

1 **Air-Ice-Ocean Coupling During a Strong Mid-Winter Cyclone, Part 1: Observing**
2 **Coupled Dynamic Interactions Across Scales**

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12 **Key Points:**

- 13 • A pair of cyclones crossed the Multidisciplinary drifting Observatory for the Study of the
14 Arctic Climate (MOSAIC) in midwinter 2020
- 15 • Detailed, multi-platform observations enable characterization of coupled air-ice-ocean
16 interactions during the passage of these cyclones
- 17 • The development of a low-level atmospheric jet is a key factor in the spatially and
18 temporally varying sea ice-ocean response to cyclones
19

20 Abstract

21 Arctic cyclones are key drivers of sea ice and ocean variability. During the 2019-2020
22 Multidisciplinary drifting Observatory for the Study of Arctic Climate (MOSAiC) expedition,
23 joint observations of the coupled air-ice-ocean system were collected at multiple spatial scales.
24 Here, we present observations of a pair of strong mid-winter cyclones that impacted the
25 MOSAiC site as it drifted in the central Arctic pack ice, with analytic emphasis on the second
26 cyclone. The sea ice dynamical response showed spatial structure at the scale of the evolving
27 atmospheric wind field. Internal ice stress and the ocean stress play significant roles, resulting in
28 timing offsets between the atmospheric forcing and the ice response and post-cyclone inertial
29 ringing in the ice and ocean. A structured response of sea ice motion and deformation to cyclone
30 passage is seen, and the consequent ice motion then forces the upper ocean currents through
31 frictional drag. The strongest impacts to the sea ice and ocean from the passing cyclone occur as
32 a result of the surface impacts of a strong atmospheric low-level jet (LLJ) behind the trailing cold
33 front. Impacts of the cyclone are prolonged through the coupled ice-ocean inertial response. The
34 local impacts of the approximately 120 km wide LLJ occur over a 12 hour period or less and at
35 scales of a kilometer to a few tens of kilometers, meaning that these impacts occur at smaller
36 spatial scales and faster time scales than many satellite observations and coupled Earth system
37 models can resolve.

38 Plain Language Summary

39 Arctic winter cyclones are an important part of the Arctic climate system. Yet, due to sparse
40 observations, processes of the coupled sea ice-ocean response to cyclones are not fully
41 understood. During the 2019-2020 Multidisciplinary drifting Observatory for the Study of Arctic
42 Climate (MOSAiC) expedition, observations of the atmosphere, sea ice, and ocean were
43 collected at a range of spatial scales. Here, we describe the atmospheric structure and coupled
44 ice-ocean response to a strong winter cyclone using data from surface weather stations, weather
45 balloons, radar, and a weather model. We then describe the sea ice motion using a large set of
46 GPS buoys and ice radar images. Finally, we examine the upper ocean currents and structure
47 using ocean buoy data. The most important part of the storm structure for the sea ice is the
48 development of an atmospheric low-level jet (LLJ), a narrow region of fast-moving air that
49 eventually circles around the storm. The sudden change in ice drift speed at the time that the LLJ

50 passes overhead enhances motion of the ice and ocean. Periodic currents in the ocean initiated by
 51 the sudden wind change of the LLJ continue for days following the passage of the storm,
 52 prolonging its effects.

53 **1 Introduction**

54 The physical environment in the Central Arctic consists of dynamically and thermodynamically
 55 coupled processes between the atmosphere, ice, and upper ocean (Brenner et al., 2023; Deser et
 56 al., 2015; Persson et al., 2017; Petty et al., 2016; Webster et al., 2019). Sea ice, and its
 57 accompanying snow cover, regulates the linkage between atmosphere and ocean through
 58 dynamics (lead opening and closing, evolving roughness of the air-ice and ice-ocean interfaces)
 59 and through thermodynamics as the ice and snow packs grow and melt (Maykut, 1982; Overland,
 60 1985; Persson, 2002, 2012; Pinto et al., 2003; Ruffieux et al., 1995; von Albedyll et al., 2022). In
 61 turn, the stability of the atmospheric and ocean boundary layers governs the evolution of
 62 turbulent eddies, affecting the magnitude of turbulent fluxes of heat and momentum (Andreas et
 63 al., 2010a, b; Grachev et al., 2007; Lüpkes et al., 2008; Lüpkes & Gryanik, 2015; Taylor et al.,
 64 2018).

65 Arctic cyclones play a large role in this air-ice-ocean turbulent exchange. The large-scale
 66 pressure and mass fields of a cyclone produce strong winds near the central low and in air-mass
 67 transport belts along fronts. Therefore, cyclone passage results in a pulse of momentum, heat,
 68 and moisture into the ice-ocean system. They represent major sources of poleward heat and
 69 moisture transport during Arctic winter (Fearon et al., 2021) and impact the surface energy
 70 budget, ice growth, and even spring melt onset (Persson, 2012; Persson et al., 2017). Cyclone
 71 passage is often accompanied by strong sea ice deformation (Itkin et al., 2017; Lindsay, 2002;
 72 Oikkonen et al., 2017) and enhanced ocean mixing (Meyer et al., 2017a, b). Cyclone impacts on
 73 sea ice depend on time of year, cyclone strength and evolutionary stage, location within the
 74 Arctic, location relative to the ice edge and coast, and the sea ice state (Aue et al., 2022).

75 The direct dynamic impacts of cyclones on the sea ice momentum equation, expressed in
 76 Equation 1 (e.g., Hibler, 1979; Hunke et al., 2015) are transferred through the atmospheric stress
 77 term, $\boldsymbol{\tau}_a$:

$$78 \quad m \frac{D\mathbf{u}}{Dt} = -mf\mathbf{k} \times \mathbf{u} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_o - mg \nabla H + \nabla \cdot \boldsymbol{\sigma} \quad (1)$$

79 The left side of (1) is the rate of change of the ice momentum with approximately constant mass
80 m (snow and sea ice mass per unit area), where \mathbf{u} is the sea ice velocity. The sum of forces on
81 the right-hand-side terms consists of the stresses on the ice due to the Coriolis force, where f is
82 the Coriolis parameter, the atmosphere and ocean stress vectors $\boldsymbol{\tau}_a$ and $\boldsymbol{\tau}_o$, the effect of gravity
83 down the slope of the ocean surface, and the divergence of the internal stress tensor. The last
84 term represents energy loss due to friction between the floes and conversion of kinetic energy to
85 potential energy, parameterized in terms of bulk and shear viscosities and ice strength. The
86 coupled inertial response following the storm passage can prolong its dynamic effects (Haller et
87 al., 2014).

88 The structure of the wind field within a cyclone imparts spatial gradients in the surface stresses,
89 resulting in gradients of ice acceleration. As a result, the thermodynamic and dynamic sea ice
90 response varies relative to the position of the low pressure center and the orientation of the storm
91 track (e.g., Brümmer, 2003; Haapala et al., 2005; Kriegsmann & Brümmer, 2014; Overland &
92 Pease, 1982). Composite analysis based on reanalysis and satellite observations demonstrate that
93 sea ice impacts have spatial structure, with dependence on distance from the storm center
94 (Kriegsmann & Brümmer, 2014) and position relative to the storm track (Clancy et al., 2022).
95 However, estimates of cyclone structure and impacts based on composite analysis are sensitive to
96 choices made in cyclone identification (Rae et al., 2017) and to the choice of reanalysis (Vessey
97 et al., 2020). Differences in cyclone properties between reanalysis composites can arise from
98 uncertainty in the physics of Arctic cyclones, differences in model implementation (including
99 choice of parametrization schemes and model resolution), and the limited long-term in situ
100 observations in the central Arctic, particularly joint observations of atmosphere, sea ice, and
101 ocean.

