enhanced by both tidal dissipation and bottom heating, suggesting that strong ITV is difficult to sustain in the presence of strong oceanic tidal dissipation and bottom heating.

3.1 Interaction between convection and baroclinic eddies.

The first interaction mechanism that we discuss involves convection and baroclinic eddies, occurring when bottom heating (§2.1) and ice forcings (§2.3) coexist. Using both

linear stability analysis and numerical simulations, Callies and Ferrari (2018) demonstrated that baroclinic instability can develop as an independent mode even in the presence of convection. Between these two dynamics, a tug-of-war forms, where baroclinic eddies act to stratify the ocean and flatten the isopycnals, as dense fluid slides underneath buoyant fluid (see §2.3), while bottom heating works to erode stratification and steepen the isopycnals (Kang, 2023; Wang, Kang, & Li, 2026).

When the bottom heat flux Q_b is weak (Fig. 8a), baroclinic eddies dominate. They stratify the upper ocean, preventing convective plumes from reaching the ice directly above. In the stratified region, the heat from below continues to be transported upward by baroclinic eddies. But since the motions associated with baroclinic eddies tend not to cross isopycnals (constant density surfaces) (Vallis, 2006), the heat follows the isopycnals and reaches the equatorial regions, where the ice shell is thick. In contrast, when Q_b is sufficiently large (Fig. 8b), convection will take over and the stratification of the ocean will be weakened or even completely destroyed. The unstratified environment allows the basal heating pattern to project onto the ice shell.

Distinguishing between these two scenarios would provide useful insight into 1) whether the ocean is well mixed or stratified, and 2) to what extent the observed ice thickness variations can be linked to seafloor heating patterns. Kang (2023) presents a formula for the critical bottom heat flux Q_c that demarcates the convective and stratified regimes, and it is jointly determined by the magnitude of the horizontal temperature contrast in the ocean interior $\Delta_h^{(i)}T$, which broadly follows the under-ice temperature difference $\Delta_h T$ set by the ITV (Eq.5) and by planetary parameters such as ocean depth D , radius a , rotation rate f and gravity g :

$$Q_c \sim \frac{C_p \rho_0}{a^3 f^2} (\alpha_T g)^{\frac{3}{2}} (\Delta_h^{(i)} T D)^{\frac{5}{2}}. \quad (8)$$

The Q_c here determines the minimum bottom heat flux required to overcome surface stratification. If Q_c greatly exceeds the upper bound of bottom heat flux Q_b (see Table. 1), which is limited by conductive heat loss through the ice assuming thermal equilibrium, we can deduce that the upper ocean should be stratified and ITV is unlikely to be sustained by nonuniform bottom heating. Note that, if the temperature jump across the top boundary layer is not suppressed by enhanced boundary mixing, such as that driven by tidal flow over boundary roughness, $\Delta_h^{(i)}T$ will be smaller than $\Delta_h T$ set by the freezing point shift (Eq. 5). This is because the boundary layer temperature jump is larger beneath thick ice, where heat deflected from other latitudes is transmitted upward, reducing the temperature contrast beneath the boundary layer.

The estimated values of Q_c for several icy satellites of interest are listed in Table 1. Taking Enceladus as an example, with $\Delta H = 20$ km and $\alpha_T = 10^{-4} \text{ K}^{-1}$, we obtain $Q_c \sim 2 \text{ W m}^{-2}$, which is significantly greater than the heat that can be conducted through an ice shell of ~ 20 km thick. As long as Enceladus does not melt rapidly (a heat surplus of 1 W / m^2 can drive a melting rate of $\sim 100 \text{ km / Myr}$), the actual bottom heat flux should be much smaller than Q_c . Since $Q_b < Q_c$, nearly all bottom heating is diverted toward the thick equatorial ice, as illustrated in Fig. 8a. Preferential heating of the equatorial regions would progressively erase the poleward thinning gradient of the ice shell. The only viable way to sustain the strong equator-to-pole ITV observed on Enceladus is thus to generate heat within the polar ice shell itself. Furthermore, with $Q_b < Q_c$, a stratified layer will form in the upper ocean, which has important implications for the tracer transport timescale within the ocean (Ames et al., 2025; Wang, Kang, Zhang, & Marshall, 2026) and thereby what can be detected in the geyser ejecta.

A similar analysis may be conducted for Titan. If a subsurface ocean existed beneath Titan's surface (which it may not; Petricca et al., 2025) then overcoming an ice thickness variation of $\Delta H \sim 5$ km (Nimmo & Bills, 2010) would require a bottom heat flux of at least $Q_c \sim 2000 \text{ W m}^{-2}$. This value far exceeds any plausible estimate of bot-

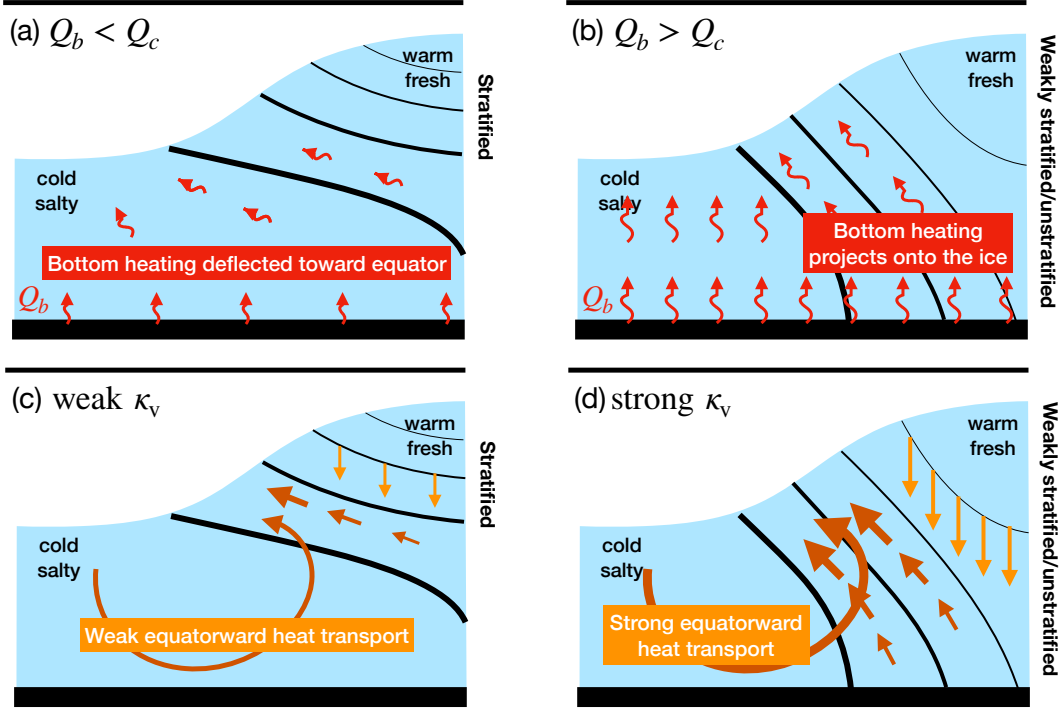


Figure 8: Panels (a,b) sketch the interaction between convection and baroclinic eddies when ice thickness varies latitudinally. Density gradients along the water-ice interface sustain a stratified layer (black contours) beneath the thin ice shell, preventing convective plumes from reaching the ice. When bottom heat flux Q_b is weaker than the critical value Q_c given by Eq. (8) (panel a), bottom heating is deflected toward the equator; when $Q_b > Q_c$ (panel b), it can reach the ice shell directly. Panels (c,d) sketch the interaction between tidal mixing and baroclinic eddies. Tidal wave breaking mixes warm fresh fluid from the surface into the deep interior (parameterized by κ_v), weakening the stratification and energizing baroclinic eddies, which then transport heat upward along constant density contours toward regions with thick ice. This heat convergence increases with the tidal dissipation rate.

tom heating, suggesting that the hypothetical Titan ocean must be stably stratified and that bottom heating cannot be communicated to the polar ice to sustain the observed ITV. Generally speaking, on larger icy satellites, the same ITV can induce a stronger shift of the freezing point under the ice, due to stronger gravity (Eq. 5), and this will deflect heat more effectively. This conclusion, however, differs from that of Kvorka and Čadež (2024), who conducted fully coupled ice-ocean simulations for Titan and found that substantial ITV can be maintained even when all heat production occurs in the silicate core. However, we suggest that their results may be consistent with our prediction given that the parameter choices made by Kvorka and Čadež (2024) imply that $Q/Q_c \sim 80\text{-}5000$ in their simulations.

