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The persistent impact of tropical cyclones over the Northwest Pacific on Arctic sea ice

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Abstract

25 Tropical cyclones (TCs) over the western North Pacific (WNP) are known to
26 intensify El Niño, and recent work has linked the preceding autumn WNP TC activity to
27 the following spring sea ice in the Beaufort Sea. Here we extend this connection into the
28 summer melt season and diagnose, term by term, how the signal reaches the western Arctic.
29 Using the Accumulated Cyclone Energy (ACE) index derived from IBTrACS (1970–2024)
30 together with ERA5, HadISST, and a regression-based decomposition, partial-correlation
31 analysis, Rossby-wave ray tracing, a perturbation hypsometric diagnosis, and a surface-
32 flux decomposition, we show that enhanced August–October ACE is associated with an
33 intensified following-winter El Niño (an amplification attributed by earlier work to the
34 tropical cyclones themselves; Wang et al. 2019), whose tropical-Pacific convective heating
35 excites a stationary Rossby wave train and a Pacific–North American-like pattern that
36 transports energy poleward to the Beaufort Sea. The resulting circulation anomalies
37 reorganize the surface energy budget: net longwave radiation dominates the response
38 during and just after polar night, whereas net shortwave radiation — amplified by the sea-
39 ice–albedo feedback — dominates from spring into summer, with latent and sensible heat
40 fluxes playing a secondary role. The ACE- and El Niño-related pathways drive broadly
41 similar surface-energy changes but differ in the relative importance of longwave versus
42 shortwave radiation. Atmosphere-only CAM5 experiments show that an ENSO-like
43 tropical-Pacific SST anomaly forces this teleconnection: the ACE-congruent forcing
44 reproduces the El Niño 500-hPa response pattern almost exactly (pattern correlation 0.95)
45 at a proportionally reduced amplitude. Finally, evaluated by cross-validation to avoid
46 overfitting on the short record, the preceding ACE index provides a modest but genuine

source of predictive information for the following spring–summer Beaufort sea ice: at long lead (≈ 7 – 10 months) the ACE index alone yields a leave-one-out skill of $r \approx 0.3$, and combining it with the observed winter sea-ice state in a shorter-lead configuration raises this to $r \approx 0.45$. The ACE contribution is complementary to sea-ice persistence and is most valuable in the melt season, where local persistence carries no memory; its skill beyond the winter Niño-3.4 index is positive but, on this short record, not statistically distinguishable from zero.

1 Introduction

Arctic sea ice is a central component of the global climate system, regulating the surface energy budget through its high albedo, insulating the ocean from the atmosphere, and modulating high-latitude atmospheric and oceanic circulation. Over recent decades the Arctic has warmed rapidly and the sea-ice cover has declined sharply in both extent and thickness (Serreze and Meier, 2019; Stroeve et al. 2012). The Beaufort Sea, in the western Arctic, is one of the regions where this decline is most pronounced and most variable from year to year, and its interannual variability has been linked to modes such as the North Atlantic Oscillation (Liu et al. 2024). Because the loss of Beaufort sea ice feeds back on regional and hemispheric climate and has large ecological and societal consequences, identifying the drivers of its interannual variability — and any sources of seasonal predictability — is of considerable interest.

Although the most familiar controls on Arctic sea ice are local (surface radiation, ocean heat, and ice dynamics), a growing body of work shows that remote forcing from the tropics can reach the Arctic through large-scale atmospheric teleconnections. The El

Niño–Southern Oscillation (ENSO), in particular, perturbs the tropical Pacific heat source, excites stationary Rossby wave trains, and projects onto the Pacific–North American (PNA) pattern, thereby influencing high-latitude circulation, poleward energy and moisture transport, and ultimately sea ice (Lee et al. 2017; Chen et al. 2019; Li et al. 2019). Indeed, tropical-Pacific SST variability has been identified as one of the most effective remote drivers of Arctic sea-ice change, both in observational attribution (Baxter et al. 2019) and in model linear-response analyses (Wu et al. 2021). Enhanced poleward transport of heat and moisture increases downward longwave radiation and can accelerate sea-ice melt, while changes in cloud and surface radiation modify the shortwave budget once polar night ends (Woods and Caballero 2016; Villamil-Otero et al. 2017); intensified moisture intrusions and atmospheric rivers, in particular, can slow the seasonal recovery of Arctic sea ice (Zhang et al. 2023). These pathways establish that low-latitude signals, given a suitable bridge, can modulate the western Arctic surface energy budget.

A more recent and less-explored link in this chain is tropical-cyclone (TC) activity over the western North Pacific (WNP). TCs pump heat and momentum into the upper ocean and atmosphere, and their integrated seasonal effect can be quantified by the Accumulated Cyclone Energy (ACE) index (Bell et al. 2000). WNP TC activity is itself strongly modulated by ENSO (Camargo and Sobel 2005; Patricola et al. 2018), but the relationship is not one-way: Wang et al. (2019) demonstrated that WNP TCs act to intensify El Niño, as the cyclone-induced westerly wind anomalies and oceanic mixing project onto the developing ENSO and strengthen the subsequent winter El Niño. WNP ACE has also been related to the Pacific–North American (PNA) pattern (Song et al. 2022), pointing to an extratropical pathway by which the TC signal can reach high latitudes. Building on this,

Ye et al. (2024) showed that the preceding August–October (ASO) ACE over the WNP is statistically linked to the following spring sea ice in the Beaufort Sea, with El Niño acting as the bridge. The physical chain implied by these studies is: WNP ACE → intensified El Niño → tropical-Pacific convective heating → Rossby wave train / PNA-like teleconnection → anomalous western-Arctic circulation and surface energy budget → Beaufort sea-ice change.

The sign and mechanism of the TC–Arctic connection, however, remain unsettled. In contrast to the sea-ice *reduction* pathway emphasized here, Zeng et al. (2024) reported that WNP TC activity can *favor* an increase in Arctic sea ice, and related work has extended the TC–Arctic link to the eastern North Pacific (Zeng et al. 2025), underscoring that the teleconnection is sensitive to season, region, and the specific energy-budget terms involved. This motivates a more complete, term-by-term diagnosis of how the ACE-related signal reaches the Beaufort Sea and which surface fluxes dominate in different seasons.

The present study extends the ACE–ENSO–Arctic framework of Wang et al. (2019) and Ye et al. (2024) in three ways. First, we follow the signal beyond spring into the summer Beaufort sea ice, the season of greatest melt. Second, we provide a term-by-term decomposition of the surface energy fluxes (net longwave, net shortwave, latent, and sensible heat fluxes), combined with a perturbation hypsometric diagnosis and Rossby-wave ray tracing, to isolate the dynamical pathway and the dominant radiative drivers in each season. Third, we assess whether the preceding ACE index, which leads the sea ice by roughly three seasons, carries any practical predictive information for Beaufort summer sea ice.

The remainder of the paper is organized as follows. Section 2 describes the data and methods, including the ACE index, the regression-based signal decomposition, the perturbation hypsometric equation, Rossby-wave ray tracing, and the surface energy budget. Section 3 documents the impact of the preceding ASO ACE on the following spring sea ice and the bridging role of El Niño. Section 4 examines the summer Beaufort sea-ice variability through the surface energy budget. Section 5 evaluates the predictive skill of the ACE, Niño-3.4, and combined indices. Section 6 summarizes the results and discusses limitations and implications.