102 Observations of the coupled air-ice-ocean system with the ability to resolve mesoscale cyclone
103 features, including fronts, are extremely rare. Thermodynamic air-ice-ocean interactions for
104 cyclones sampled during the Surface Heat and Energy Budget of the Arctic expedition (SHEBA;
105 Persson, 2002; Uttal et al., 2002) have been analyzed at least in part in numerous studies (e.g.,
106 Lindsay, 2002; Persson, 2012; Persson et al., 2017; Richter-Menge et al., 2001; Shaw et al.,
107 2009), providing case studies and seasonal and annual analysis. Both Lindsay (2002) and
108 Richter-Menge et al., (2001) identify periods of enhanced mid-winter sea ice deformation that
109 coincided with significant cyclone activity; however, the sea ice deformation observations lack

110 sufficient resolution to examine air-ice dynamic coupling in detail. Measurements of sea ice
111 motion and deformation show patterns related to the storm structure (Brümmer et al., 2008;
112 Haller et al., 2014), with ice tending to diverge on average as the cyclone passes. The location of
113 the ice edge and the local history of deformation is an important factor (Oikkonen et al., 2017).

114 The ocean response to wind forcing is strongly modulated by seasonal changes in ice thickness,
115 roughness, and concentration (Gallaher et al., 2016; McPhee, 2002, 2008; Meyer et al., 2017a;
116 Shaw et al., 2009; Stanton et al., 2012; Yang, 2004). Cyclones, and the strong gradients in winds
117 associated with them, result in changes in momentum transfer to the ocean that can excite inertial
118 oscillations in the ocean and ice (Brümmer & Hoerber, 1999; Hunkins, 1967), where the ice and
119 ocean move together in an inertial ringing (Toole et al., 2010). The presence of ice can damp the
120 ocean response (Brenner et al., 2023; Rainville & Woodgate, 2009), however inertial oscillations
121 are observed in all seasons, including under consolidated winter ice pack (Martini et al., 2014).

122 This momentum transfer and the inertial motion enhances mixing in the upper ocean and may
123 also excite internal waves that enhance deeper mixing (McPhee & Kantha, 1989). High wind
124 speeds over sea ice have been observed to produce increased ocean friction velocity (Shaw et al.,
125 2009) and enhanced turbulent dissipation in the upper ocean (Meyer et al., 2017a). The winter
126 ice cover impedes momentum transfer from the wind to the ocean, reduces the inertial response
127 of the ocean, and likely sets the shallow winter mixed layer depth in parts of the Arctic Ocean
128 (Toole et al., 2010). Buoy observations of sea ice drift suggest that the inertial response of the ice
129 has been increasing (Gimbert et al., 2012; Yuan et al., 2022). It has been hypothesized that an
130 increase in sea ice inertial response may arise due to thinning of the ice pack (Gimbert et al.,
131 2012; Kwok et al., 2013) as well as increased cyclonic activity (Roberts et al., 2015).

132 To date, the full momentum transfer from wind, through ice to the ocean has not been observed
133 directly on the temporal and spatial scales that clearly define the roles of the spatial structure of a
134 cyclone for the associated ice and ocean response. To that end, we consider the detailed
135 observations of the coupled-air-ice-ocean system obtained during the Multidisciplinary drifting
136 Observatory for the Study of Arctic Climate (MOSAiC) expedition (Shupe et al., 2020; Shupe &
137 Rex, 2022). This study examines the relative roles of the atmospheric stress, ocean stress
138 (shearing between ice motion and upper-ocean currents), and the internal ice stress (via
139 consideration of sea ice deformation) in the momentum balance from MOSAiC observations
140 during the passage of two atmospheric cyclones that traversed the study area between 30 January

141 and 2 February, 2020. While there were numerous cyclones during the MOSAiC year (Rinke et
142 al., 2021), these two cyclones (especially the second cyclone) are of particular interest. They
143 were intense, with high wind speed and low minimum sea level pressure. The sea ice response
144 included the fastest winter drift speeds in the MOSAiC drift and strong deformation of the ice
145 pack. The proximity of the storm track to the MOSAiC observatory allowed detailed
146 observations of the cyclone development and ice-ocean response. Furthermore, the cyclones
147 occurred during the consolidated ice season in the high Arctic, when the internal ice stress term
148 is expected to be an important part of the response.

149 The study highlights the atmospheric features producing the atmospheric stress characteristics,
150 and the impacts of these stress terms on the sea ice and ocean motion. While the atmospheric
151 stress is generally regarded as the primary forcing mechanism for ice motion, it is shown that
152 both the internal ice stress and the ocean stress play significant roles in changing the typical air-
153 ice interaction characteristics, including producing timing offsets between the atmospheric
154 forcing and the ice response and producing post-cyclone inertial “ringing” responses in the ice
155 and ocean. The MOSAiC observations and additional data are described in section 2. Sections 3-
156 5 describe the observations of atmosphere, sea ice, and ocean, respectively. Discussion and
157 conclusions follow in Section 6.

158 **2 Data and methods**

159 The MOSAiC Central Observatory (CO) and its surrounding distributed network (DN) of
160 automated observational platforms and buoys were deployed in residual ice north of the Laptev
161 Sea in early October 2019, and drifted across the Central Arctic during the subsequent winter,
162 entering the Fram Strait in June 2020 (Krumpfen et al., 2020). Maps showing the track of the
163 drifting station and more details of the atmospheric, ice, ocean and DN observations along this
164 drift track can be found in a series of MOSAiC overview (Nicolaus et al., 2022; Rabe et al.,
165 2022, 2024; Shupe et al., 2022) and domain-specific (e.g., Fer et al., 2022; Krumpfen et al., 2021;
166 Peng et al., 2023; von Albedyll et al., 2022; Watkins et al., 2023) publications. Figure 1 shows a
167 map of the relative positions of the CO and the DN sites on Jan 31, 2020, at which time the CO
168 was located at 87.5° N, 96.0° E (275 km from the North Pole).

169 2.1 Atmospheric observations

170 Atmospheric observations used in this study were made at the CO (both on board the *R/V*
171 *Polarstern*, and at the “Met City” site located on the ice approximately 400 m from the ship), and
172 at the three “L” sites located 10-25 km from the ship (Figure 1). Key measurements from the
173 *R/V Polarstern* include the 6-hourly rawinsondes providing profiles of temperature, humidity,
174 and horizontal winds, and the vertically-pointing Ka-band radar providing profiles of radar
175 reflectivity and radial velocity. The 30-s radar data profiles were averaged to 10-min time
176 intervals for this study. A DOE/ARM scanning Ka-band radar provided volumes of radar
177 reflectivity and radial velocity approximately every 12 minutes, providing data for plane-parallel
178 indicator (PPI) displays characterizing clouds and precipitation. Post-field program reflectivity
179 calibrations were applied. Analyses of fronts and mesoscale features in the time-height cross
180 sections and horizontal displays relied on standard subjective analyses of thermodynamic (e.g.,
181 temperature, virtual potential temperature (θ_v), equivalent potential temperature (θ_e)), kinematic
182 (e.g., wind speed and direction), and radar reflectivity observations, not all of which are shown.
183 Changes in θ_e and wind direction and minima in SLP were key markers for determining frontal
184 boundaries. The Arctic inversion (AI) was defined as the height of the maximum temperature in
185 each sounding, and varied distinctly as synoptic conditions changed. Surface-based layers of
186 constant θ_v defined the surface mixed-layer (SML) depth for each sounding.

187 Sonic anemometers and basic meteorology sensors at Met City provided time series of
188 temperature, humidity, winds, mean sea level pressure (SLP), and turbulence (including
189 momentum flux) at 3 different levels (nominally 2, 6, and 10 m) and 4-component broadband
190 radiative fluxes at ~2.5 m height. Atmospheric Surface Flux Stations (ASFS) located at the three
191 L-sites provided measurements of temperature, humidity, pressure, and 4-component broadband
192 radiative fluxes at ~2 m above the sea ice, and winds and turbulence (including momentum flux)
193 at 3.8 m above the ice. The ASFS and Met City data used in this study are 10 minute average
194 values. Unless otherwise stated, the Met City wind and turbulence data shown represents the 10
195 m height while those at the ASFS represent the 3.8 m height.