The ITV on Europa is less constrained. The only constraint available based on measurements of the limb profile is consistent with a rather flat ice shell with an equator-to-pole ITV less than 2 km (Nimmo et al., 2007)). Thus, it is more useful to ask what is the minimum ITV ΔH_c needed to stratify the ocean for a given bottom heat flux Q_b (see Table. 1). For Europa, if we assume the bottom heat flux $Q_b = 40 \text{ mW/m}^2$, Eq. (8) would predict that an ITV of merely $O(100)$ m is sufficient to stratify the ocean. Since the presence of an ITV of a few hundred meters cannot be excluded, whether Europa ocean is convective or stratified remains elusive.

3.2 Interaction between tides and baroclinic eddies.

Baroclinic tides and eddies are similarly coupled. Tidal mixing (an important contributor to vertical diffusivity κ_v) steepens isopycnals and energizes baroclinic eddies and ocean circulation. In turn, baroclinic eddies drive heat transport, shaping the ocean's temperature and salinity profiles and, thus influencing the tidal dissipation rate (Tyler, 2020; Rovira-Navarro et al., 2019; Hay & Matsuyama, 2019; Requier et al., 2019; Idini & Nimmo, 2024).

Fig. 8c,d illustrates how ocean circulation and tidal mixing (parameterized by vertical diffusivity κ_v) balance each other and together determine ocean stratification. Transport by baroclinic eddies can be equivalently represented by an overturning circulation (brown swirly arrow) that pushes isopycnals upward. This motion lowers the center of mass of the ocean, releasing gravitational potential energy. In contrast, tidal mixing induces a downward buoyancy flux by diffusing the light fluid downward. This process tilts the isopycnals downward, raising the center of mass of the ocean and thereby replenishing its gravitational potential energy. At equilibrium, the upward buoyancy flow associated with baroclinic eddies ($\mathcal{F}_{b,bci}$, straight brown arrows) exactly balances the downward diffusive flux (straight orange arrows) (Lobo et al., 2021; Zhang et al., 2024, 2026; Kang & Zhang, 2026),

$$\mathcal{F}_{b,bci} = \kappa_v N^2 (2\pi a^2), \quad (9)$$

where κ_v is the vertical diffusivity induced by the tide, N is the Brunt-Väisälä frequency of the ocean, and a is the radius of the planet.

Because baroclinic eddies transport heat and salinity primarily along isopycnals (Vallis, 2006), the heat diffused downward by tidal mixing ultimately returns to the ocean-ice interface, concentrated beneath thicker ice (the equator in this example), as sketched in Fig. 8d. The magnitude of this meridional heat transport increases with the strength of the diffusive flux. Building on pioneering work by meteorologists on geostrophic turbulence (Held & Larichev, 1996; Jansen & Ferrari, 2013; Chang & Held, 2019; Gallet & Ferrari, 2020), recent studies (Zhang et al., 2024, 2026; Kang & Zhang, 2026) have investigated ocean heat transport (OHT) in icy satellites and derived scaling laws for OHT, circulation strength Ψ , and ocean stratification N^2 as functions of key parameters, such as the variation of equator-to-pole ice thickness ΔH , thermal expansivity α , planetary properties including radius a , rotation rate Ω , and gravity g , and turbulent vertical diffusivity κ_v , which is in turn expected to be controlled by tidal energy dissipation ϵ .

Assuming that the breaking of the tidal wave is the primary mechanism of vertical mixing, the vertical diffusivity κ_v can be related to the tidal dissipation rate ϵ through the Osborn relationship (Osborn, 1980; Wunsch & Ferrari, 2004):

$$\kappa_v = \frac{\Gamma \epsilon}{N^2}, \quad (10)$$

where Γ is the so-called mixing efficiency, which represents the fraction of tidal energy converted into gravitational potential energy rather than heat (i.e, $\Gamma < 1$). In the interior of Earth's oceans, $\Gamma \approx 0.2$ on average (Gregg et al., 2018), a value that is characteristic of internal wave breaking due to Kelvin-Helmholtz instability in a stratified ocean (Peltier & Caulfield, 2003). In a weakly stratified (or unstratified) ocean, however, the mixing efficiency could be much lower (or zero), as a statically stable stratification ($N^2 > 0$) is required for a positive mixing efficiency.

Combining the Osborn relationship (Eq. 10) with Eq.9 and converting the buoyancy flow $\mathcal{F}_{b,bci}$ into the heat flow $\mathcal{F}_{h,bci}$ using $\mathcal{F}_{h,bci}/\mathcal{F}_{b,bci} = C_p \rho_0 / (\alpha g)$, we can calculate the ratio between the OHT $\mathcal{F}_{h,bci}$ and the dissipated tidal energy within one hemisphere $\rho_0 \epsilon (2\pi a^2 D)$

$$\eta_{bci}^{-1} \equiv \frac{\mathcal{F}_{h,bci}}{\rho_0 \epsilon (2\pi a^2 D)} = \frac{\Gamma C_p}{\alpha g D} \sim 10^3 - 10^4 (\text{Enceladus}), 10^2 (\text{Europa}) \quad (11)$$

On a small icy moon such as Enceladus, each watt of tidal dissipation within the ocean can in principle drive several thousand watts of equatorward heat transport. Similarly to a heat pump for HVAC, this high heat-to-work ratio arises because the ocean operates as a very inefficient heat engine, which is manifested by its low efficiency of the heat engine η_{bci} , and therefore becomes an efficient heat pump (see §4). However, because strong equatorward heat convergence would act to erase the observed poleward thinning of the ice shell, the tidal dissipation within Enceladus's ocean must be negligible to maintain the large observed variations in ice thickness, assuming mixing efficiencies $\Gamma \sim O(0.1)$.

4 The energy perspective

Given the complexity of the interactions between the three types of dynamical responses, it is useful to view the system from an energy perspective (c.f. Jansen et al., 2023). To derive the energy equation, Note we start from the momentum and buoyancy (b) equations for a Boussinesq fluid:

$$\frac{D\mathbf{v}}{Dt} + 2\mathbf{\Omega} \times \mathbf{v} + \nabla p' = b\hat{z} - \nabla \times \nu \boldsymbol{\zeta} - \nabla \Phi_{\text{tide}} \quad (12)$$

$$\frac{Db}{Dt} = \nabla \cdot \kappa \nabla b \quad (13)$$

where $b = -g\rho'/\rho_0$ represents buoyancy, p' is pressure anomaly, $\boldsymbol{\zeta} = \nabla \times \mathbf{v}$ is vorticity, and Φ_{tide} is tidal potential. The boundary conditions are

$$\mathbf{v} = \mathbf{v}_b \quad (14)$$

$$\kappa \partial_n b = \mathcal{H}_B, \quad (15)$$

where $\mathbf{v}_b$ denotes the velocity of the boundaries due to tidal deformation and libration, and ∂_n denotes the derivative normal to the boundaries. $\mathcal{H}_B$ is the boundary buoyancy flux (positive when the fluid gains buoyancy).

Multiplying $\mathbf{v}$ to Eq. (12) and $(-z)$ to Eq. (13), adding them together and substituting the boundary conditions yields

$$\begin{aligned} \frac{\partial}{\partial t} \iiint \left(\frac{|\mathbf{v}|^2}{2} - bz \right) = & \underbrace{\left[\iiint \mathbf{v} \cdot \nabla \Phi_{\text{tide}} - \oint p' v_{bn} + \oint \mathbf{v}_{bt} \cdot \mathcal{H}_M \right]}_{\text{tidal work } (\dot{W}_T)} \\ & + \underbrace{\oint -\mathcal{H}_B z}_{\text{buoyancy work } (\dot{W}_B)} + \underbrace{\iiint \kappa_v \frac{\partial b}{\partial z}}_{\text{vertical diffusion work } (\dot{W}_\kappa)} + \underbrace{\iiint -\nu |\zeta|^2}_{\text{dissipation } (\dot{D})}, \end{aligned} \quad (16)$$

where v_{bn} and $\mathbf{v}_{bt}$ denotes the normal and tangential velocities set at the boundaries (Eq.14), possibly induced by ice/core shell deformation, libration or ocean current across topography, and the tangential stress along the boundaries $\nu \zeta \times \hat{n}$ is symbolically written as $\mathcal{H}_M$. The function in the volume integral on the left-hand side of the equation is the total mechanical energy (including kinetic energy $\frac{|\mathbf{v}|^2}{2}$ and potential energy $-bz$), and the terms on the right hand side of the equation are the contributions by tides, buoyancy fluxes, vertical diffusion and dissipation, respectively.