2 Data and Methods

2.1 Data

The International Best Track Archive for Climate Stewardship (IBTrACS) from the National Oceanic and Atmospheric Administration (NOAA; Knapp et al. 2010, 2018) is used to obtain TC data from 1970 to 2024. Monthly $1^{\circ} \times 1^{\circ}$ SST data for the period 1970–2025 are taken from the Met Office Hadley Centre Sea Ice and Sea Surface Temperature data set (HadISST). Monthly $1^{\circ} \times 1^{\circ}$ atmospheric, $0.25^{\circ} \times 0.25^{\circ}$ surface radiation, and sea-ice concentration (SIC) datasets are obtained from the fifth-generation European Centre for Medium-Range Weather Forecasts Reanalysis (ERA5; Hersbach et al. 2020). All datasets were smoothed using a three-month running mean over the 1970–2024 period. El Niño events are selected following the NOAA standard (<http://ggweather.com/enso/oni.htm>), where the Niño-3.4 (N3.4) index is the running three-month mean of SST anomalies in the Niño-3.4 region (5°S – 5°N , 120° – 170°W). The

Beaufort sea-ice index analysed throughout is the area-averaged, linearly detrended SIC over the box 69°–75°N, 210°–240°E (i.e. 150°–120°W).

2.2 Accumulated Cyclone Energy index

The ACE index is an indicator representing the cumulative kinetic energy of TCs (Bell et al. 2000; Wang et al. 2019; Wang and Li 2022a, b, c; Ye et al. 2024). The ACE in each 5°×5° grid cell is defined as the sum of the squares of the estimated maximum sustained surface wind speed (in knot²) for all TCs occurring within the grid cell over all 6-hourly periods in which the maximum sustained surface wind speed exceeds 30 knot. The ACE index is defined as the anomaly of the sum of the ACE over all grid cells in the selected region (10°–25° N, 120°–160° E). Strong ACE events are defined when the ACE index exceeds 0.7 standard deviations, which is used to characterize years of ACE enhancement.

2.3 Atmospheric indicators affected by ACE

Following previous studies (Wang et al. 2019; Wang and Li 2022b, c), El Niño intensity can be enhanced by the ASO ACE over the WNP (Fig. 1b). To highlight the impact of ACE on El Niño, ACE-enhanced El Niño events are defined when both ACE and ENSO are strong events (Fig. 1c). Thus, the contribution of ASO ACE over the WNP to atmospheric variables via El Niño ($y_{d_{N3.4}}$) can be expressed as follows:

$$d_{N3.4} = a_0 x_{ACE} + b_0, \quad (1)$$

$$y_{d_{N3.4}} = a_1 d_{N3.4} + b_1, \quad (2)$$

where a_0 and a_1 are regression coefficients, b_0 and b_1 are constants, x_{ACE} indicates the WNP ACE anomalies in ASO, and $d_{N3.4}$ is the contribution of ACE to the November–December–January (NDJ) N3.4 index.

Meanwhile, atmospheric variables affected by the N3.4 index can be calculated as:

$$y_{N3.4} = a_1 x_{N3.4} + b_1, \quad (3)$$

where a_1 and b_1 are the regression coefficient and a constant, respectively, and $x_{N3.4}$ indicates the SST anomalies for the N3.4 region in NDJ. The two-tailed Student’s t -test is employed to determine statistical significance throughout this study; because the analysed series are three-month running means, we verified that significance is essentially unchanged when the effective number of degrees of freedom (accounting for lag-1 autocorrelation) is used.

We emphasize that Eqs. (1)–(2) provide a *statistical* decomposition, not a direct proof of causality. Autumn WNP ACE is itself strongly related to the concurrently developing El Niño (the ASO ACE index correlates with the concurrent ASO Niño-3.4 at $r \approx 0.6$), and that developing signal persists into winter. In our data, controlling for the concurrent ASO Niño-3.4 reduces the apparent autumn-ACE $\rightarrow$ winter-N3.4 relationship from $r \approx 0.65$ to a non-significant $r \approx 0.18$. The regression term $d_{N3.4}$ should therefore be read as the ACE-*congruent* component of the winter ENSO signal rather than as evidence that TCs force ENSO. The proposition that WNP TCs *amplify* a developing El Niño is not established by the present observational analysis; it rests on the independent mechanism and modelling of Wang et al. (2019). Our own CAM5 experiments (Section 4) do not test this upstream step — they prescribe the tropical-Pacific SST anomaly and diagnose only

the *downstream* (SST $\rightarrow$ Arctic) response. Accordingly, we treat the ACE-related signal throughout as the fraction of the winter El Niño anomaly that is congruent with the preceding ACE, and confine our causal statements to the downstream teleconnection that the experiments can support.

2.4 Other methods

2.4.1 Perturbation hypsometric diagnosis

According to the hypsometric equation, the thickness ΔZ of the atmospheric layer between the pressure surfaces p_1 and p_2 can be calculated as follows (Holton and Hakim 2013; Li et al. 2022):

$$\Delta Z = Z_2 - Z_1 = \frac{R \bar{T}}{g_0} \ln \frac{p_1}{p_2}, \quad (4)$$

where p_1 (p_2) indicates pressure, Z_1 (Z_2) is the geopotential height at p_1 (p_2), $g_0 = 9.80665 \text{ m s}^{-2}$ is the global-average gravity at mean sea level, and $R = 287 \text{ J kg}^{-1} \text{ K}^{-1}$ is the gas constant for dry air. Here $\bar{T}$ denotes the mean temperature of the layer. Letting $\bar{T}'$ and $\Delta Z'$ be the deviations (anomalies) from their time averages, one obtains the perturbation hypsometric equation (Li et al. 2022):

$$\bar{T}' = \frac{g_0}{R \ln(p_1/p_2)} \Delta Z'. \quad (5)$$

The mean temperature of the perturbation layer therefore has a proportional relationship with the thickness of the perturbed layer bounded by isobaric surfaces: when the perturbation-layer thickness increases, the perturbation-layer mean temperature rises,

and vice versa. This perturbation equation is used to discuss the influence of tropospheric circulation anomalies on surface air temperature variations.

2.4.2 Rossby wave ray tracing

In addition, to verify the pathways by which SST anomalies in the Niño-3.4 region (induced by both El Niño and ACE) affect the sea-ice cover of the Beaufort Sea, Rossby wave ray-tracing theory is used to identify the trajectory of the stationary Rossby wave train (Li et al. 2015; Zhao et al. 2015, 2019). Based on the original theory of Hoskins and Karoly (1981), Zhao et al. (2015) assume a horizontally nonuniform basic flow. Unlike the classical theory of Hoskins and Karoly (1981), in a horizontally nonuniform basic flow the horizontal wavenumber along a wave ray is not fixed but varies as the ray propagates. Through this Rossby-wave ray-tracing equation, the propagation path of the Rossby waves and the degree of concentration of Rossby-wave energy can be clearly identified.

2.4.3 Atmospheric heat transport

Atmospheric heat transport is regarded as one of the crucial factors influencing sea ice (Woods and Caballero 2016; Gong et al. 2017; Lee et al. 2017): the energy exchange between sea ice and the atmosphere directly determines the local energy budget and thereby affects sea-ice variability. The surface net energy flux over the Beaufort Sea can be obtained from the following equation:

$$Q = Q_{nlw} + Q_{nsw} + Q_{lh} + Q_{sh}, \quad (6)$$

where Q_{nlw} is the net downward longwave radiation flux, Q_{nsw} is the net downward shortwave radiation flux, Q_{lh} is the latent heat flux, and Q_{sh} is the sensible heat flux; the

downward direction is taken as positive for all variables. From this, the surface net energy flux anomalies over the Beaufort Sea associated with the boreal summer–autumn ACE and the winter N3.4 can be calculated.