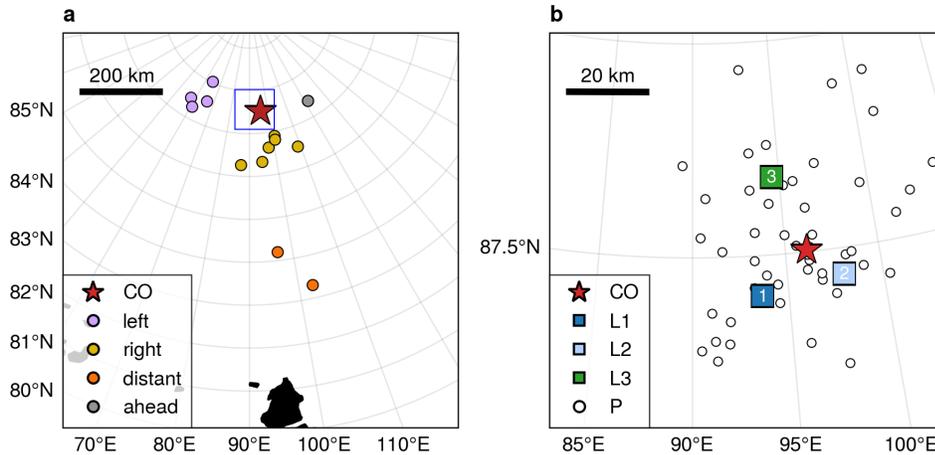
196 Atmospheric stress was obtained through covariance calculations using the 10 Hz three-
197 component (u, v, w) measurements from the sonic anemometers at Met City and the ASFS. First,
198 the earth coordinate system is rotated into streamwise coordinates through a double rotation

199 (Kaimal and Finnigan, 1994). The 10-min values of friction velocity $u_a^* = (u_s'w')^{1/2}$ were then
 200 obtained from the integration of the cross-spectral density for a 13.65 minute window centered
 201 on the 10 minute period. Here, u_s' and w' are perturbation values of the streamwise and vertical
 202 wind speeds, respectively. The observed atmospheric stress is then calculated by $\tau_a = \rho_a u_a^{*2}$,
 203 where ρ_a is the atmospheric density. More detailed descriptions of the data processing,
 204 turbulence calculations and the atmospheric measurements on the *R/V Polarstern*, at Met City,
 205 and at the ASFS sites are provided by Shupe et al. (2022) and Cox et al. (2023).

206 Time series of low-level atmospheric divergence are calculated from the winds at the three L-
 207 sites using the assumption that the winds vary linearly between the three sites. With this
 208 assumption, the area-averaged low-level atmospheric divergence div_a can be calculated using the
 209 area-normalized divergence theorem and by integrating the winds normal to the sides of the
 210 polygon such that

$$211 \quad \text{div}_a \approx \frac{1}{A} \left[\sum_{i=1}^{n_s} (\bar{u}_i dy_i - \bar{v}_i dx_i) \right] \quad (2)$$

212
 213 where $n_s = 3$ is the number of sides, \bar{u}_i, \bar{v}_i are the mean u and v wind components on side i , and
 214 (dx_i, dy_i) are the component lengths of each side i , and A is the area of the polygon. Because this
 215 calculation is sensitive to errors in the installation and manual orientation of the sonic
 216 anemometers, we assume that the long-term mean divergence between ASFS sensor alignments
 217 is zero, and subtract this mean value from each value in the time series. For this case, the long-



218 term mean divergence was $-1.1 \times 10^{-5} \text{s}^{-1}$ calculated between 30 November, 2019, and 5
219 February, 2020. Hence, magnitudes of div_a larger than $2 \times 10^{-5} \text{s}^{-1}$ are likely significant.

220

221 **Figure 1.** MOSAiC domain and instrument locations on 1 February 2020 at 00:00 UTC. Shown
222 are a) the extended DN and b) the DN, defined as buoys within 60 km of the CO. The extent of
223 panel b is shown by the open blue square in panel a. The Central Observatory (red star labeled
224 CO) and the 3 “L-sites” with the ASFS, SIMB and AOFB (squares, right hand panel) measure
225 complete atmospheric, ice and upper-ocean parameters. The GPS ice buoys (circles) measure
226 position, and their colors in panel a correspond to groups defined and highlighted in Figures 7
227 and 9.

228 2.2 ERA5 atmospheric reanalysis

229 To obtain additional spatial atmospheric information, we supplement the atmospheric
230 observations with 0.25° resolution data from the fifth-generation European Center for Medium-
231 range Weather Forecasting reanalysis (ERA5; Hersbach et al., 2020) obtained from the
232 Copernicus Data Store (Hersbach et al., 2023b, 2023a). Prior to analysis, the data was
233 reprojected on a regular 25 km north polar stereographic grid with central longitude of 90° . The
234 ERA5 reanalysis performs well relative to other reanalyses in the Arctic domain (Graham et al.,
235 2019a, b). ERA5 is known to have a surface warm bias in the Arctic (C. Wang et al., 2019; Yu et
236 al., 2021), and ERA5 low-level jets are slightly weaker and slightly elevated (López-García et
237 al., 2022). Here, we use ERA5 mean sea level pressure, 10-m winds, and 925 hPa temperature
238 and humidity, and 950 hPa winds. Rawinsonde and surface observations from the MOSAiC
239 central observatory were assimilated by ERA5, as were surface measurements of temperature
240 and air pressure from a few buoys from the MOSAiC DN. The ERA5 4D-var assimilation
241 method uses a centered 12-h window, allowing impacts of observations to spread spatially and
242 temporally. Hence, the ERA5 atmospheric structure should well represent the true atmospheric
243 structure near the CO (as demonstrated for LLJs by López-García et al., 2022) and likely is the
244 best available estimate of storm structure further away from the CO.

245 2.3 Sea ice observations

246 An array of drifting buoys, comprising the DN, track sea ice motion in the vicinity of the CO.
247 Figure 1b shows the positions of the DN sites on February 1st, 2020 within 60 km of the central
248 observatory. An additional 13 buoys comprise the “extended DN” (ExDN) and provide
249 information on larger-scale ice motion (Figure 1a, colored circles). Each buoy reports positions
250 via the Global Positioning System (GPS) with time resolution ranging from 10 minutes to 4
251 hours; the majority of buoys sampled at least once every hour. We only use buoys with (a) time
252 resolution of three hours or less and (b) at least 80% data coverage between 25 January and 5
253 February 2020. Initial buoy processing is described in Bliss et al. (2023). In addition, anomalous
254 points due to large random GPS errors were identified and removed by calculating the Z-score of
255 the minimum of velocities estimated by forward and backward differences relative to a 3 day
256 centered window. Observations were aligned to a 30 minute grid using natural cubic spline
257 interpolation. During the study period, 64 buoys were operating, out of which 57 fulfill study
258 criteria. These sites are referred to as position sites, or “P-sites”. An additional 11 sites
259 (including the L-sites and the CO) contain multiple instruments. We selected a reference buoy
260 from each of these sites, preferring those with higher sampling rate and data precision (Table
261 S1). Choice of reference buoys is arbitrary in most cases, as the buoys at each site are closely
262 situated.

263 The buoy trajectories provide information about the divergence, shear and vorticity of sea ice
264 inside the DN. We calculate strain-rate components using a Green’s Theorem method, following
265 Hutchings et al., (2012, errata 2018). The area over which deformation is estimated can be varied
266 depending on sites chosen to surround the region of interest. We consider deformation on a
267 variety of scales including: the triangle with L-sites at its vertices, a set of 5 polygons with length
268 scales of 15-30 km covering the DN, a polygon enclosing the full DN with length scale 57 km
269 (Figure 10), and two polygons for the left and right sections of the ExDN (Figure 9). Hence, the
270 DN Full array is an estimate of average deformation within the DN, while the smaller polygons
271 in Figure 10 give an indication of the variability within the array.