Dissipation is the only sign-definite term among all. In a fast rotating system, fluid motions tend to be invariant along the rotating axis, following the Taylor-Proudman theorem. This will give rise to turbulent boundary layers near the top and bottom boundaries, where most of the dissipation tends to occur. Following boundary layer theory (Monin & Obukhov, 1954), the energy dissipation rate in the ocean boundary layers per square meter can be estimated by

$$\rho_0 \dot{D}_{\text{boundary}} = -C_d \rho_0 |\mathbf{v}_t|^3, \quad (17)$$

where $C_d \sim 0.002-0.02$ is the drag coefficient (Lu et al., 2011) and $\mathbf{v}_t$ is the tangential velocity just outside the boundary layer.

The boundary buoyancy flux work increases the total energy of the system when buoyancy sources are located at lower elevations than buoyancy sinks, the configuration of a boiling pot. In the context of icy moon oceans, three primary sources contribute to buoyancy fluxes: (1) heat flux from the core, (2) heat flux from the ice shell, and (3) salinity flux from the ice.

When thermal expansivity is positive, the bottom heat flux drives convective motions by injecting energy into the ocean. When thermal expansivity is negative, it instead stratifies the ocean, reducing the total energy. With a bottom heat flux Q passing through an ocean with depth D , the total energy of the ocean per square meter will increase at a rate of

$$\rho_0 \dot{W}_{B,\text{conv}} \equiv \frac{\alpha_T g Q D}{C_p} \quad (18)$$

This energy injection rate is usually orders of magnitude smaller than Q . In fact, we may think of the convective process as a heat engine, where the work medium is heated at the bottom, cools down at the top, and in between drives convective motion through pressure work. The efficiency of the heat engine η can be estimated by the ratio of $\dot{W}_{B,\text{conv}}$ over Q .

$$\eta_{\text{conv}} = \frac{\alpha_T g D}{C_p} \sim 10^{-4} (\text{Enceladus}), 10^{-3} (\text{Europa}). \quad (19)$$

Following the definition, it is clear that η_{conv} increases with a deeper ocean, stronger gravity and a more compressible fluid, and it cannot exceed $\Delta T/T_{\text{max}}$ by virtue of the second law of thermodynamics. In fact, using the Maxwell relationship, one can show that $\eta_{\text{conv}} = \frac{(\frac{\partial \ln V}{\partial T})_P g D}{T (\frac{\partial S_e}{\partial T})_P} = \frac{(\frac{\partial T}{\partial P})_{S_e} \rho g D}{T} \leq \Delta_v T/T_{\text{max}}$. Here, S_e denotes entropy. $\Delta_v T$ denotes the vertical in-situ temperature difference and T_{max} is the maximum temperature at the bottom, both on the Kelvin scale.

For icy moons such as Enceladus and Europa, η_{conv} is less than 10^{-3} , indicating that even when the majority of heat is produced in the silicate core, the energy used to drive the ocean dynamics is much less than the heat flux transmitted through the ocean. Since the bottom heat flux Q_b should not exceed the total heat generation in the satellite $Q_m \sim \mathcal{O}(10)$ mW/m², the corresponding energy injection rate is merely $\mathcal{O}(10^{-6})$ W/m². Summarized in Table 1 is the maximum flow speed U_Q and the corresponding Rossby number Ro_Q for various icy satellites, assuming that they are forced only by convection ($Q_b = Q_m$) and dissipated only in the boundary layer (Eq.17):

$$C_d \rho_0 U_Q^3 = \frac{\alpha_T g Q_m D}{C_p}. \quad (20)$$

For the three icy satellites considered here, the current speed sustained by bottom heating U_Q does not exceed a few cm/s. If part of the tidal heating occurs in the ice shell and/or if additional kinetic energy dissipation occurs in the ocean interior, U_Q would further weaken. More detailed discussion on why this ocean current upper bound is significantly lower than what is reported by numerical simulations (e.g., Soderlund et al., 2014) can be found in Jansen et al. (2023). The magnitude of ocean current speed has important implication for whether certain particles can be suspended and redistributed by ocean current (Schoenfeld et al., 2023).

The upper bound on ocean current speed estimated here is two orders of magnitude lower than the $\mathcal{O}(1)$ m/s upper bound suggested by Cabanes et al. (2024). This discrepancy arises from the different assumptions made about the boundary layer. In Cabanes et al. (2024), a laminar Ekman boundary layer is assumed, so viscosity provides only limited dissipation and allows stronger interior flow to develop. However, if the flow speed indeed reaches $\mathcal{O}(1)$ m/s, the Reynolds number of the boundary layer exceeds 10^5 , indicating that the boundary layer would in fact be highly turbulent. In a turbulent boundary layer, much of the dissipation arises from the conversion of mean kinetic energy into turbulent kinetic energy through shear instabilities, followed by a cascade to smaller scales, where it is ultimately dissipated by molecular viscosity. This turbulence-induced dissipation can be parameterized as $C_d |U|^3$ based on the wall-bounded turbulence theory (Monin & Obukhov, 1954). Accounting for the enhanced boundary-layer dissipation therefore leads to a lower estimate of the interior flow speed.

In addition to buoyancy forcing from the seafloor, when heat and salt exchanges with the ice shell correlate with the topography of the water-ice interface, they can change the ocean's total energy. The salinity flux from the ice is proportional to the freezing/melting rate, denoted q , and the mean salinity of the ocean S_0 . Assuming the thickness of the ice is near equilibrium, q can be estimated from the thickness variations driven by ice flow from the thick to thin regions of the shell (Kang, Mittal, et al., 2022). As noted by Jansen et al. (2023), under this condition, high-salinity water should form beneath thick ice and low-salinity water beneath thin ice, always reducing the total energy of the ocean. However, the heat flux from the ice shell can increase or decrease the kinetic energy of the ocean depending on the sign of thermal expansivity. When the thermal expansivity is negative (positive), the heat flux from the ice shell increases (decreases) the energy of the ocean by cooling the water under thick ice and warming the water under thin ice. Although buoyancy flux from the ice may not directly energize the ocean, the existence of horizontal buoyancy gradients can still strongly modulate the ocean dynamics, determining how heat and momentum are transported (see §3).

Diffusion can consume or inject energy into the ocean depending on whether the ocean is convective or stratified. When the ocean is stably stratified, diffusion can increase the potential energy of the ocean by mixing dense fluid upward. As explained in §2.3 and §3, stratification can easily form in icy moon oceans in the presence of modest variations in ice thickness, allowing diffusion to potentially energize the ocean.

Tidal forcing contributes to the energy budget of the ocean in three ways: 1) the interior tidal forcing $\iiint \mathbf{v} \cdot \nabla \Phi_{\text{tide}}$, 2) the pressure work done by the ice shell and silicate core, $-\iint p' v_{bn}$ and 3) the frictional drag at boundaries $\iint \mathbf{v}_{bt} \cdot \mathcal{H}_M$, as they librate relative to the ocean in the presence of under-ice or seafloor topography, and/or as the two interfaces deform. If the ocean is stratified, like the Earth's ocean, some of the tidal energy will likely be used to generate internal gravity waves (IGWs), and these IGWs subsequently break down as they approach critical levels (Booker & Bretherton, 1967) or near rough boundaries or critical slopes (Dauxois & Young, 1999). As a result, a portion ($\Gamma \sim 20\%$ on Earth) of the wave energy will be used to vertically mix a stratified ocean, and the remaining will be used to produce heat (Wunsch & Ferrari, 2004). The vertical mixing then steepens the isopycnals, providing energy for baroclinic eddies (see §2.3). If the ocean is unstratified, inertial waves can be generated by responding to the tidal potential and enhanced through the elliptical instability (Favier et al., 2015; Le Bars et al., 2015), leading to ocean dissipation without generating potential energy.

Unlike convection, where only a small fraction of the heat flux is converted into kinetic energy ($\eta \sim 10^{-4}$ for Enceladus), tidal forcing directly drives a mechanical response, and the energy dissipated is comparable to the total tidal work input. For fixed damping, heat generated internally by tidal dissipation therefore sustains a much more energetic ocean circulation than an equivalent bottom heat flux. In Table. 1, we present the upper bound on the speed of ocean currents $U_T = (Q_m/\rho_0/C_d)^{1/3}$ ($\Leftarrow C_d \rho_0 U_T^3 = Q_m$). By adopting this formula, we again assume that 1) all heating is produced in the ocean through tidal processes, and 2) all dissipation takes place in the boundary layer through friction. Compared to the speed of the ocean current for the core-heating scenario U_Q , U_T is indeed much stronger.