2.4.4 Additional methods

The statistical significance of correlations between highly autocorrelated time series was assessed using the effective degrees of freedom (Pyper and Peterman 1998; Li et al. 2013; Xie et al. 2014; Sun et al. 2015). Composite analysis and partial correlation analysis are also employed.

3 The impact of the preceding ACE (ACE in August–October) on spring sea ice

The occurrence of tropical cyclones in the western North Pacific (WNP) peaks between August and October. The intensity of tropical-cyclone activity can influence sea-ice variability over the Beaufort Sea (Ye et al. 2024). Here, the ACE index is used to represent the climate effect of the WNP tropical cyclones. The composite difference between strong and weak ACE events (Fig. 1a, b) shows that years of enhanced preceding ACE are followed by reduced sea ice in the Beaufort Sea. Consistent with this, the standardized ACE and SIC time series covary, with enhanced ACE preceding reduced Beaufort SIC (Fig. 1c). Because previous research shows that ACE can significantly enhance ENSO after about three months (Wang et al. 2019; Ye et al. 2024), El Niño is a natural candidate for the bridge linking the autumn ACE to the following Beaufort sea ice.

We next examine the relationship among the preceding ASO ACE over the WNP, the NDJ N3.4 index, and the following SIC in the Beaufort Sea. During the period of

significant negative correlation between the preceding ASO ACE over the WNP and the following SIC in the Beaufort Sea, a substantial reduction in correlation is observed when the effect of the NDJ N3.4 signal is removed (the ASO-ACE-SIC correlation weakens from $r = -0.38$, $p = 0.005$, to a non-significant $r = -0.17$, $p = 0.24$, once NDJ Niño-3.4 is partialled out; Fig. 1d). Thus, the NDJ N3.4 signal may play a crucial role in the process by which ASO TCs over the WNP affect SIC in the Beaufort Sea. At the same time, the lag correlation between NDJ N3.4 and Beaufort SIC is strongly affected by removal of the ASO ACE signal (Fig. 1e), and the magnitude of the correlation coefficient changes markedly, though it remains marginally significant (from $r = -0.42$ to $r = -0.26$, $p = 0.06$). In both partial correlations the effective number of degrees of freedom is essentially equal to the sample size, so the reduced significance reflects the shared ACE-ENSO variance (the two indices correlate at $r \approx 0.6$) rather than autocorrelation. Hence, the ASO ACE and NDJ N3.4 signals are strongly intertwined, and the ACE-Beaufort-SIC relationship is mediated to a large degree by the winter El Niño (consistent with the confounding discussed in Section 2.3). Statistically, these results are consistent with the hypothesis that the preceding enhanced ASO ACE over the WNP is associated with a reduction of sea ice in the Beaufort Sea in the following spring and summer, with the winter El Niño acting as the bridge. The following dynamical diagnosis provides further evidence for this hypothesis.

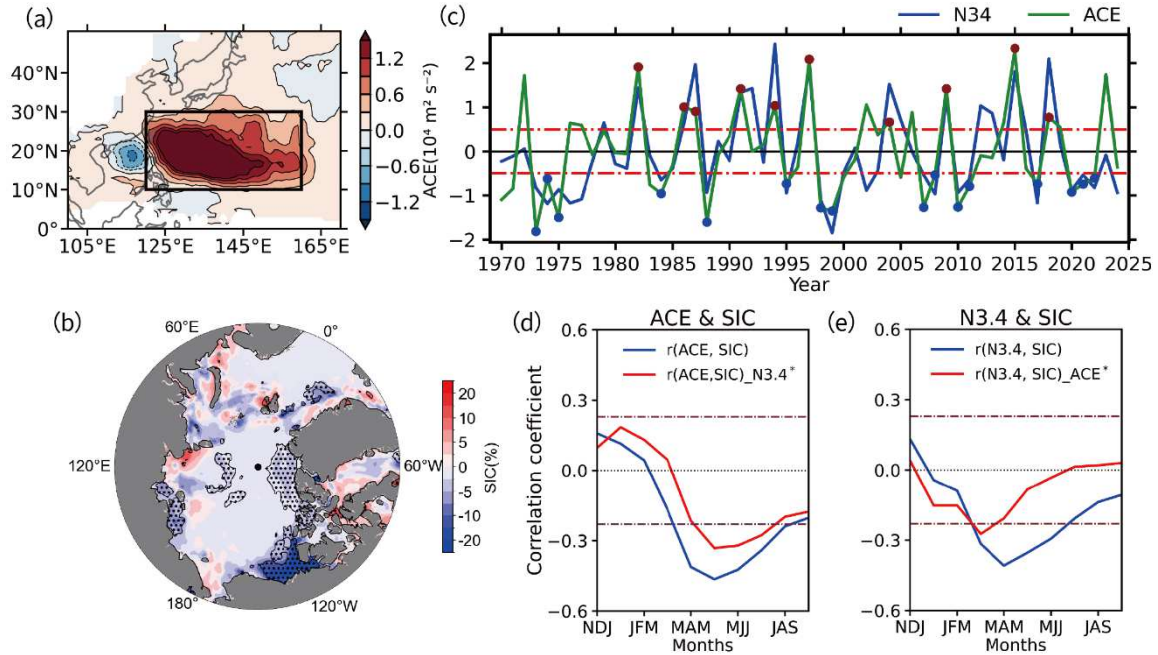


Figure 1. Composite anomalies of the climatological anomaly of (a) August–September–October (ASO) accumulated cyclone energy (ACE) pattern over the western North Pacific (WNP), and (b) the following April–May–June (AMJ) Arctic sea-ice concentration (SIC) between strong and weak ACE events from 1970 to 2024. (c) Standardized time series of ACE and SIC. (d) Lag correlation coefficients (blue) between the preceding ASO ACE and SIC in the Beaufort Sea, and their partial correlation after removing the NDJ N3.4 (red). (e) Lag correlation coefficients (blue) between the NDJ N3.4 and the SIC in the Beaufort Sea, and their partial correlation after removing the preceding ASO ACE over the WNP (red). The red and blue dotted lines denote significance at the 95% confidence level.

The observations are consistent with a preceding-ACE influence on Beaufort sea-ice melt. Under the mechanism proposed by Wang et al. (2019), in which the preceding ASO ACE over the WNP enhances the intensity of the winter (November–December–January, NDJ) El Niño (Fig. 2a), the associated tropical-Pacific convective activity intensifies. These anomalous convective activities excite Rossby wave trains propagating

toward the Arctic, which trigger a Pacific/North American (PNA)-like teleconnection pattern in the following spring. This process transports energy from the tropical Pacific to the Beaufort Sea and leads to anomalous atmospheric circulation in this area, including anomalies of atmospheric thickness and the surface pressure gradient. These anomalies affect sea-ice melt in the Beaufort Sea in the following spring through surface temperature anomalies, water-vapor anomalies, radiation anomalies, and sea-ice transport anomalies. Specifically, Fig. 2 shows the eastward and poleward evolution of the ACE-related tropical SST (Fig. 2a), the 500-hPa geopotential-height response and the traced Rossby-wave path into the western Arctic (Fig. 2b), and the resulting thickness, near-surface temperature, humidity, and water-vapor-flux-divergence anomalies over the Beaufort Sea (Fig. 2c–f).