272 2.4 Sea ice radar imaging

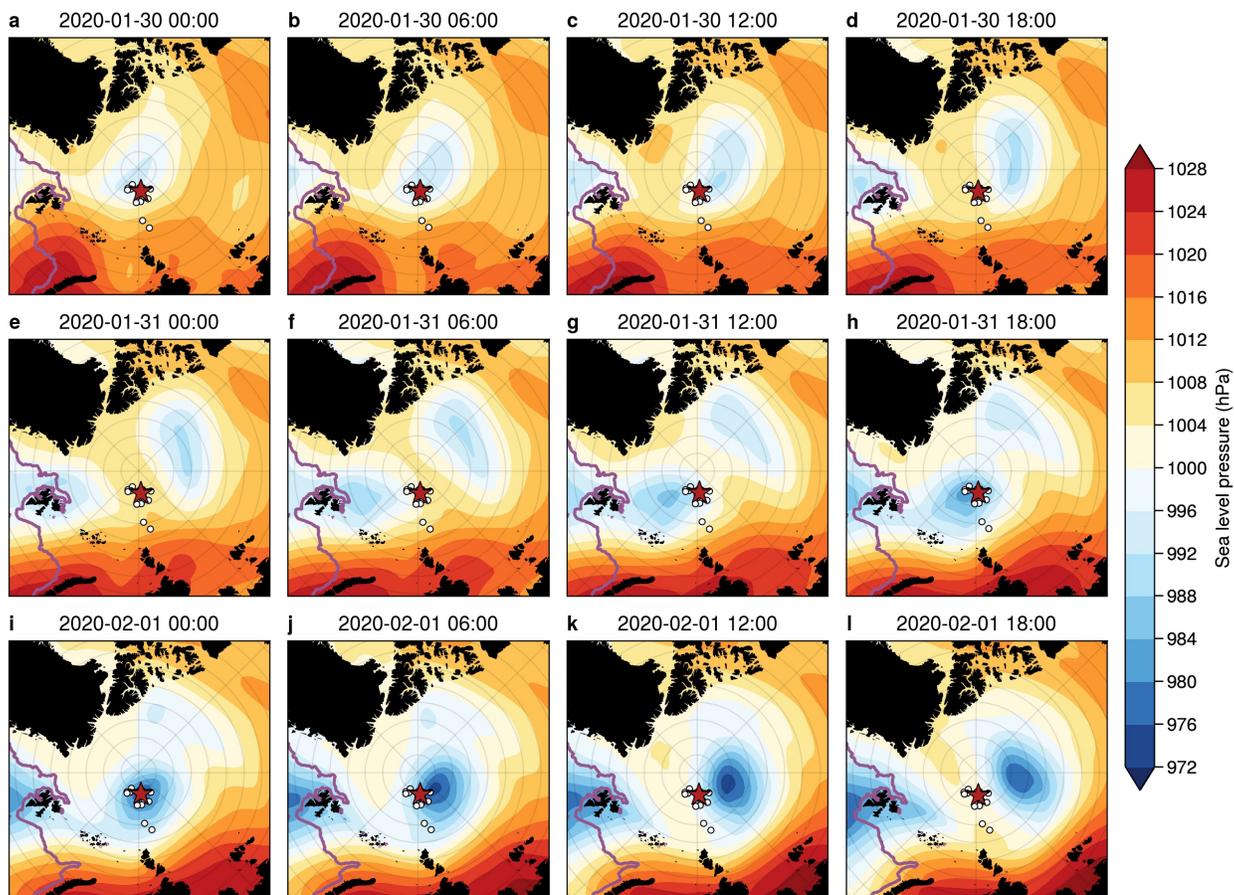
273 Local sea ice deformation observations were obtained from a ship radar-image digitizing system.
274 The system was connected to the 9.4 GHz X-band radar mounted at the roof of the *R/V*

275 *Polarstern* (Hessner et al., 2019). Images of sea ice backscatter were collected with 8.3 m
276 resolution every 2.5 seconds. We use a set of processed and georeferenced images (Krumphen et
277 al., 2021a) downsampled to approximately 15 minute resolution. Images are centered at the *R/V*
278 *Polarstern* and extend radially to a distance of 3 nautical miles (approximately 5.4 km).

279 2.5 Upper-ocean turbulence and current measurements

280 Ocean timeseries observations were made from Autonomous Ocean Flux Buoys (AOFBs;
281 Stanton et al., 2012) adjacent to the CO Met Tower and from the three L-sites. Each of these
282 buoys supported a 4 m deep eddy-correlation turbulence sensor package providing direct heat,
283 salt and momentum fluxes every 2 hours from 35-minute ensemble co-spectra of the 2 Hz
284 sampled 3-component velocity, temperature and conductivity timeseries. Ocean friction
285 velocities $u_o^* = (\langle u'w' \rangle^2 + \langle v'w' \rangle^2)^{\frac{1}{2}}$ from these co-spectra are used to infer the upper ocean
286 stress ($\rho_o u_o^{*2}$) at 4 m. A co-located Acoustic Doppler Current Profiler (ADCP) measured current
287 profiles from 6 m to 80 m depth sampling every 2.5 s and every 2 m in depth, and reporting 15-
288 minute ensembles with $<1 \text{ cm s}^{-1}$ noise levels. Earth-referenced absolute current profiles were
289 calculated from the instrument-coordinate ADCP measurements by first rotating the component
290 profiles into true north coordinates using declination-corrected fluxgate compass measurements
291 in the ADCP and flux package, and, where possible, comparison with shipboard and ASFS GPS-
292 based heading observations. The AOFB / ice floe horizontal motion was then removed using the
293 AOFB GPS timeseries to form absolute u/v vector current profiles.

294 In this study, we use ocean measurements from the CO site adjacent to the main Met City tower.
295 Water density profiles were calculated from the intermittent ship CTD and microstructure
296 profiling program at the CO. Difficult operating conditions during this period of very high winds
297 limited CTD sampling at the CO to as little as once per day. Seasonal mixed layer depths are
298 estimated from the depth in each profile where there is a 0.2 kg m^{-3} potential density increase
299 from the 8 m near-surface values. These sparse-in-time mixed layer depths are linearly
300 interpolated in time and smoothed with a 12-hr period running average filter to estimate the
301 depth of the top of the strong halocline observed across much of the Arctic. A much more
302 sensitive density threshold of 0.01 kg m^{-3} is used as an indicator of the base of the active surface
303 boundary layer in the analysis in section 5.

304 **3 Atmospheric structure and evolution**305 **3.1 Synoptic evolution**

306

307 **Figure 2.** Sequence of ERA5 mean sea-level pressure (SLP) analyses at 6-hour intervals from 30
 308 January 00:00 UTC to 1 February 18:00 UTC. The red star indicates the location of the CO, and
 309 the white circles show the buoys in the DN and ExDN. The position of the ice edge from the
 310 daily NSIDC 12.5 km AMSR2 sea ice concentration (SIC), defined as the 15% SIC isopleth, is
 311 indicated with a purple line.