General circulation models typically cannot capture tidal dissipation, and may not include tidal forcing at all when the focus is on thermally driven dynamics such as convection and baroclinic eddies. In such simulations, the effect of tidally driven turbulence on the large-scale circulation is represented through an additional turbulent diffusivity, which drives diffusive work $\dot{W}_\kappa$. One must choose this turbulent diffusivity to be consistent with the expected ocean tidal energy dissipation rate in icy moons. Unfortunately, this dissipation rate remains poorly constrained (c.f. §2.2 and §3.2).

5 Ocean heat transport (OHT) and ice thickness variations (ITV)

As the only liquid layer in icy satellites, the subsurface ocean can redistribute heat in ways that the ice shell and silicate core cannot, driving distinct patterns of freezing and melting that leave detectable signatures in the ice-thickness variations (ITV). Here, using Enceladus as an example, we discuss what the observed ITV can tell us about the interior dynamics.

Enceladus's ice shell seems to present a strong poleward thinning trend, as inferred from gravity and topography measurements (Hemingway & Mittal, 2019; Schenk & McKinnon, 2024; Park et al., 2024) and direct measurements of infrared emission from the polar regions (Howett et al., 2011; Miles et al., 2025). This pattern is consistent with the poleward-amplified tidal dissipation expected in the ice shell (Ojakangas & Stevenson, 1989b). Yet many theoretical tidal calculations at Enceladus's current eccentricity have struggled to produce enough dissipation in the ice to balance the observed heat loss (Robuchon et al., 2010; Shoji et al., 2013; Běhouňková et al., 2013; McCarthy & Cooper, 2016; Beuthe, 2019; Soucek et al., 2019; Gevorgyan et al., 2020). One possible resolution is that Enceladus had a higher eccentricity in the past, driven by long-term orbital evolution coupled to internal structural changes from ice freezing and melting (P. Goldreich et al., 2025), though direct evidence for such a history remains elusive.

This shortfall has motivated the consideration of additional heat sources in the silicate core and the ocean (Beuthe, 2019; Soucek et al., 2019; Gevorgyan et al., 2020). The partitioning of heat production among the ice shell, ocean, and silicate core is critical to understanding the potential for hydrothermal activity, the habitability of the ocean, and whether the ocean is convective or stratified. Most theoretical and numerical studies of ocean tides suggest that tidal dissipation within the Enceladus ocean contributes only a secondary fraction of the total heat budget (Chen et al., 2014; Hay & Matsuyama, 2019; Requier et al., 2019; Rovira-Navarro et al., 2019), except when turbulent dissipation induced by libration is considered (Wilson & Kerswell, 2018). Tidal dissipation in the silicate core (Choblet et al., 2017) has the potential to sustain the current rate of heat loss, but is subject to substantial uncertainties arising from poorly constrained core rheology and complex dissipation mechanisms within such a porous-viscoelastic material (Liao et al., 2020; Rovira-Navarro et al., 2022; Kamata, 2023; Hay et al., 2025).

Although directly measuring the heat production rate in each layer is challenging, it may be inferred from precise measurements of tidally driven ice-shell deformation, together with gravity and topography anomalies, which also reflect deformation of the ocean and core. Combining these observations could allow the contributions of each layer to be estimated (Bagheri et al., 2025). Until data of sufficient quality become available, however, the observed geometry of the ice shell provides an alternative constraint. To sustain a poleward-thinning ice shell, either the polar ice must generate more heat internally, or the ocean must transport more heat to the polar regions than to the equatorial regions. Although heat production in the core may also be amplified poleward (Choblet et al., 2017), enhanced heat generation at high latitudes in the core does not necessarily imply that this excess heat would be delivered to the polar ice shell.

As discussed in §2.3 and §3.1, a poleward-thinning ice shell induces a temperature gradient along the water-ice interface that tends to stratify the ocean beneath the polar regions, preventing heat produced in the silicate core from reaching the polar ice (see Fig.8a). To completely overcome this barrier, the average bottom heating must exceed a critical threshold $5Q_c$ (Eq.8), which is roughly 10 W/m^2 for Enceladus (Kang, 2023).

Hydrothermal vents can locally concentrate large amounts of heat, potentially exceeding this threshold. However, due to weak gravity and rapid rotation on Enceladus, concentrated plumes become unstable within a few kilometers above the seafloor, cease rising and spread laterally at heights $h_p \lesssim 10 \text{ km}$ (Zhang et al., 2025; Wang et al., 2025). In addition, zonal jets can effectively homogenize heat along a given latitude (Lemasquerier et al., 2023). As a result, even if the peak bottom heat flux initially exceeds the 10 W/m^2 threshold, by the time the plume reaches $\sim 10 \text{ km}$ altitude, the heat flux per unit area will have diluted to approach the average heat flux over the polar regions, which should be much lower than 10 W/m^2 .

If thermal expansivity $\alpha_T < 0$, convection beneath the ice is completely suppressed, and bottom heat is again preferentially delivered to the equatorial ice, which extends deeper into the warm ocean interior (Zeng & Jansen, 2024; Kang, Bire, & Marshall, 2022). Therefore, regardless of the sign of the thermal expansivity, a poleward-thinning ice geometry intrinsically inhibits the delivery of heat generated in the deep ocean or silicate core to the polar ice. Thus, maintaining such an ice geometry requires heat to be generated within the polar ice shell itself.

To be more quantitative, we need to consider the heat budget of the ice:

$$\mathcal{H}_{\text{cond}} = \mathcal{H}_{\text{ocn}} + \mathcal{H}_{\text{ice}} + \mathcal{H}_{\text{lat}}. \quad (21)$$

From left to right are conductive heat loss through the ice, heat flux from the ocean, heat produced in the ice shell, and latent heat release associated with freezing/melting, which in turn is set to balance the ice flow. Writing the budget separately for the equatorial and polar regions and then taking their ratio leads to the following constraint on the core-

ice heat partition.

$$\frac{\overline{\mathcal{H}_{\text{ice}}}}{\overline{Q_b}} \geq \frac{H_e}{H_p}, \quad (22)$$

where $\overline{Q_b}$ and $\overline{\mathcal{H}_{\text{ice}}}$ denote the average heat flux generated in the core and ice, respectively. As elaborated in Appendix B, the derivation takes advantage of the results that the ITV on Enceladus is strong enough to completely stratify the polar ocean and deflect all the heat generated in the core Q_b towards the equatorial ice.

For Enceladus, $H_e/H_p \sim 3$ (Hemingway & Mittal, 2019; Schenk & McKinnon, 2024; Iess et al., 2014; Park et al., 2024); then to sustain such ice thickness contrast requires at least 75% of the total heat production to be generated in ice shell (Fig.9a, Kang, 2023).

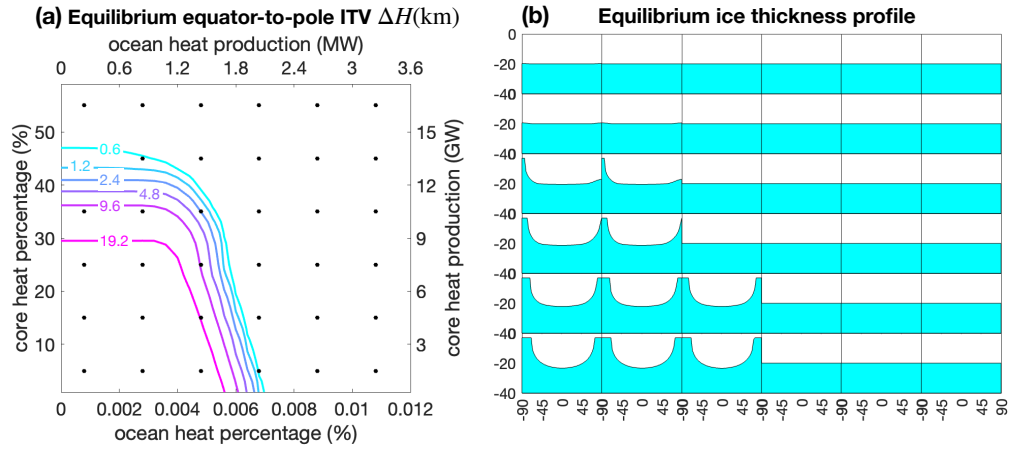


Figure 9: (a) Equilibrium equator-to-pole ice thickness difference ΔH plotted as a function of the proportions of energy dissipated in the ocean and in the core of Enceladus. (b) presents a subset of equilibrium ice thickness profiles, which are indicated by black dots in panel (a). The white and blue colors in each panel indicate ice and ocean, respectively.

We can also compute the equilibrium ITV ΔH under different heat-partition scenarios and identify which combinations are consistent with observations. Following Kang and Jansen (2022), we integrate an ice evolution model under Enceladus configurations to obtain ΔH as a function of the fractions of energy dissipated in the ocean and in the core, as shown in Fig.9a, with a representative subset of the resulting equilibrium ice-thickness profiles shown in Fig. 9b (marked by dots in panel a). Details of the calculation are provided in §Appendix C.