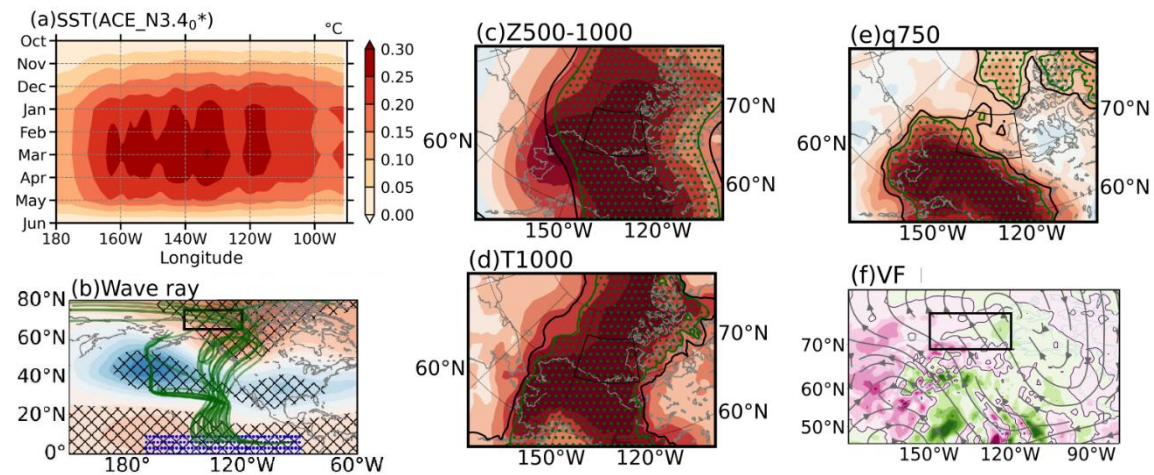


Figure 2. (a) Evolution of composites of the meridional-mean tropical-Pacific (5°S – 5°N) SST anomalies ($^{\circ}\text{C}$; shading) associated with the preceding ACE anomalies, after removing the signal of the ASO SST anomalies in the tropical Pacific. (b) 500-hPa geopotential height (gpm; shading) and stationary Rossby-wave trajectories with zonal wavenumbers in the range 1–4 at 500 hPa in a horizontally nonuniform climatological flow (green curves),

(c) 500–1000-hPa atmospheric thickness (gpm; shading), (d) surface temperature (K; shading), (e) 750-hPa specific-humidity anomalies (10^{-2} g kg⁻¹), and (f) 700–1000-hPa water-vapor flux divergence (kg m⁻² s⁻¹; shading) in the Beaufort Sea in the following El Niño decaying year.

4 The variability of the sea ice over the Beaufort Sea in summer

The surface heat-transport anomalies over the Beaufort Sea reveal the subsequent energy changes. From the FMA (February–March–April) season of the year following strong ACE years, the surface net energy flux anomalies over the Beaufort Sea associated with the summer–autumn ACE show positive anomalies over part of the region (Fig. 3a). In the following months, this region of positive anomalies spreads rapidly from the southern Beaufort Sea to the entire Beaufort Sea (Fig. 3b–d); the surface net energy flux is positive, and the whole Beaufort Sea becomes covered by relatively strong positive surface radiation anomalies.

In the year following strong ACE years, the responses of the individual surface energy-flux terms associated with the preceding summer–autumn ACE are shown in Fig. 3. In the FMA season, as Fig. 3e–h show, the positive surface net energy flux anomaly associated with the summer–autumn ACE is dominated by the downward longwave radiation anomaly. By the MAM (March–April–May) season, the net longwave radiation anomaly (Fig. 3i) is still the leading cause of the surface net energy flux anomaly over the Beaufort Sea; at this time polar night ends, the subsolar point moves progressively northward, and the net shortwave radiation (Fig. 3j) anomaly increases significantly along the southeastern coast of the Beaufort Sea. In the AMJ (April–May–June) season,

313 significant shortwave radiation anomalies cover most of the Beaufort Sea; as the subsolar
314 point continues to move northward and sea ice continuously decreases, the albedo is
315 reduced, the ocean surface absorbs more and more shortwave radiation, and the downward
316 shortwave radiation flux comes to dominate the change in surface net energy (Fig. 3n). The
317 latent heat flux reflects the phase changes of sea ice (melting and freezing), while the
318 sensible heat flux is determined mainly by the air–ice temperature difference and the wind
319 speed and represents a delayed atmospheric feedback to sea-ice reduction. Over most of
320 the Beaufort Sea the magnitudes of these two terms are small — especially after the rapid
321 melting of sea ice — so we conclude that latent (Fig. 3g, k, o) and sensible heat fluxes (Fig.
322 3 h, l, p) play only a minor role in the influence of the summer–autumn ACE-related
323 tropical-Pacific SST anomalies on the surface net energy flux of the Beaufort Sea.

324 In summary, we conclude that the tropical-Pacific SST warming associated with
325 increased summer–autumn ACE and the increased winter N3.4 can induce broadly similar
326 surface-energy changes over the Beaufort Sea during El Niño decaying years, favoring the
327 melting of sea ice in this region. However, the two processes differ in their mechanisms
328 and controlling factors. When the SST anomalies associated with the summer–autumn
329 ACE increase, the anomalous increase in atmospheric humidity during polar night is
330 comparatively strong and the influence of the net longwave radiation is larger, which is
331 more favorable for the melting of Beaufort sea ice in the following year; after polar night
332 ends, the net shortwave radiation anomaly plays a more important role and, through the
333 sea-ice–albedo feedback, favors the maintenance and amplification of the response. By
334 contrast, when the winter N3.4 anomaly increases, the net longwave radiation anomaly is

smaller, and the influence on sea-ice melt occurs mainly through a process dominated by the net shortwave radiation anomaly.