312

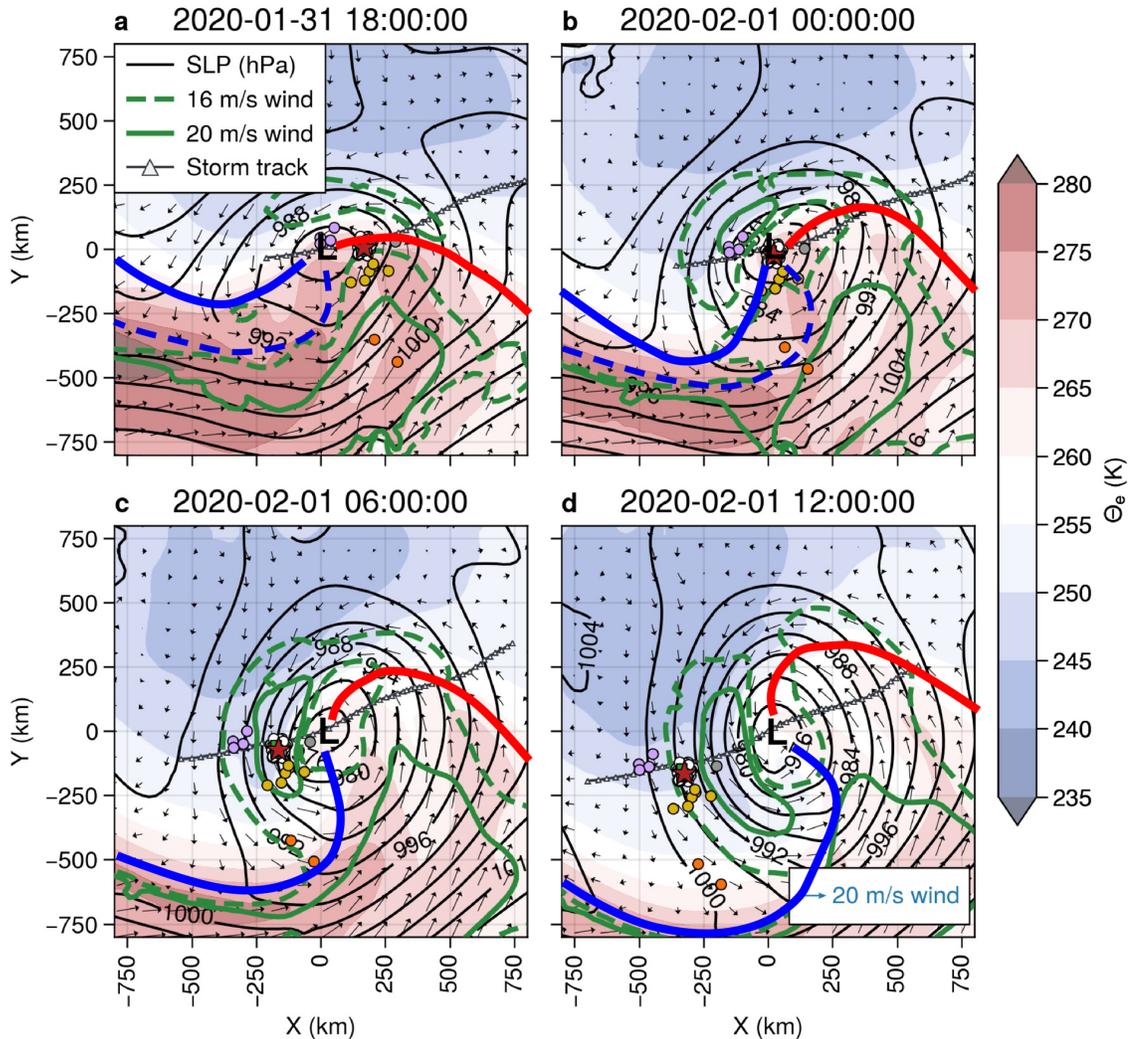
313 Two cyclones crossed the MOSAiC domain in short succession between January 29th and
 314 February 1st, 2020 (Figure 2). The first cyclone (C_1) developed along the NE coast of Greenland
 315 on 29 January, strengthening slightly as it moved northward over the North Pole (Figure 2a-c).
 316 Soundings at the *R/V Polarstern* suggest that a warm front/cold front couplet developed with the

317 system (discussed in Section 3.2) and that the warm sector passed over the MOSAiC domain. As
318 this first cyclone was passing the MOSAiC domain, a second cyclone (C_2) developed along the
319 west coast of Svalbard on 30 January and strengthened along Svalbard's north coastline as it
320 moved northward (Figure 2c). While C_1 only deepened by about 7 hPa along its track, C_2
321 deepened by nearly 20 hPa, becoming one of the deepest cyclones to pass over the MOSAiC
322 domain during the year (Figure 2i). The observed SLP minimum (974 hPa) in the MOSAiC
323 domain during C_2 was 4 hPa lower than the minimum central pressure in the ERA5 fields,
324 indicating that the observed cyclone was slightly stronger than in ERA5. A warm front/cold front
325 couplet also developed with this system, both of which passed over the MOSAiC observatory.

326 3.2 Key mesoscale structures

327 Figure 3 shows an atmospheric frontal analysis of C_2 based on the 6-hourly ERA5 mean sea-level
328 pressure, 10 m wind vectors, 925 hPa equivalent potential temperature (θ_e), and 950 hPa wind
329 speed. In this and following figures, references to cardinal directions are relative to the CO. The
330 polar stereographic maps are oriented so that north from the CO is in the positive y direction and
331 east is in the positive x direction; note that the North Pole is 267 km north of the CO, so the
332 direction of true north will vary substantially throughout the figure. The storm deepened by 8
333 hPa during the 18 hours shown and has clear spatial structure, with northward warm-air
334 advection in the warm sector primarily to the right of the storm track ahead of the low center and
335 southward cold-air advection in the cold sector primarily to the left of the storm track and behind
336 the low center. The surface warm front passes over the CO (red star) on 31 January between 14
337 UTC and 16 UTC (Figure 3a), while a cold front aloft passes over the CO on 31 January near 23
338 UTC and a surface cold front passes over the CO near 02 UTC. The surface low passes very
339 close to the CO but just to its west and north, such that the CO is initially in the warm sector air
340 before being affected by the trailing surface cold front. Strong low-level wind speeds indicating a
341 low-level jet (LLJ) initially occur in the warm sector between the warm front and the cold front
342 aloft. By 1 February 00 UTC (Figure 3b), a LLJ (indicated here by the 16 m s⁻¹ isotach at 950
343 hPa) encircles the surface low and remains as a nearly axisymmetric annulus through the rest of
344 the time period as the system occludes with bands of warm and cold air wrapping around the low
345 center (Figure 3c,d). Figure 2 suggests that C_2 was more axisymmetric (circular) than C_1 . While
346 C_2 is quasi-axisymmetric initially and becomes even more axisymmetric as it strengthens, C_1
347 starts out more elongated and becomes even more so with time. C_1 appears eventually to be

348 absorbed into C₂. We hypothesize that the symmetry of the storm is an important factor in the
 349 development of the axisymmetric mesoscale LLJ annulus.