Consistent with the rough estimate of Eq. 22, the fraction of total heat production occurring in the core must be $\leq 35\%$ to maintain an ITV comparable to that currently inferred for Enceladus. The upper bound on the fraction of ocean dissipation is far more stringent, at $\leq 0.007\%$, which is unsurprising given that even weak tidal mixing in the ocean can drive vigorous heat transport by baroclinic eddies (§3.2). In the transitional zone between strong and weak ITV, hemispherically asymmetric ice-thickness profiles emerge (Fig.9b), consistent with the findings of Kang and Flierl (2020).

In scenarios that reproduce the observed ITV, global heat production in the ice shell ranges from 15 to 27 GW (average heat flux of 19 to 34 mW/m²). While this is consistent with some models of heating in equilibrium ice shells (Kamata & Nimmo, 2017), it exceeds the maximum values found in others (Beuthe, 2019; Soucek et al., 2019; Gevor-

909 gyan et al., 2020), so whether a fully self-consistent picture can be constructed remains
 910 an open question. The upper bound on ocean heating, of order 1 MW, is meanwhile con-
 911 sistent with existing estimates of ocean tidal heating (Chen et al., 2014), which are of
 912 order 10 kW.

913 Besides heat partitioning, strong ITV can also constrain the equation of state of
 914 seawater and, by extension, the salinity of the ocean. For small icy satellites like Ence-
 915 ladus, the thermal expansivity of seawater can become negative at low mean salinity S_0
 916 (Fig. 1b), reversing the equator-to-pole density contrast and the direction of circulation
 917 (Fig. 7, column 1). Such low-salinity scenarios, however, are unlikely to be compatible
 918 with strong ITV (Kang, Mittal, et al., 2022; Zeng & Jansen, 2024; Kang, Bire, & Mar-
 919 shall, 2022), because unlike salty oceans, thermal forcing from the ice can directly en-
 920 ergize the ocean circulation (§4), driving strong OHT that would effectively erase the
 921 equator-to-pole ITV. Since salinity and heat partitioning have degenerate effects on ITV,
 922 the two must be constrained together (Kang & Zhang, 2026). Independent constraints
 923 on ocean salinity would be helpful in resolving this degeneracy.

924 Beyond Enceladus, each icy satellite has its own unique characteristics that war-
 925 rant a case-by-case investigation. Nevertheless, some general conclusions can be drawn,
 926 particularly when strong ITV is observed. Strong ITV is difficult to maintain unless most
 927 of the heat is produced within the ice shell itself. Heat generated in the silicate core tends
 928 to be diverted toward regions with thicker ice by ocean dynamics, as long as the aver-
 929 age bottom heat flux Q is smaller than Q_c (Eq. 8). Ice flow transports ice from thick to
 930 thin regions, and conductive heat loss is more efficient through thinner ice. Together,
 931 these processes all act to reduce ITV, suggesting that a strongly varying ice shell gen-
 932 erally requires ice-internal heating, or strong variations in surface temperature or poros-
 933 ity, to dominate the heat budget, particularly for the Galilean moons.

934 It is also worth noting that any larger icy satellites that exhibit significant ITV will
 935 differ from smaller ones in two key respects: (1) anomalous expansion is completely sup-
 936 pressed by high pressure, (2) for a given ITV, the under-ice temperature contrast is stronger
 937 because of the larger gravity (Eq. 5) and (3) the bottom of the ocean may be covered
 938 by high-pressure ice (Vance et al., 2018). (1) and (2) enhance ocean stratification and
 939 increase the critical heat flux Q_c (Eq. 8), which means a bigger barrier for bottom heat
 940 to reach the thinner part of the ice. In the meanwhile, the stronger gravity on larger icy
 941 moons, however, may allow individual, concentrated convective plumes to overshoot the
 942 relatively thin stratified layer beneath that thinner ice (Wang, Kang, Zhang, & Marshall,
 943 2026).

944 The presence of a high-pressure ice layer beneath the ocean introduces additional
 945 complexity to the system. An uneven interface between this layer and the ocean can drive
 946 circulation analogous to that driven by the upper low-pressure ice shell. Sustaining ice-
 947 thickness variations in the high-pressure ice layer similarly requires inhomogeneous heat
 948 sources within, to compensate for processes such as OHT and ice flow that tend to re-
 949 duce these variations. The two ice layers bounding the ocean can also interact with each
 950 other by modifying the temperature along their respective interfaces with the ocean, thereby
 951 affecting the heat delivered to the opposite ice layer. These topics have not yet been sys-
 952 tematically investigated and remain open.

953 **6 Ocean momentum transport and non-synchronous rotation (NSR)** 954 **of the ice shell**

955 Almost all regular icy moons rotate synchronously, to first order, meaning that one
 956 side permanently faces the tide-raising central planet. Long-time-scale deviations from
 957 this state have long been hypothesized, as torques on a tidally deformed satellite in an
 958 eccentric orbit act to accelerate its spin rate (P. Goldreich & Soter, 1966). A state of equi-

librium, where the spin-rate is constant in time, may be reached with the ice shell rotating slightly faster than synchronous, slowly reorienting the surface over time (Greenberg & Weidenschilling, 1984). Various lines of geological evidence support the existence of slow ice-shell reorientation of Ganymede and Europa, including the distributions of tectonic features (e.g., McEwen, 1986; Geissler et al., 1998; Rhoden & Hurford, 2013; Kattenhorn & Hurford, 2009) and impact craters (Zahnle et al., 2003). Further theoretical work has shown that additional torques are caused by out-of-equilibrium variations in ice shell structure due to tidal heating (Ojakangas & Stevenson, 1989b), while gravitational torques acting on mass asymmetries (Greenberg & Weidenschilling, 1984; Bills et al., 2009) and elastic restoring torques due to shearing of the ice shell (P. M. Goldreich & Mitchell, 2010) act to contour non-synchronous rotation. Most recently, Burnett and Hayne (2023) have found that NSR may only be possible for a narrow range of rheological and interior structures, especially when more realistic viscoelastic rheologies are considered (Makarov & Efroimsky, 2013). However, these studies neglect the viscous torque exerted at the ice-ocean interface as the ocean flows beneath it (Hay et al., 2023; Ashkenazy et al., 2023; Hay et al., 2024; Kang, 2024). In this section, we recap how the ocean dynamics outlined in §2 can lead to a net torque on the ice shell, transporting angular momentum from the silicate core to the ice shell, and hence reorientation of the ice over time. We use estimates of this torque in different dynamical regimes to place bounds on the kinetic and dissipated energy in the upper ocean.

As mentioned in §2, there are three mechanisms to transport momentum between the silicate core and the ice shell. First, convection can trigger thermal Rossby waves near the equator (outside the tangent cylinder), transporting momentum from the core to the shell (Busse & Or, 1986; J. Aurnou et al., 2007; Soderlund, 2019; Bire et al., 2022), when planetary rotation dominates ocean turbulence ($Ro \ll 1$), which is generally true for icy moon oceans (see Table. 1). Similarly, tidal waves due to obliquity can induce a single weak retrograde jet in the equatorial region, removing angular momentum from the ice shell (Hay et al., 2024), as discussed in §2.2. Finally, a meridional density gradient in the ocean induced by ice thickness variations can sustain flow shear along the axis of planetary rotation, following the thermal wind balance (Eq. 7). If the ice shell is poleward-thinning and the thermal expansivity is positive, the upper ocean would usually flow in a retrograde direction with respect to the bottom of the ocean, driving a retrograde NSR of the ice shell. The shear and resultant NSR could reverse if one of the following two conditions is met: (1) the ice shell thickens poleward, or (2) the thermal expansivity of water is negative (Kang, 2024). The latter condition can occur in low-salinity, low-pressure environments, such as those found on small icy moons orbiting Saturn and Uranus.

Type of forcing	NSR direction	Reference
convection	prograde	(Ashkenazy et al., 2023; Hay et al., 2023)
tides	retrograde	(Hay et al., 2024)
ice topography	retrograde if thermal expansivity $\alpha_T > 0$; prograde if $\alpha_T < 0$, assuming ice shell is poleward thinning. Opposite is true if ice shell thickens poleward	(Kang, 2024)

Table 2: Ocean-induced NSR direction driven by different types of forcing.