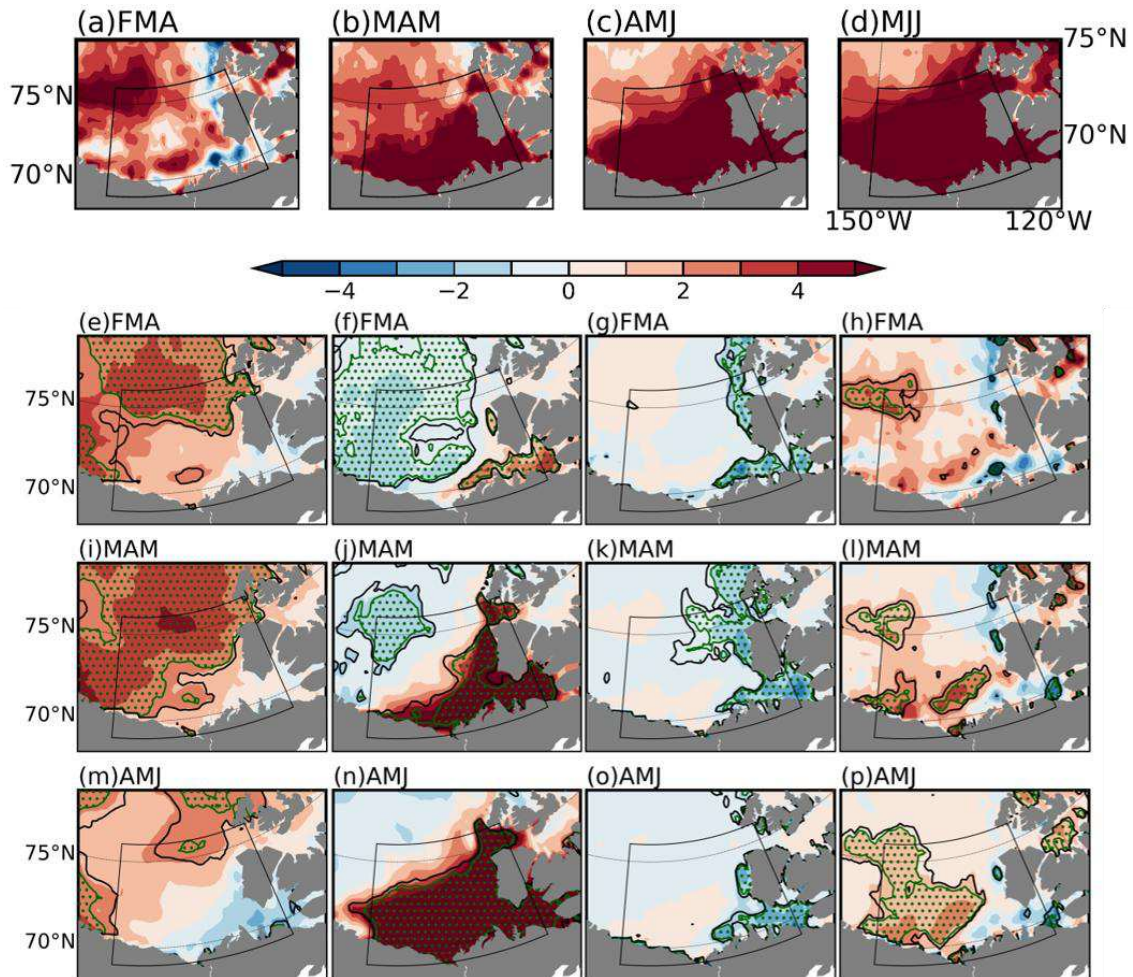


Figure 3. Composite patterns of the contribution of the enhanced SST anomalies related to the preceding ACE to the Beaufort Sea surface energy budget, by season. Top row: net surface heat flux (W m^{-2} ; shading) for (a) FMA, (b) MAM, (c) AMJ, and (d) MJJ. Subsequent rows show the individual flux components for the same four seasons: net longwave radiation (e, i, m), net shortwave radiation (f, j, n), latent heat flux (g, k, o) and sensible heat flux (h, l, p), in the following El Niño decaying year.

345 To test whether the tropical-Pacific SST anomaly associated with ACE can drive
346 this Beaufort response — that is, to isolate the *downstream* (SST to Arctic) half of the chain
347 — we performed five 35-year atmosphere-only experiments with the Community
348 Atmosphere Model version 5 (CAM5) at $1.9^\circ \times 2.5^\circ$ horizontal resolution with 30 vertical
349 levels (finite-volume dynamical core; Table 1). We emphasize at the outset that these are
350 **not** experiments in which tropical cyclones are present; they impose a prescribed SST
351 anomaly and therefore test only the atmospheric response to that anomaly, not the upstream
352 generation of ENSO by TCs. A control run uses climatological SST globally. In the ACE-
353 forced experiment, the SST anomaly obtained by regressing tropical-Pacific SST onto the
354 ASO WNP ACE index is superimposed over the central–eastern tropical Pacific (5°S – 5°N ,
355 170° – 90°W), with climatological SST elsewhere; the N3.4-forced experiment is identical
356 but uses the SST anomaly regressed onto the NDJ Niño-3.4 index. Because the two
357 regression patterns are nearly identical (spatial correlation ≈ 0.99 , the ACE-regressed
358 anomaly being $\approx 70\%$ the amplitude of the N3.4-regressed one; §4), the ACE-forced run is
359 best understood as a **scaled ENSO-like forcing** rather than as a physically distinct “ACE”
360 SST fingerprint — a point we return to below. Each is run with positive and negative-
361 amplitude counterparts (denoted ACE/ACE_ne and NINO/NINO_ne). In both experiments
362 the imposed anomaly follows the seasonal evolution of its parent index — near zero in
363 summer, growing through autumn, peaking in winter (Niño-3.4 amplitude ≈ 1.2 K in
364 December for the ACE run) and decaying through the following spring. Because the WNP
365 ACE index is itself closely tied to the developing ENSO (Section 3), the ACE-regressed
366 SST anomaly closely resembles the canonical El Niño pattern, at a somewhat weaker

amplitude (roughly 70% of the N3.4-regressed anomaly); the two forcings therefore differ mainly in amplitude rather than in spatial structure.

Table 1. Design of the CAM5 experiments (atmosphere-only; prescribed SST; 35 years each).

Experiment	Prescribed SST boundary condition
CTRL	Global climatological SST
ACE	CTRL + the tropical-Pacific SST anomaly regressed onto the ASO WNP ACE index, imposed over 5°S–5°N, 170°–90°W (positive amplitude)
ACE_ne	As ACE, but with the sign of the imposed anomaly reversed (negative amplitude)
NINO	CTRL + the tropical-Pacific SST anomaly regressed onto the NDJ Niño-3.4 index, same region (positive amplitude)
NINO_ne	As NINO, but negative amplitude

The imposed anomaly follows the seasonal evolution of its parent index (near zero in summer, peaking in winter).

Relative to the control, the positive-ACE run reproduces the *spatial pattern* of the El Niño Arctic teleconnection with high fidelity: the Northern-Hemisphere (20°–90°N) 500-hPa height response is correlated with that of the NINO run at 0.95 (95% bootstrap CI [0.84, 0.97]), a robust PNA-like wave train. Consistent with the $\approx 70\%$ amplitude of the ACE forcing, the *magnitude* of the response scales down accordingly. The full El Niño (NINO) run produces a statistically significant spring Beaufort skin-temperature response of +1.17 K (95% CI ± 0.62 K over the 35 model years, box 72°–80°N, 200°–230°E). The ACE-run response over the same box, however, is not distinguishable from internal variability on a single 35-year realization ($\Delta T_s = -0.08 \pm 0.72$ K; the positive-minus-

negative ACE contrast, which isolates the forced linear response, is $+0.18 \pm 0.66$ K — of the expected sign but not significant). We therefore rest the mechanistic conclusion on the two results that *are* robust: the highly significant full-El-Niño Beaufort response, and the near-identical teleconnection pattern excited by the ACE-congruent forcing. Together these establish that an ENSO-like tropical-Pacific SST anomaly — of which the ACE-congruent anomaly is a scaled version — forces a PNA-like wave train and, in the significant full-El-Niño run, a positive Beaufort surface-energy anomaly that favours ice melt (Figs. 4, 5); as with skin temperature, the corresponding flux response in the weaker ACE run is of the expected sign but not individually distinguishable from internal variability. The negative-amplitude runs reverse the sign of the pattern, consistent with a near-linear dependence on the imposed forcing. These experiments provide direct dynamical support for the downstream half of the proposed chain — that the ACE-congruent tropical-Pacific SST anomaly forces a PNA-like wave train and a positive Beaufort surface energy anomaly. A Takaya–Nakamura (2001) wave-activity-flux diagnosis of the CAM5 response confirms coherent poleward wave-energy propagation from the forced North Pacific source into the Arctic-Pacific gateway (60° – 75° N) in both winter and spring. Notably, from winter to spring the mid-latitude source weakens faster than the flux entering the gateway (the North Pacific source flux drops by a factor of ~ 6 for the full El Niño run while the gateway flux falls only by a factor of ~ 2), indicating that the tropical–Beaufort wave pathway persists into the spring melt season even as the underlying SST forcing decays — consistent with the DJF-to-MAM persistence of the response noted above. We emphasize two limitations of the design. First, the forcing is imposed as a tropical-Pacific SST anomaly (with no anomaly over the WNP), so these

experiments test the ENSO to Arctic pathway and do not simulate tropical cyclones forcing ENSO. Second, because the ACE-regressed and N3.4-regressed SST anomalies are nearly co-linear in pattern, the ACE run should be read as testing the response to *the fraction of the El Niño SST anomaly associated with ACE*, rather than to a physically distinct “ACE” SST fingerprint.