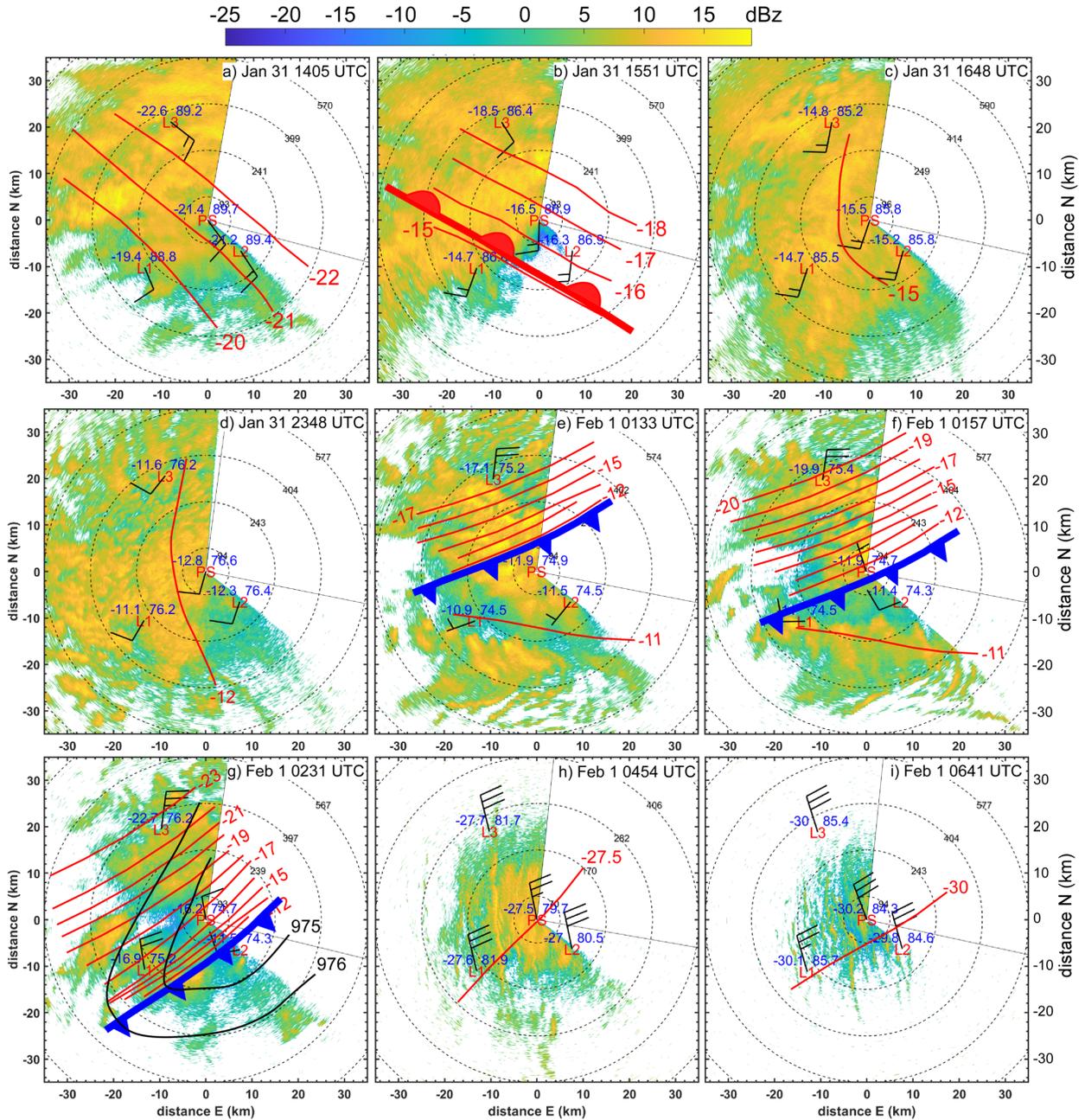
350

351 **Figure 3.** ERA5 reanalyses centered on the SLP minimum for cyclone C₂. Shown are SLP (hPa;
 352 black isopleths), 925 hPa equivalent potential temperature θ_e (K; colors), 10 m wind vectors, and
 353 select 950 hPa isotachs (16 and 20 m s⁻¹; green). Every fourth wind vector is plotted for clarity;
 354 vector length is proportional to wind magnitude. The heavy red and blue lines show the positions
 355 of the warm and cold fronts, respectively. Dashed front lines indicate thermal features aloft,
 356 while the solid lines depict fronts at the surface. The light gray line shows the track of the low

357 center within the domain, while the colored circles show buoy positions. Colors for DN and Ex
358 DN sites are as in Figure 1.

359 Utilizing the near-surface observations at the three ASFS sites and at the CO, rough
360 observational surface analyses are possible on the ~ 50 km scale of the DN. Figure 4 shows
361 isotherm analyses centered on the CO between 31 January 14:05 UTC, just prior to the passage
362 of the surface warm front, and 1 February 06:41 UTC, nearly 5 hours after the passage of the
363 surface cold front. These are overlaid on low-elevation radar-reflectivity PPI scans to provide an
364 indication of the spatial distribution and structure of the clouds ($< \sim 0$ dBZ) and precipitation ($>$
365 ~ 0 dBZ).

366 Moderate ($5\text{-}10\text{ m s}^{-1}$) southeasterly surface winds were present throughout the domain as the air
367 temperatures warmed with the approaching surface warm front (Figure 4b). Within the warm
368 sector, temperatures gradually warmed to $-11\text{ }^{\circ}\text{C}$. Winds were initially moderate from the SSW
369 but decreased in magnitude as the low-pressure center neared the CO, particularly after the cold-
370 front aloft passed overhead. The more cellular nature of the clouds and precipitation after the
371 cold front aloft passed can be seen in comparing Figure 4c and 4d. The trailing surface cold
372 front entered the DN from the NW, marked by a sudden wind shift to the N and a trailing, very
373 strong temperature gradient (Figure 4e-g). Air temperatures dropped to between $-27\text{ }^{\circ}\text{C}$ and -30
374 $^{\circ}\text{C}$. The cold front took about 1.5 h to traverse the L-site triangle. The northerly winds
375 increased throughout this frontal zone, reaching near-surface speeds of $12\text{-}15\text{ m s}^{-1}$ as the LLJ
376 behind the front passed overhead (Figure 4h-i). High wind speeds combined with strong cold-air
377 advection leads to strong mixing near the surface, producing what appears to be horizontal roll
378 vortices in the atmospheric boundary layer (suggested by the linear, along-wind, cloud and
379 precipitation features). Horizontal roll vortices are an effective mechanism for vertical mixing in
380 the atmospheric boundary layer (Etling & Brown, 1993; LeMone, 1973).



381

382 **Figure 4.** Near-surface meteorological observations of air temperature ($^{\circ}\text{C}$), SLP (hPa),
 383 downwelling longwave radiation (W/m^2) and wind barbs from the three L-site ASFSs (3.8 m)
 384 and at Met City (6 m)) during the warm-frontal (heavy red, lobed line) passage on 31 January (a-
 385 b), cold frontal (heavy blue toothed line) passage on 1 February (e-g), and in the post-cold-
 386 frontal sector (h-i). Manual analysis of air temperature is shown in red lines with 1°C isotherm
 387 interval. Panels with only one isotherm represent times when the spatial temperature difference
 388 between sites is less than 1°C . Isobar analysis is depicted with black lines in panel g. The

389 background shows PPIs of the low-elevation scanning Ka-band radar reflectivity (color, dBZ).
390 Range rings (black dashed lines) are shown at 5, 15, 25, 35, and 45 km distances, and are labeled
391 with the radar-beam height (m) above the local surface. The thin black radii bracket the region
392 not scanned by the radar. North is upwards for each panel.