Although all three types of forcing can drive momentum transport in the ocean and thereby induce NSR, the resulting NSR direction can differ among mechanisms (Table 2), allowing us to distinguish end-member scenarios. Assuming that the ice topography thins toward the poles and that the thermal expansivity of seawater is positive (generally holds on large icy moons), a prograde NSR due to ocean dynamics would point to convection-dominated dynamics. In contrast, a retrograde NSR could be evidence that tides and/or baroclinic dynamics (driven by the ice thickness gradient) outweigh convection, consistent with weak core heating and a stratified ocean. Interestingly, the latest reconstruction of Enceladus’s ice-shell thickness profile identifies a possible westward drift of the tidal bulge relative to the sub-Saturn point (Schenk & McKinnon, 2024), which may be explained by oceanic torque driven by tides or baroclinic eddies. Since the ITV inference discussed in §5 is only possible when ITV can be detected, which requires the majority of heat to be produced within the ice, the direction of NSR can help break the degeneracy in ITV inference.

As mentioned at the beginning of this section, beyond ocean torque, other processes, such as gravitational coupling with neighboring satellites and shell drift driven by tidal torque (Greenberg & Weidenschilling, 1984; Ojakangas & Stevenson, 1989a; Hoppa et al., 1999), can also generate NSR in the prograde direction. Furthermore, the work of deformation resulting from NSR provides a counter-torque (P. M. Goldreich & Mitchell, 2010). These contributions of gravitational torque need to be taken into account and removed before attributed to NSR to ocean torque (Ashkenazy et al., 2023), particularly if the observed NSR is prograde.

If the rate of ocean-driven NSR can be determined, it can be used to put constraints on the energetics of the ocean, as the torque exerted by the ocean necessarily consumes the energy of ocean currents, which has to be replenished by convection or tides if an equilibrium state is to be maintained.

To infer ocean’s energetics based on the NSR rate, the first step is to estimate the component of the ocean torque aligned with the rotation axis, Γ_s , from the observed NSR rate V_d (m/s), assuming that all other contributions to NSR can be taken into account (Greenberg & Weidenschilling, 1984; Ojakangas & Stevenson, 1989b; P. M. Goldreich & Mitchell, 2010; Jara-Orué & Vermeersen, 2011). Following the analysis by Ashkenazy et al. (2023), V_d should be proportional to the driving torque Γ_s and the viscous relaxation rate of the ice shell. If the ice shell adapts to its new orientation faster due to its low viscosity η_i and/or its low moment of inertia I , the gravitational torque that resists further NSR will be reduced, allowing the ice shell to undergo faster NSR. The relation between V_d and Γ_s is given by

$$\Gamma_s \equiv \int_{-\pi/2}^{\pi/2} \tau_s a \cos \phi (2\pi a^2 \cos \phi) d\phi = \frac{Ik_i \tau_M V_d}{a}, \quad (23)$$

where τ_s denotes the zonal-mean zonal stress exerted by the ocean current, k_i characterizes the elastic strength of the ice shell which is defined in (Ashkenazy et al., 2023) and $\tau_M = \eta_i/\mu$ is the Maxwell timescale of ice, η_i denotes the depth-averaged viscosity of the ice shell, $\mu = 3 \times 10^9$ Pa is the elastic shear modulus of ice. Note that Eq.23 is the same as Eq. 7 in (Ashkenazy et al., 2023), except the r_o factor in the denominator is dropped. By so doing, we assume the drift speed of ice is negligible compared to the ocean current. In this step, the largest uncertainty comes from the viscosity of the ice η_i . Observational constraints on the imaginary component of the tidal Love number may provide useful constraints on η_i (Roberts et al., 2023a).

The dissipation along the water–ice interface can be estimated by the product of torque and ocean flow speed U_s

$$\dot{E} = \int_{-\pi/2}^{\pi/2} \tau_s U_s (2\pi a^2 \cos \phi) d\phi. \quad (24)$$

If the water-ice interface is overall flat, the stress there follows the quadratic law

$$\tau_s = C_d U_s^2 \text{sign}(U_s), \quad (25)$$

where C_d is a dimensionless drag coefficient that varies with the roughness of the interface. On Earth, C_d under sea ice is generally between 2×10^{-3} and 2×10^{-2} (Lu et al., 2011). When the interface has zonal topography, an additional wave drag associated with the excitation of lee waves must also be accounted for (Legg, 2021).

Substituting Eq. (25) into Eq. (24), we get

$$\dot{E} = \int_{-\pi/2}^{\pi/2} |\tau_s|^{3/2} C_d^{-1/2} (2\pi a^2 \cos \phi) d\phi \geq C_d^{-1/2} \left| \int_{-\pi/2}^{\pi/2} \tau_s (2\pi a^2 \cos \phi) d\phi \right|^{3/2} \geq C_d^{-1/2} (\Gamma_s/a)^{3/2}. \quad (26)$$

The first inequality holds as an equality if τ_s is a constant across the globe, and the second inequality reduces to an equality if all stress concentrates at the equator. These two conditions cannot be satisfied simultaneously, and therefore $\dot{E}$ is indeed strictly smaller than the RHS of Eq. (26)). If Γ_s can be constrained by the observed NSR rate V_d (Eq. (23)), Eq. (26) allows us to put a lower bound on the ocean energy dissipation rate, as it has to be balanced by energy injection by either bottom heating or tides as discussed in §4. Since the bottom heating is bounded by the heat loss through the ice sheet, a large ocean-driven NSR hence would provide a lower bound for the ocean tidal energy dissipation since it neglects dissipation within the ocean.

7 Potential for future observational constraints

Future spacecraft measurements can provide additional constraints on the properties of subsurface oceans. Five principal techniques have proven useful: static gravity and shape, magnetic induction, tidal response, rotation state, and seismology. Here, we discuss what each technique can constrain, at what accuracy, and how they may complement one another.

Gravity and surface topography have been used to constrain the mean thickness of the ice/ocean layers and the ice thickness variations (ITV) (Nimmo et al., 2007; Iess et al., 2014; Hemingway & Mittal, 2019; Nimmo & Bills, 2010). At long wavelengths, ice shells are expected to be isostatically compensated, in which case the gravity signals are expected to be small and can be affected by core deformation. On the other hand, for such shells the topography provides a direct measurement of shell thickness variations. With sufficiently good measurements the gravity due to surface topography can be subtracted (Bouguer correction), with any residual gravity anomalies likely indicating shell thickness variations, as at the Moon (Wieczorek et al., 2013) and at Enceladus (Hemingway & Mittal, 2019). To reduce artifacts associated with spherical harmonic representations, high-order harmonics are required to reconstruct credible ITV. In the coming decade, the Europa Clipper mission will combine gravity measurements with ice-penetrating radar observations from REASON to constrain Europa’s mean ice thickness and lateral ITV. Current mission design studies suggest that the mean ice thickness may be constrained to about ± 20 -50% (Roberts et al., 2023b), unless the radar is able to detect clear reflections from the ice-ocean interface. The Jupiter Icy Moons Explorer (JUICE) mission is expected to recover Ganymede’s gravity field to roughly spherical harmonic degree 35-40 and measure global topography at meter-level precision with the Ganymede Laser Altimeter (GALA), so observational errors in gravity-topography analyses should be small. The source of uncertainty in inferred ice-shell thickness variations comes from poorly known ice-shell and ocean densities, such that a plausible $\sim 5\%$ density uncertainty would translate to roughly 5% uncertainty in the inferred thickness variations.

Magnetic induction has enabled the detection of subsurface oceans on several Galilean moons, as induced currents are generated when a satellite moves through a spa-

tially varying magnetic field (K. K. Khurana et al., 2009; K. Khurana et al., 1998; Kivelson et al., 1999; Weiss et al., 2021). Future missions using multi-frequency techniques will be able to determine both the thickness of the ocean and its mean conductivity (a proxy for salinity) (Seufert et al., 2011, e.g.). A conductor in motion (such as an ocean current) will also generate an induction signal even in a steady background field. On Earth, this idea has been used to measure oceanic flow speeds (Tyler et al., 2003) and in principle the same approach could be used on icy moons (Tyler, 2011a). In practice, however, this technique is likely to be very difficult, even with an orbiter, because the expected signals are so small (Šachl et al., 2025).

Direct measurements of tidal responses through gravity anomalies and strain rates provides constraints on both the ocean dissipation and the ice shell properties (Wahr et al., 2006). For most cases, the gravitational and deformation signals of these oscillations are likely small. However, it has been proposed that a stratified ocean can exhibit an anomalously large (or even negative) tidal response (Idini & Nimmo, 2024, e.g.) if it happens to be in a resonant configuration (see §2.2); such a response would be readily measurable by gravity, but not in surface deformation. Also, the higher-frequency deformation induced by tides raised by one moon on another becomes dynamically enhanced relative to the daily tide (Hay et al., 2020, 2022). The ESA’s JUICE mission will attempt to measure such signals (De Marchi et al., 2022; Van Hoolst et al., 2024). In particular, constraints on the complex Love numbers (both real and imaginary components) would allow the tidal dissipation rate in the outer ice shell to be distinguished from that in the underlying ocean or core layers (Bagheri et al., 2025). These constraints can then be compared with those inferred from ITV and ocean heat transport (§5). On the other hand, maps of regional-to-local tidally driven strain may be obtained through surface deformation measurements. Such maps would provide direct constraints on ITV, as regions with thinner ice would deform differently from regions with thicker ice (Berne et al., 2023).