CAM5 winter (DJF) response to imposed tropical-Pacific SST forcing (run - control)

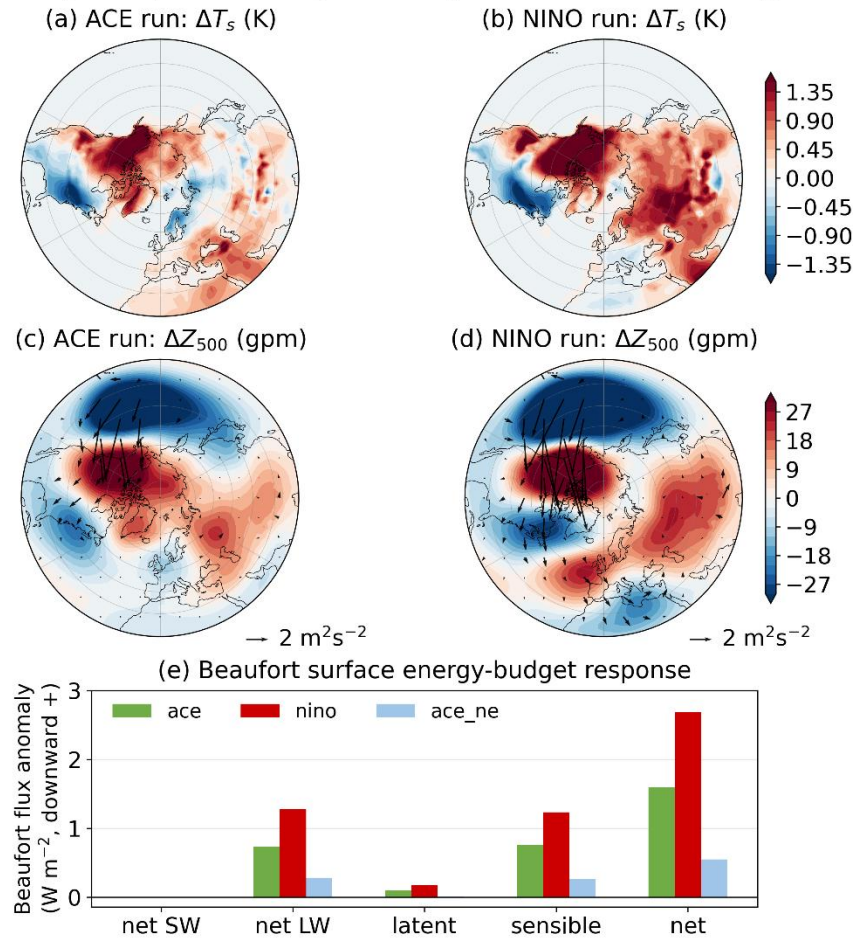


Figure 4. (CAM5 winter/DJF response, the El Niño peak season.) Northern-Hemisphere and Beaufort response (experiment minus control), averaged over December–February. The ACE (N3.4) run imposes the tropical-Pacific SST anomaly regressed onto the ASO ACE (NDJ Niño-3.4) index; the two patterns are nearly identical in structure, with the ACE

anomaly $\approx 70\%$ of the N3.4 amplitude. (a, b) skin-temperature anomaly ΔT_s (K) for the positive-ACE and positive-N3.4 runs; (c, d) 500-hPa geopotential-height anomaly ΔZ_{500} (gpm) showing the PNA-like wave train, with Takaya–Nakamura (2001) horizontal wave-activity-flux vectors (black arrows; scale at lower right) indicating the direction of stationary wave-energy propagation; (e) surface energy-budget decomposition (net shortwave, net longwave, latent, sensible, net downward flux) over the Beaufort box (72–80°N, 200–230°E) for the ACE, N3.4 and negative-amplitude (ACE_ne) runs.

The teleconnection follows a physically coherent seasonal evolution. In the El Niño peak season (December–February, Fig. 4), the imposed tropical-Pacific SST anomaly forces a well-developed PNA-like wave train and a strong positive surface energy flux into the Beaufort Sea (net $\approx +1.6 \text{ W m}^{-2}$ for the ACE run, +2.7 for the full El Niño run), dominated by downward longwave radiation and sensible heat under polar night. By the following spring (March–May, Fig. 5) — the season of the sea-ice response analysed above — the wave train and surface energy anomaly persist but weaken, and net shortwave radiation begins to contribute as polar night ends. In both seasons the ACE run reproduces the sign and spatial structure of the full El Niño response at reduced amplitude, and the negative-amplitude runs reverse it, confirming a near-linear dependence on the imposed tropical-Pacific SST forcing.

CAM5 spring (MAM) response to imposed tropical-Pacific SST forcing (run - control)

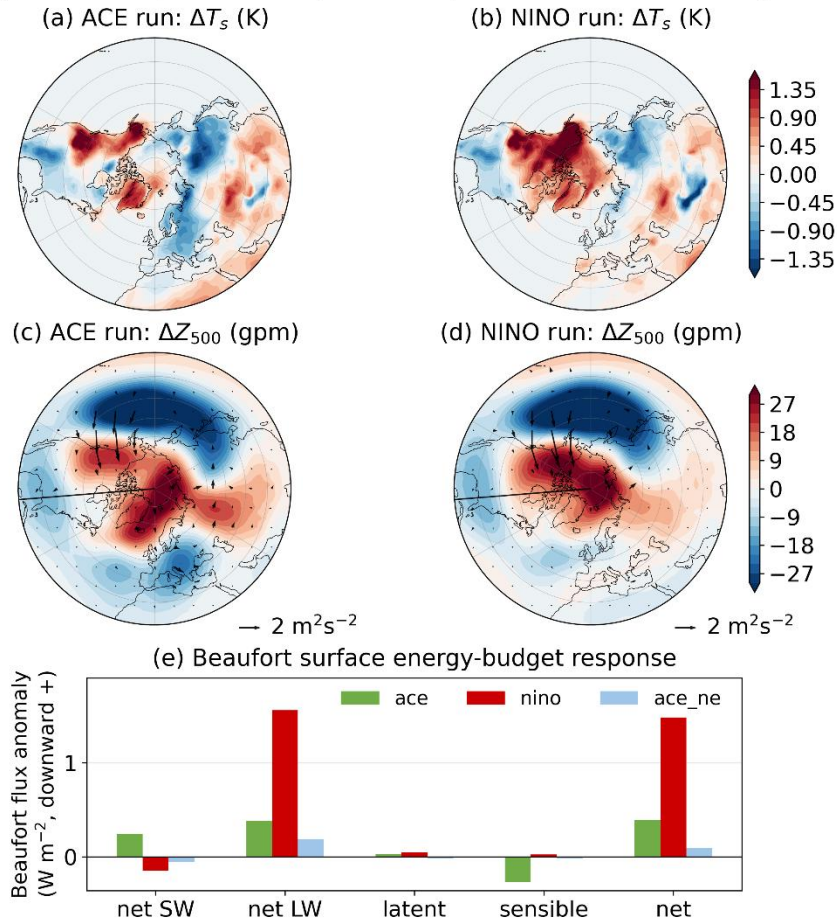


Figure 5. (CAM5 spring/MAM response, matching the sea-ice target season.) As Fig. 4, but averaged over March–May. The wave train and positive Beaufort surface energy anomaly persist from winter but weaken, with net shortwave beginning to contribute as polar night ends.

5 Sea-ice prediction based on the preceding ACE index

Skillful seasonal prediction of Arctic sea ice remains challenging, and existing skill derives largely from coupled dynamical forecast systems and ocean/ice initial conditions (Wang et al. 2013; Bushuk et al. 2022). A remote tropical precursor that leads the sea ice

by several seasons would therefore be of practical interest. Because the preceding ASO ACE leads the following spring–early-summer Beaufort SIC by roughly three seasons (Section 3), we test whether it provides genuine predictive information. To avoid overfitting on a short (≈ 50 -year) record, skill is assessed by **leave-one-out cross-validation (LOO-CV)** rather than by a single fixed fit/hindcast split. Detrending and predictor standardization are performed within each training fold to avoid information leakage; the cross-validated skills reported below are, if anything, slightly conservative relative to a full-record-detrended estimate.