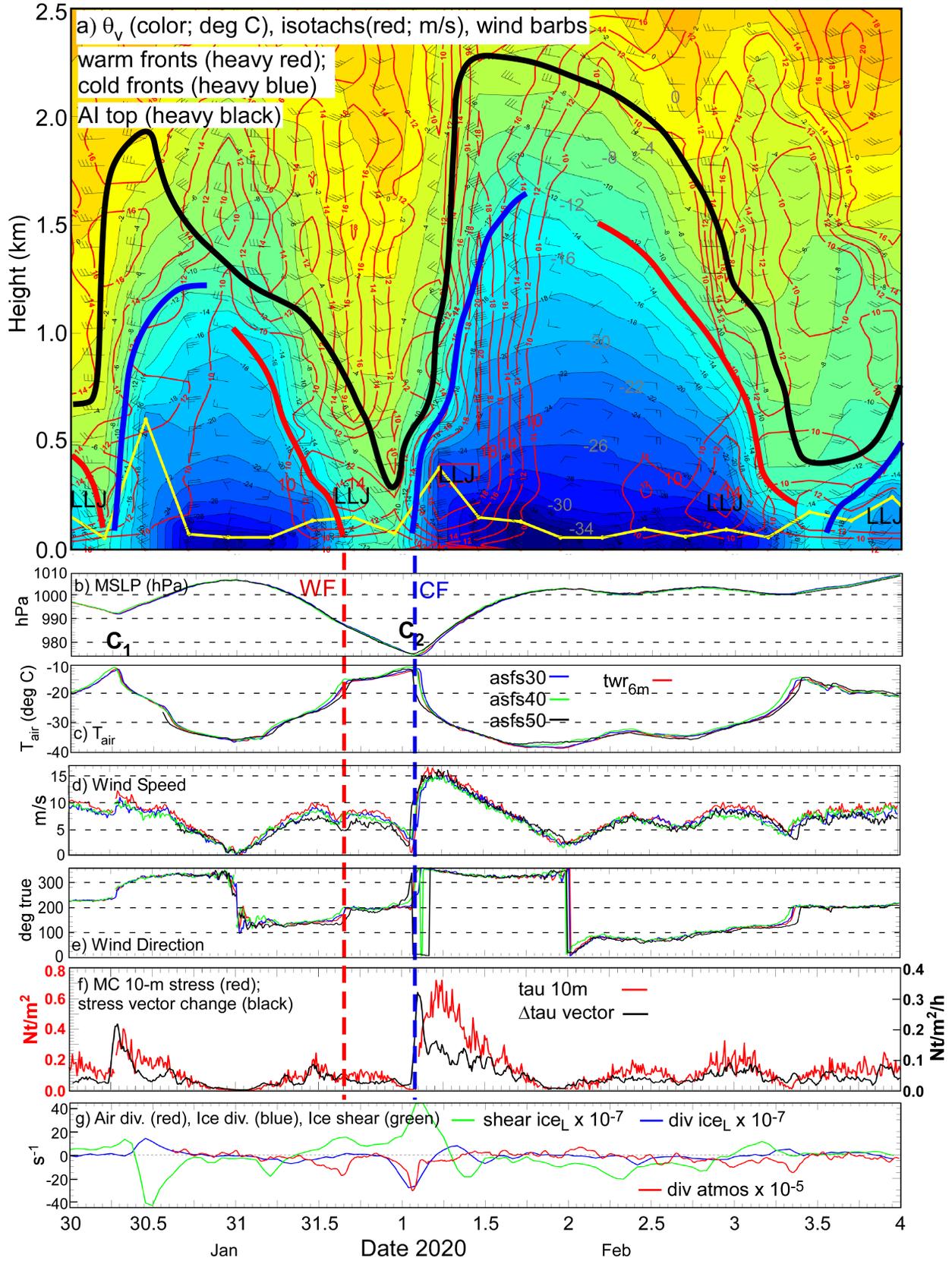
393 A time-height section of serial rawinsonde data and near-surface time series of various
394 parameters (Figure 5) confirm the features passing over the MOSAiC domain discussed above.
395 The passage of the first cyclone (C_1) and its associated narrow warm-sector are clearly seen, with
396 the brief but distinct warm air peak in the warm sector, and the cooling and veering of the
397 surface wind with the passage of the cold front. A LLJ is present at approximately 250 m above
398 the surface near the time of the warm-frontal passage. The second cyclone (C_2) is deeper with a
399 broader warm sector over the CO. The air warms only slightly in the warm sector between the
400 warm front and the cold front, but the thermal wind effect from this thermal gradient, with the
401 warmest air closest to the cold front, is a likely cause for the observed LLJ within the warm
402 sector at 250-300 m height. The rawinsondes show the warm-sector LLJ wind speed maximum
403 near 15 m s^{-1} just above the surface mixed layer (SML), with associated near-surface wind
404 speeds of $7\text{-}8 \text{ m s}^{-1}$ (Figure 5a, d; Figure 4c).

405 The passage of the cold front with C_2 near 02 UTC on 1 February marks the time of the lowest
406 observed central pressure (974 hPa), a very sharp drop in surface temperature (-11°C to -38°C in
407 only 12 h), a minimum in surface winds, and a rapid change in surface wind direction (Figure
408 5b-e; Figure 4e-g). A second LLJ is observed at $\sim 350\text{-}400$ m height just behind the cold front
409 with a core speed of $21\text{-}22 \text{ m s}^{-1}$ and with temporarily deeper SML as indicated by the constant
410 θ_v with height. Just after the cold-frontal passage, the near-surface wind speed increases with
411 the arrival of the LLJ above, reaching speeds of $14\text{-}16 \text{ m s}^{-1}$ across the four observational sites
412 between 04 and 06 UTC on 1 February (see also Figure 4h-i). The timing differences in the
413 wind direction shifts, wind speed increases, and temperature decreases between sites (Figure 5d-
414 e) represent the progression of the cold front across the DN from the northwest. Stability
415 differences in the sub-jet layers may cause the higher surface wind speed relative to its core
416 strength for this second post-cold-frontal LLJ when compared to the warm-sector LLJ. A peak
417 in the observed covariance surface stress (τ_a) at Met City (Figure 5f) occurs just after the cold
418 frontal passage, and is coincident with the deepening of the SML just below the LLJ (Figure 5a)
419 and the appearance of the likely horizontal roll vortices (Figure 4h-i). It is unclear whether

420 enhanced turbulence caused by the LLJ or the roll vortices have deepened the SML, or if the
421 deeper SML has weakened the near-surface winds thereby producing a LLJ just above the SML.
422 All of these features indicate significant, efficient, vertical momentum transport at this time.

423 The presence of the LLJ behind the cold front is likely due to the LLJ being quasi-axisymmetric
424 around C_2 . This “wrap-around” LLJ is likely an extension of the LLJ observed in the warm
425 sector as this warm air wraps around the strong but compact cyclone center, as seen in Figure 3a-
426 d. Note that the cold sector LLJ is at a slightly higher altitude than the LLJ in the warm sector,
427 consistent with some lifting as the LLJ has wrapped itself around the cyclone center. There is
428 likely a thermal wind contribution over a ~ 100 m deep layer at the core of this second LLJ. We
429 infer that the presence of two LLJs in fairly rapid succession, possibly parts of a wrap-around
430 LLJ within a rapidly moving cyclone, produces strong, rapid surface wind speed and wind
431 direction changes as it translates across the MOSAiC domain. This is the key forcing for the
432 significant ice motion, ice deformation and upper-ocean current changes observed during the
433 passage of C_2 . Such a double LLJ (wrap-around LLJ) was also observed in other MOSAiC
434 cyclones with significant ice deformation (e.g., Persson et al., 2023).

435 Low-level atmospheric divergence is one way to quantify these wind transitions. Indeed, we
436 observe significant atmospheric convergence with the passage of the warm and cold fronts of
437 cyclone C_2 , with the strongest convergence ($\sim 30 \times 10^{-5} \text{ s}^{-1}$) occurring with the cold-frontal
438 passage (Figure 5g). Ice convergence and shearing also show peaks. There is no appreciable ice
439 divergence/ convergence within the L-site triangle associated with the warm frontal passage,
440 though there is significant shearing of the ice (Figure 5g). Note that other substantial wind
441 transition events also show some atmospheric divergence/convergence and some ice deformation
442 (e.g., near 08 UTC on 3 February, also associated with a LLJ). Nevertheless, ice deformation
443 events can have multiple local and nonlocal causes, and therefore are only sometimes associated
444 with local atmospheric divergence.



446 **Figure 5.** a) Time-height section of virtual potential temperature (θ_v , colors; °C, gray labels),
 447 isotachs (red), and select wind barbs from serial rawinsondes at the *R/V Polarstern*. Heavy (red,
 448 blue, black) lines mark warm and cold fronts and the top of the Arctic inversion (base of the free
 449 troposphere), respectively. The thin yellow line marks the top of the surface mixed-layer. Low-
 450 level jets (LLJ) are also marked. Rawinsondes were launched every 6 hours, launch times
 451 indicated by the origin of the wind barbs. Lower 6 panels: Time-series from the Met City tower
 452 and the ASFSs of b) MSLP; c) T_a ; d) 10 m (Met City) and 3.8 m wind speed; e) 10 m and 3.8 m
 453 wind direction; f) 10-m atmospheric stress (red), stress vector change (black) at MC; and g) 3.8
 454 m atmospheric divergence (red), ice divergence (blue), and ice shear (green). The vertical dashed
 455 lines show the times when the warm (red) and cold (blue) fronts with the second cyclone pass
 456 over. C_1 and C_2 mark the minima in MSLP with first and second cyclone, respectively.