The rotational state of the ice shell, including NSR and true polar wander, can in principle be estimated either by comparing images or limb profiles acquired during a single long mission or between two separate missions (Hoppa et al., 1999). The latter approach requires careful calibration and may involve substantial uncertainties. A key challenge of this method is separating NSR from long-period librations. Alternatively, NSR can be estimated by identifying the offset between the tidal bulge and the tidal axis (Schenk & McKinnon, 2024). This approach primarily constrains the direction of NSR but has limited sensitivity to the NSR rate. As shown in §6, ocean currents can drive NSR in either a prograde or retrograde direction. Therefore, determining the direction of NSR can still provide useful information about subsurface ocean dynamics. The retrograde drift inferred for Enceladus is one example (Schenk & McKinnon, 2024). The rate of NSR reflects not only the torque exerted by the ocean but also the opposing torque arising from deformation of the ice shell, which depends on the shell’s thickness and rigidity. As a result, estimates of the NSR rate are subject to large uncertainties associated with ice rheology. Nonetheless, measuring the NSR rate could provide a constraint on the average zonal flow speed in the upper ocean.

The orientation of the rotation axis (i.e. the obliquity) should be coplanar with the orbit pole and the invariable pole: a Cassini state. Departures from an exact Cassini state are the result of dissipation, either at the ocean-solid interfaces or within the solid layers (Downey & Nimmo, 2025). An obliquity measurement would, therefore, provide at least an upper bound on ocean dissipation.

Directly probing interior structure and dynamics is possible through seismology. Although some attention has been paid to the utility of seismic waves for probing the structure of ice shells (Maguire et al., 2021, e.g.), less attention has been paid to probing the dynamics of the underlying ocean (Panning et al., 2018). On Earth, internal ocean gravity waves have been detected by seismometer (Shaddox et al., 2021), so the topic of passive seismic characterization of ocean dynamics (Wunsch, 2022) seems

ripe for further exploration. The *Dragonfly* mission to Titan will have a seismometer as part of its payload and may provide constraints on Titan’s ice-shell thickness, tidal dissipation rate, and potentially even the dynamics of the underlying ocean.

8 Conclusions

We have reviewed recent progress in understanding the dynamics of the subsurface ocean on icy moons, driven by bottom heating, tidal forcing, and buoyancy fluxes at the ice-ocean interface. In particular, we have emphasized how these processes interact with each other (§3), rather than acting in isolation. Although ocean circulation is primarily energized by tidal work and basal heating, water-ice heat and salinity exchanges associated with ice thickness variations (ITV) act energetically to establish and maintain a stratified layer beneath the ice shell (§4). Once formed, any such stratification strongly modulates the transport of heat, momentum, and tracers within the ocean, with important implications for habitability.

Although subsurface ocean dynamics and their forcings cannot be directly observed, they may be inferred from the more accessible observables of ITV and, potentially, non-synchronous rotation (NSR). The sustainability of any observed ITV places constraints on ocean heat transport (OHT), and therefore on the processes that control OHT. NSR, in turn, reflects ocean momentum transport (OMT) and thus constrains the mechanisms responsible for transferring angular momentum between the ocean and the ice shell, and between the ocean and the silicate core.

Taking Enceladus as an example, we show that the maximum plausible heat production in the core is likely insufficient to overcome the stratification imposed by its poleward-thinning ice shell ($Q_m < Q_c$, see Table. 1). As a result, any heat generated in the core is redirected toward regions of thick equatorial ice, rendering it incapable of sustaining the observed ITV. Tidal dissipation within the ocean drives vertical mixing. As dense fluid is pumped upward, the gravitational potential energy of the ocean is replenished, sustaining baroclinic eddies. These eddies drive efficient equatorward heat transport that converges beneath thick ice, which again acts to erase existing ITV. If the inferred equator-to-pole ITV on Enceladus are robust, these results imply the following for its subsurface ocean.

- Heat production in the core should not exceed $\sim 35\%$ of the total;
- Heat production within the ocean should be negligible;
- The majority of heat production should occur within the ice shell, with polar amplification (Fig. 9);
- Enceladus’ ocean is likely stratified over the polar regions (Fig. 8a), accompanied by strong compositional gradients — the strength and thickness of stratified layer can be estimated using scaling laws;
- This stratification may significantly prolong tracer transport time from the seafloor to the polar ice beyond decadal timescales (Zhang et al., 2025).

Achieving the high dissipation rate within the ice shell suggested here requires either an ice rheology that differs from current assumptions, or a higher orbital eccentricity in the past (P. Goldreich et al., 2025).

However, the data used to constrain Enceladus’ ITV are quite sparse (Iess et al., 2014; Park et al., 2024), highlighting the need for future missions to better constrain them. Furthermore, when observational constraints on ITV become available for other icy satellites (such as Europa), similar inferences can be made. The inferred heat partitioning can be independently tested if the tidal Love numbers k_2 and h_2 can be better constrained through future gravity and radar measurements (Bagheri et al., 2025), as h_2 helps distinguish gravity signals associated with ice-shell deformation from those arising from de-

1190 formation of the solid interior. Beyond heat partitioning, mean ocean salinity S_0 also
 1191 modulates OHT and therefore equilibrium ITV. Possible degeneracies between salinity
 1192 and heat transport may be resolved through independent constraints on ocean salinity,
 1193 potentially obtained from magnetic induction measurements (Biersteker et al., 2023), or
 1194 from direct sampling of plume ejecta (Postberg et al., 2009), combined with appropri-
 1195 ate modeling to relate ejecta salinity to ocean salinity, and to assess the sustainability
 1196 of the south polar geysers (Ingersoll & Nakajima, 2016).

1197 Non-synchronous rotation (NSR) provides an additional constraint on subsurface
 1198 ocean dynamics. Although the basic mechanisms by which convection, tides, and baro-
 1199 clinic eddies drive NSR have begun to be explored in recent years, quantitative scaling
 1200 laws that would enable robust observational inference remain largely undeveloped. Nev-
 1201 ertheless, because convection and tides tend to drive NSR in a direction opposite to that
 1202 driven by baroclinic eddies, the observed direction of NSR alone can potentially serve
 1203 as a diagnostic of the dominant oceanic regime (Table. 2). The magnitude of NSR can
 1204 further constrain the ocean’s momentum transport and energetics, and can be cross-validated
 1205 against constraints derived from ITV, providing an independent test of inferred ocean
 1206 properties.

1207 One factor that has not been considered in the current literature is the influence
 1208 of the high-pressure ice layer present at the base of the oceans of large icy satellites such
 1209 as Ganymede and Callisto. When the depth of the interface between the high-pressure
 1210 ice and the overlying water varies, we hypothesize that circulation may develop near the
 1211 ocean floor, analogous to the circulation driven by topography at the upper ice-ocean
 1212 interface. Such circulation could modify the OHT and thereby induce freezing or melt-
 1213 ing of both the low-pressure and high-pressure ice layers. This process has not yet been
 1214 explored and may be particularly important for large icy moons that host high-pressure
 1215 ice layers.

Appendix A Previous results on the interaction between convection and baroclinic eddies

To compare our results with Kvorka and Čadek (2024), we recast the model configurations from Kvorka and Čadek (2024) into Q/Q_c . To do so, we first rewrite Q/Q_c using the horizontal and vertical Rayleigh numbers, which measure the strength of the under-ice buoyancy gradient and the strength of the bottom heating, respectively,

$$\frac{Q}{Q_c} \sim \frac{\text{Ra}_v}{\Gamma_y^{-3} \text{Ra}_h^{\frac{5}{2}}}, \quad \text{Ra}_h \equiv \frac{\alpha_T g \Delta_h T}{D f^2}, \quad \text{Ra}_v \equiv \frac{\alpha_T g Q}{\rho_0 C_p D^2 f^3} \quad (\text{A1})$$

where the aspect ratio $\Gamma_y \equiv a/D_w$, where a denotes the radius of the planet, and D is the thickness of the ocean.