We distinguish two forecast configurations by their issue date. In the **long-lead configuration**, issued at the end of the ASO tropical-cyclone season, only the ASO ACE index (and, equivalently, the developing-ENSO signal it carries) is available — the following winter and spring ice states do not yet exist. Used alone in this configuration, the ASO ACE index predicts the following detrended March–May (MAM) Beaufort SIC anomaly with a cross-validated correlation of $r \approx 0.31$ (sign-agreement 57%); the NDJ Niño-3.4 index gives comparable skill ($r \approx 0.29$), consistent with the two indices sharing the ENSO-bridged signal of Section 3.

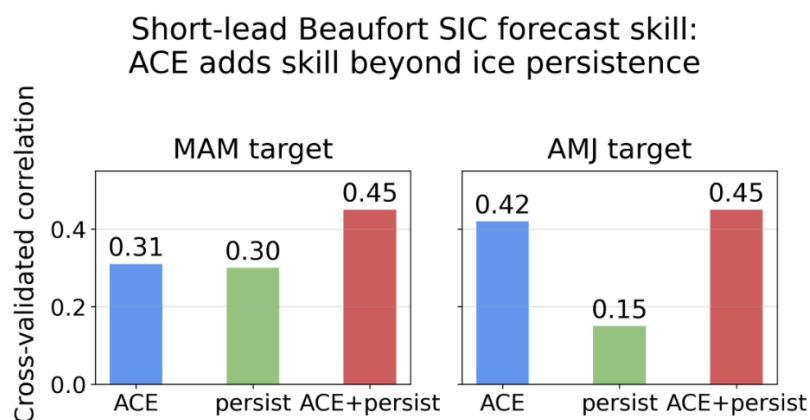
In the **short-lead configuration**, issued after the following winter (e.g. in February), the just-observed winter sea-ice state becomes available as an additional predictor. Because ACE is essentially independent of the Beaufort sea-ice state (its correlation with the preceding-winter, JFM, SIC is only 0.03), we combine the two as a simple two-predictor regression (ACE + preceding-winter SIC). This raises the cross-validated skill to $r \approx 0.45$ (95% bootstrap CI [0.30, 0.60], resampling the out-of-fold prediction–observation pairs; sign-agreement $\approx 74\%$), with the ACE index contributing a

point-estimate increment of +0.15 beyond persistence (though with a wide confidence interval; see below). We stress that these two numbers describe different forecasts: the $r \approx 0.31$ skill is available at $\approx 7\text{--}10$ months lead, whereas the $r \approx 0.45$ combined skill is a shorter-lead ($\approx 1\text{--}4$ month) forecast that additionally exploits winter ice persistence. One caveat attaches to the combined skill: the JFM persistence predictor and the MAM target share the month of March, so the MAM persistence — and hence the MAM combined skill — is modestly inflated by this overlap; the AMJ target, on which the finding that ACE is most valuable where persistence fails principally rests, shares no month with the JFM predictor and is unaffected.

The relative contributions vary with target season in a physically interpretable way. For spring (MAM) ice, persistence is comparatively strong (LOO-CV $r \approx 0.30$) and ACE adds a modest increment. For early-summer (AMJ) ice, persistence weakens sharply ($r \approx 0.15$) as melt erodes the winter memory, and the ACE/ENSO dynamical signal then supplies most of the skill (increment $\approx +0.30$). In other words, the ACE precursor is most valuable in the melt season, precisely where sea-ice persistence fails.

Two further points reinforce the practical value of this remote precursor. First, its lead time is genuinely long: issued at the end of the ASO tropical-cyclone season, an ACE-based forecast anticipates the following spring–early-summer ice by roughly seven to ten months. At that lead, the local sea-ice state carries essentially no linear memory — the preceding autumn (SON) Beaufort ice concentration is uncorrelated with the following-year MAM/AMJ ice ($r \approx -0.05$, not significant), because the intervening freeze–melt cycle erases it. The ACE index therefore supplies predictive information in a season and at a lead where conventional persistence provides none. Second, there is some indication that the

493 ACE signal is not merely a proxy for winter ENSO: its partial correlation with the
 494 following early-summer (AMJ/MJJ) ice, after removing the NDJ Niño-3.4 index, is $r \approx$
 495 -0.28 ($p \approx 0.05$), and adding ACE to a predictor set already containing both the winter
 496 ENSO index and the winter ice state raises the cross-validated skill by a point estimate of
 497 ≈ 0.08 (AMJ/MJJ). We stress, however, that on the short observational record ($n \approx 54$) this
 498 incremental skill is not statistically robust: a bootstrap 95% confidence interval on the
 499 increment beyond ENSO-plus-persistence spans zero (AMJ $+0.07$, 95% CI $[-0.06, +0.20]$;
 500 MAM $+0.02$, 95% CI $[-0.04, +0.09]$). The increment beyond persistence alone is larger
 501 ($+0.15$ for MAM, $+0.30$ for AMJ) but likewise has a wide interval reaching (or nearly
 502 reaching) zero. We therefore conclude, conservatively, that the preceding ACE index
 503 provides a modest cross-validated source of predictive information for Beaufort spring–
 504 summer sea ice — comparable to the winter ENSO index and complementary to sea-ice
 505 persistence, and most useful at long lead in the melt season where persistence fails — but
 506 that its independent contribution beyond ENSO, while consistently positive in the point
 507 estimates, cannot be established as significant on the present record and warrants
 508 confirmation with longer or independent datasets.



509

Figure 6. Leave-one-out cross-validated forecast skill (correlation between predicted and observed detrended Beaufort SIC anomaly, 1971–2024) for two target seasons (MAM, AMJ) and three predictor sets: ACE alone, preceding-winter SIC persistence alone, and their combination. Fold-wise detrending is applied within each cross-validation fold to avoid leakage. ACE adds skill beyond ice persistence in both seasons.

6 Conclusions and discussion

This study investigated a persistent, three-season teleconnection by which tropical-cyclone activity over the western North Pacific influences Beaufort Sea ice. Using the ACE index together with regression-based signal decomposition, a perturbation hypsometric diagnosis, Rossby-wave ray tracing, and a surface-flux decomposition, we assembled the following chain of reasoning. (i) Enhanced preceding ASO ACE over the WNP is associated with an intensified following-winter El Niño — an amplification attributed to the tropical cyclones by Wang et al. (2019), though not independently established here (Section 2.3) — and partial-correlation analysis shows that the NDJ N3.4 signal bridges the ACE–Beaufort SIC relationship (Section 3). (ii) The intensified tropical-Pacific convective heating excites a stationary Rossby wave train and a PNA-like pattern that carries energy poleward to the Beaufort Sea, producing anomalies of atmospheric thickness, near-surface temperature, humidity, and water-vapor flux (Section 3, Fig. 2). (iii) These circulation anomalies reorganize the Beaufort surface energy budget: net longwave radiation dominates during and just after polar night, whereas net shortwave radiation — amplified by the sea-ice–albedo feedback — dominates from spring into

summer, while latent and sensible heat fluxes remain secondary (Section 4). (iv) The ACE-related and N3.4-related pathways produce broadly similar surface-energy changes but differ in the relative roles of longwave and shortwave radiation (Section 4). (v) In a cross-validated statistical prediction framework, the preceding ACE index provides a modest source of predictive skill for the following spring–summer Beaufort sea ice (leave-one-out $r \approx 0.45$ combined with winter ice persistence), most valuable in the melt season and at long lead where local sea-ice persistence carries no memory; its skill beyond the winter Niño-3.4 index is positive but not statistically significant on the present record (Section 5).

A schematic summary of this mechanism is given in Fig. 7.

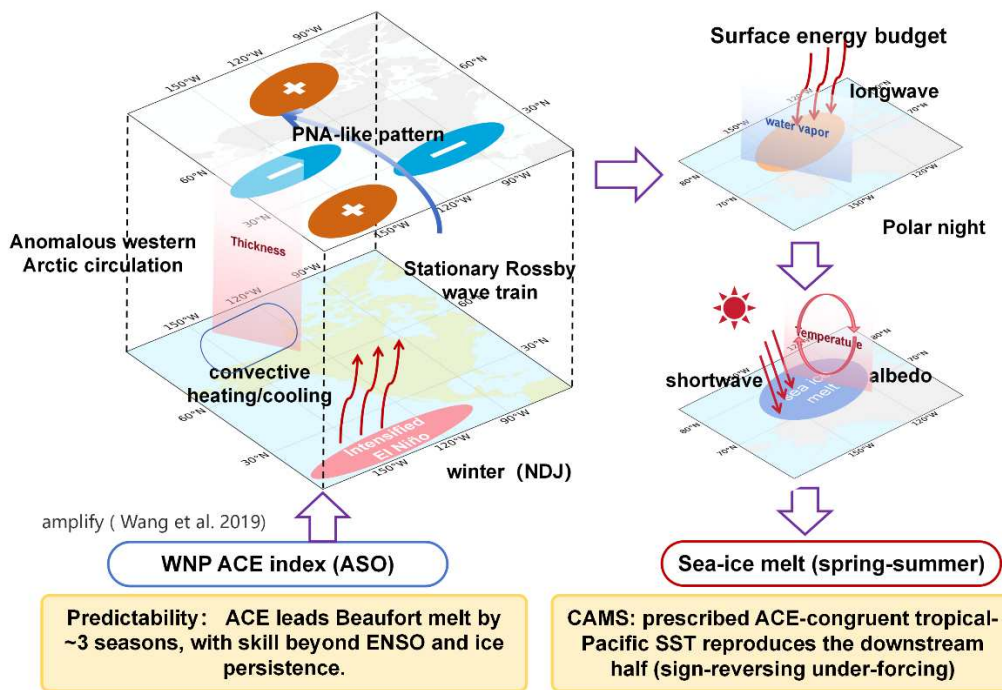


Figure 7. Schematic of the proposed mechanism. Enhanced autumn (ASO) western North Pacific tropical-cyclone activity, quantified by the ACE index, is associated with an intensified following-winter (NDJ) El Niño (an amplification attributed to the cyclones by

Wang et al. 2019, not established here); the associated tropical-Pacific convective heating excites a stationary Rossby wave train and a PNA-like pattern that propagates energy into the western Arctic, reorganizing the Beaufort surface energy budget (net longwave dominant during and just after polar night; net shortwave and the ice–albedo feedback dominant from spring into summer) and promoting spring–summer sea-ice melt. The CAM5 experiments confirm the downstream (tropical SST to Arctic) half of this chain.

These results extend the framework of Wang et al. (2019) and Ye et al. (2024) by following the ACE–ENSO signal into the summer melt season and by providing a term-by-term surface-energy diagnosis of how the signal reaches the Beaufort Sea. It is important to distinguish the present pathway from the apparently opposite-sign result of Zeng et al. (2024), who reported that WNP TC activity favors an *increase* in Arctic sea ice. The two are not contradictory: they operate on different timescales, in different regions, and through different bridges. Zeng et al. (2024) describe a *direct, synoptic* teleconnection (≈ 30 -day lag) in which TC-excited Rossby waves drive an anomalous *cyclonic* circulation over the Arctic-Pacific sector that deepens the polar vortex, cools and dries the lower Arctic, *reduces* downward longwave radiation, and thereby increases sea ice near the Fram Strait. The present study, by contrast, describes an *indirect, seasonal* pathway (≈ 6 –9-month lag) in which the ACE signal is first stored in and amplified by El Niño, whose tropical-Pacific heating then forces a PNA-like response that *warms* and moistens the Beaufort sector, *increases* downward longwave radiation, and promotes sea-ice melt. The two mechanisms can therefore coexist, with the direct synoptic pathway dominating at short lead times and high-latitude Pacific longitudes and the ENSO-bridged pathway dominating

at seasonal lead times over the Beaufort Sea. Quantifying their relative contributions is an important direction for future work.

Several limitations should be noted. First, the causal chain rests on the premise that WNP ACE *intensifies* El Niño (Wang et al. 2019). This direction of causality is the load-bearing assumption of the study, yet it remains debated: the dominant view attributes ENSO growth to large-scale westerly wind bursts and ocean dynamics (e.g. Hu and Fedorov 2019), of which organized TC activity is only one possible contributor, and the more established relationship runs from ENSO to WNP TC activity (Camargo and Sobel 2005). Because the present analysis is correlative and built on regression-based signal decomposition, it establishes a consistent, physically plausible pathway but cannot by itself prove that the ACE-related portion of the El Niño signal is causal; targeted modeling experiments (e.g. TC-permitting or TC-suppressed simulations) would be needed to close this gap. Second, the analysis rests on a relatively short observational record (about 50 years, 1970–2024), which limits the number of strong-ACE composite members and the robustness of the significance tests, particularly given the strong autocorrelation of the smoothed series. Third, the dynamical and energy-budget diagnostics rely on a single reanalysis (ERA5) and a single SST product (HadISST); the sensitivity of the results to reanalysis choice has not been assessed. Fourth, although the cross-validated prediction skill in Section 5 is statistically robust, it is modest (leave-one-out $r \approx 0.45$), so the ACE index should be regarded as a complementary source of predictive information — valuable chiefly at long lead and in the melt season, where conventional predictors fail — rather than as a stand-alone quantitative predictor. Future work could test the mechanism in coupled-model and large-ensemble experiments, evaluate the robustness across multiple

reanalyses, and separate the ACE-induced component of ENSO from internal ENSO variability more cleanly. In addition, tropical cyclones in the eastern Pacific may exert similar influences on Beaufort Sea ice, and whether their ACE provides complementary predictive information warrants further investigation.

Data Availability

ERA5 reanalysis data are available from the Copernicus Climate Change Service (C3S) Climate Data Store (<https://cds.climate.copernicus.eu>). IBTrACS tropical-cyclone best-track data (version 4) are available from NOAA NCEI (<https://doi.org/10.25921/82ty-9e16>). HadISST sea surface temperature and sea-ice concentration data are available from the UK Met Office Hadley Centre (<https://www.metoffice.gov.uk/hadobs/hadisst/>). Oceanic Niño Index (Niño-3.4) values are available from NOAA (https://origin.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_v5.php). The Community Atmosphere Model version 5 (CAM5) is distributed as part of the Community Earth System Model (CESM) by NCAR (<https://www.cesm.ucar.edu>). The analysis code and derived data that support the findings of this study are available from the corresponding author upon reasonable request.

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Author Contributions

Dingjia Ye and Yanping Li conceived the main idea of this study. All authors participated in data collection and analysis, as well as graphical analysis and visualization. The initial draft of the manuscript was written by Dingjia Ye and Yanping Li. All authors provided comments on previous versions of the manuscript. All authors contributed to reviewing and revising the manuscript.

Declarations

Conflict of interest The authors have no relevant financial or non-financial interests to disclose.

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