We then find the relationships between Ra_h , Ra_v and the nondimensional parameters (Ra, Nu, E, Pr, K, Γ_y , $\Delta H/D$) used in Kvorka and Čadek (2024),

$$\text{Ra}_h = \text{Ra Nu E}^2 \text{Pr}^{-1} K \frac{\Delta H}{D}, \quad \text{Ra}_v = \text{Ra Nu E}^3 \text{Pr}^{-2}, \quad (\text{A2})$$

where Nu denotes the Nusselt number and E the Ekman number. ΔH represents the thickness variation of the ice–water interface, read from their Fig. 5b. The parameter $K \equiv -\frac{\partial T_m / \partial r}{\partial T_w / \partial r}$, defined in Kvorka and Čadek (2024), measures the ratio between the vertical freezing-temperature gradient $\frac{\partial T_m}{\partial r} = -\frac{dT_m}{dp} \rho_0 g$ arising from the pressure dependence of the melting point and the boundary-layer temperature gradient $\frac{\partial T_w}{\partial r}$ imposed by the bottom heat flux. A larger K corresponds to a stronger melting-point shift. In Kvorka and Čadek (2024), K ranges from 0 to 10. Substituting the parameters for experiments 13–17, we find that Q/Q_c varies from 80 to 5000. Table A1 summarizes the parameter choices of Kvorka and Čadek (2024) on the left and the recast Ra_v , Ra_h and Q/Q_c on the right.

Exp#	Nu	K	$\Delta H/D$	$\Delta Q/Q_0$	Ra_h	Ra_v	Q/Q_c
13	45.3	0.3	4.5×10^{-3}	0.02	0.020	4.42×10^{-4}	4978
14	45.3	1.0	3.2×10^{-3}	0.05	0.047	4.42×10^{-4}	575
15	45.3	3.0	1.8×10^{-3}	0.075	0.079	4.42×10^{-4}	156
16	45.6	6.0	1.0×10^{-3}	0.1	0.089	4.44×10^{-4}	118
17	45.4	10.0	7.0×10^{-4}	0.1	0.103	4.43×10^{-4}	81

Table A1: Parameters used in Kvorka and Čadek (2024) are summarized in the left panel, and the corresponding Ra_h , Ra_v , and Q/Q_c are shown on the right. The shared parameters: $\text{Ra} = 3.61 \times 10^7$, $\text{E} = 3 \times 10^{-4}$, $\Gamma_y = 5.1$, and $\text{Pr} = 10$ are omitted from the left panel to save space. $\Delta Q/Q_0$ is the heat deflection rate.

Appendix B Derivation of Eq.22

We first write down the heat budget for equatorial and polar ice separately and take the ratio between the two to get

$$\frac{\mathcal{H}_{\text{cond},p}}{\mathcal{H}_{\text{cond},e}} = \frac{\mathcal{H}_{\text{ocn},p} + \mathcal{H}_{\text{ice},p} + L_f q_p}{\mathcal{H}_{\text{ocn},e} + \mathcal{H}_{\text{ice},e} + L_f q_e}, \quad (\text{B1})$$

where the subscript e, p denotes equatorial and polar, respectively, and L_f denotes the latent heat of freezing and q denotes the freezing rate in $\text{kg/m}^2/\text{s}$.

As discussed in §3, the ITV on Enceladus is strong enough to stratify the ocean even with all of the heat generated in the silicate core ($Q_b < Q_c$ in Table. 1). This means that all heat generated in the core and ocean $\mathcal{H}_{\text{int}}$ will converge toward the equatorial regions, while the polar ice loses heat into the ocean,

$$\mathcal{H}_{\text{ocn,p}} \lesssim 0, \quad \mathcal{H}_{\text{ocn,e}} \gtrsim 2\overline{\mathcal{H}_{\text{int}}} > 2\overline{Q_b} \quad (\text{B2})$$

In addition, we recognize that maintaining a poleward-thinning ice geometry against ice flow (Ashkenazy et al., 2018) requires net melting in the polar regions and freezing in the equatorial regions, i.e., $q_p < 0$, $q_e > 0$. Furthermore, while dissipation within the ice can be amplified by polar waves, the equatorial region must still experience positive dissipation, i.e. $\mathcal{H}_{\text{ice,p}} < 2\mathcal{H}_{\text{ice}}$, $\mathcal{H}_{\text{ice,e}} > 0$. Substituting all of these into Eq(B3), we get

$$\frac{\overline{\mathcal{H}_{\text{ice}}}}{\overline{Q_b}} > \frac{\mathcal{H}_{\text{ocn,p}} + \mathcal{H}_{\text{ice,p}} + L_f q_p}{\mathcal{H}_{\text{ocn,e}} + \mathcal{H}_{\text{ice,e}} + L_f q_e} = \frac{\mathcal{H}_{\text{cond,p}}}{\mathcal{H}_{\text{cond,e}}} \approx \frac{H_e}{H_p} \quad (\text{B3})$$

Appendix C Compute the equilibrium ice thickness profile for a given heat partition.

To compute the geometry of the equilibrium ice shell, we integrate an ice evolution model, which is modified from the one in Kang and Flierl (2020). The model calculates the melting induced by the tidal heating in the ice $\mathcal{H}_{\text{ice}}$ (Beuthe, 2019), the down-gradient ice flow $\mathcal{Q}$ (Ashkenazy et al., 2018), the heat loss to space by conduction $\mathcal{H}_{\text{cond}}$, and heat transmitted upward by the ocean $\mathcal{H}_{\text{ocn}}$, and evolves the ice thickness H over time. The total thickness tendency can be symbolically expressed as

$$\frac{dH}{dt} = \frac{\mathcal{H}_{\text{cond}}(H) - \mathcal{H}_{\text{ice}}(H) - \mathcal{H}_{\text{ocn}}(H)}{L_f \rho_i} + \frac{1}{a \sin \phi} \partial_\phi (\sin \phi \mathcal{Q}(H)), \quad (\text{C1})$$

where L_f and ρ_i are the latent heat of freezing and density for ice, a is the moon's radius and ϕ denotes latitude. The analytical forms of $\mathcal{H}_{\text{cond}}$, $\mathcal{H}_{\text{ice}}$ and $\mathcal{Q}$ can be found in Kang and Flierl (2020). The latitudinal surface temperature variations associated with the poleward weakening solar forcing is accounted. This leads to an equator-to-pole surface temperature contrast of $\sim 10\text{K}$. When the ice shell is thinner than 3 km, we assume that the ice shell will crack open under tidal stress and the resulting geysers will absorb large amounts of heat, preventing further melting. In the evolution model, we overwrite the thickness tendency with zero when $H < 3$ km, to implicitly represent this additional heat loss.

The ocean-ice heat exchange $\mathcal{H}_{\text{ocn}}$ should be parameterized. Here, we consider two components. The first is the deflected bottom heating (§3.1)

$$\mathcal{H}_{\text{ocn,Qb}} = \overline{Q_b} \frac{H'^2_{>}}{H'^2_{>}} \quad (\text{C2})$$

where H' is the deviation from the prescribed global mean ice thickness $H_0 = 20$ km, $H'_{>} = \max\{H', 0\}$ and $\overline{(\cdot)}$ denotes the global average. In this formulation, bottom heating is applied to ice thicker than average, with a first-order smooth transition between heated and unheated regions.

The second component is heat transport from thin ice regions to thick ice regions due to tidal mixing in the interior of the ocean (§3.2). Eq. (11) can be rewritten as

$$\mathcal{H}_{\text{ocn,bci}} \equiv \frac{\mathcal{F}_{\text{h,bci}}}{\pi a^2} = \frac{2C_p \rho_0 \Gamma \epsilon}{\alpha T g}. \quad (\text{C3})$$

It can be shown that the salinity effect will contribute another term to $\mathcal{H}_{\text{ocn,bci}}$, but the influence is negligible (Kang & Zhang, 2026). The complete $\mathcal{H}_{\text{ocn}}$ is the summation of (C2)

and (C3). We caution that this parameterization assumes that the ITV is sufficiently strong to stratify the upper ocean, i.e., $\Delta H > \Delta H_c$ (Table 1). Although this assumption is typically justified given the smallness of ΔH_c , when ΔH does not exceed ΔH_c , the ITV will be underestimated. Developing a more general formulation for ocean heat transport remains an important task for future work.

Open Research Section

Fig. 3, Fig. 4, Fig. 6 and Fig. 7 are adapted from previous works as indicated in the caption. Data to reproduce Fig. 2 are available through (Song et al., 2024b), (Song et al., 2024a) and (Gastine et al., 2016). Part of data used to generate Fig. 5 are available through (Hay & Matsuyama, 2019). The remaining data together with the data for Fig. 9 will be shared through Zenodo and can be accessed at <https://doi.org/10.5281/zenodo.21650412>.

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