457 **4 Sea ice dynamics**

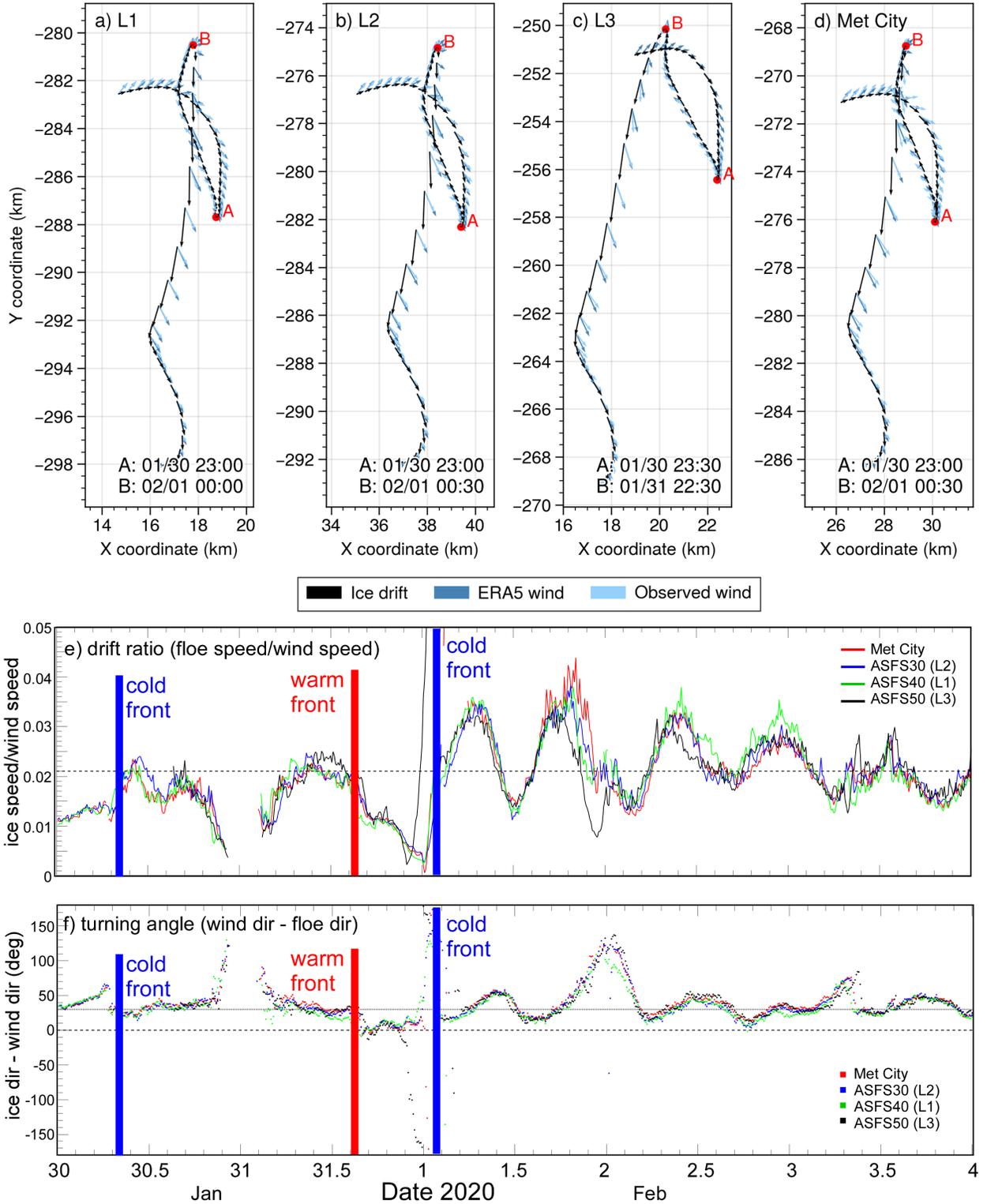
458 4.1 Atmosphere-ice interaction

459 Sea ice motion during the passage of the cyclones is broadly coherent with the time-varying
 460 wind forcing. The drift speed ratio (α , the ratio between the local drift and wind speeds) and net
 461 turning angle (θ , the difference between the local wind and drift directions) vary significantly
 462 over time (Figure 6e, f). Note that the wind velocity in ERA5 is in good agreement with the
 463 observed winds (dark and light blue arrows in Figure 6a-d). The drift speed ratio and turning
 464 angle are empirical measures of the relationship between the ice drift and the wind speed. In
 465 steady state free drift, θ is a function of the boundary layer structure and the ice surface
 466 roughness, and α is a function of the air-ice and ice-ocean drag coefficients and the densities of
 467 each medium. For the period shown in Figure 6e-f, average values of α and θ are 0.021 and 35,
 468 respectively, consistent with previous studies (Leppäranta, 2007; Schweiger & Zhang, 2015;
 469 Womack et al., 2022). Drift speed ratios are low in the warm sectors of both cyclones, while they
 470 increase after the arrival of the cold fronts. The turning angle in the warm sector is slightly larger
 471 than the mean for C_1 , but below the mean for C_2 . Drift speed ratio remains high after the cold
 472 front on 1 February, which suggests that a larger fraction of atmospheric momentum is being
 473 converted into ice motion rather than adding to the internal ice stresses. Increases in α following
 474 cyclone passage has been observed previously (e.g., Itkin et al., 2017). The drift speed ratio

475 following the second cyclone passage oscillates near the inertial frequency (~ 12 hours),
476 suggesting the possibility of inertial oscillations following the storm.

477 Drift trajectories of the three L-sites and the CO (Figure 6a-d) generally illustrate the expected
478 right hand turning rule ($\theta > 0$) first noted by Nansen (1902). At all four sites, the ice drift arcs to
479 the right and slows as the cyclone C_1 moves away from the MOSAiC site. During the passage of
480 the pressure ridge between cyclones C_1 and C_2 , the wind direction abruptly reverses (Figure 5e;
481 Figure 6a-d). This reversal occurs at 23 UTC on 30 January at all sites, marked by the red letter
482 A. As SLP decreases and C_2 approaches the CO, the ice drifts northward due to southerly winds
483 until slowing to a halt and again reversing direction. The cusp in the trajectory marking the
484 reversal is indicated by the letter B. Notably, this reversal precedes a rapid acceleration, and
485 occurs at different times at each site: first at 22:30 UTC on 31 January at L3, next at 0 UTC on 1
486 February at sites L1 and at the CO (Met City), and last at 0:30 UTC on 1 February at L2. These
487 times are all about 2 h prior to the passage of the cold front and the large change in wind
488 direction at each site. Cusps in DN buoy trajectories are identified by local minima in drift speed.
489 Cusp timestamps display a west-east gradient spanning a 3-hour period consistent with ~ 25 km/h
490 (~ 7 m s $^{-1}$) cyclone propagation speed (Figure 7b), with some deviations likely due to propagating
491 internal ice stresses from nonlocal forcing. That is, it appears as if the 2 h difference in wind and
492 ice drift direction changes and the deviations in the propagation of the ice drift reversals may be
493 due to internal ice stresses caused by non-local wind forcing behind the cold front. Further from
494 the storm center, 48-hour drift trajectories show clockwise motion to the right of the storm
495 (Figure 7c, d) and counter-clockwise motion to the left (Figure 7a), with the sharpness of the turn
496 increasing nearer the storm center due to the smaller radii of the quasi-annular wind around the
497 low center.

498

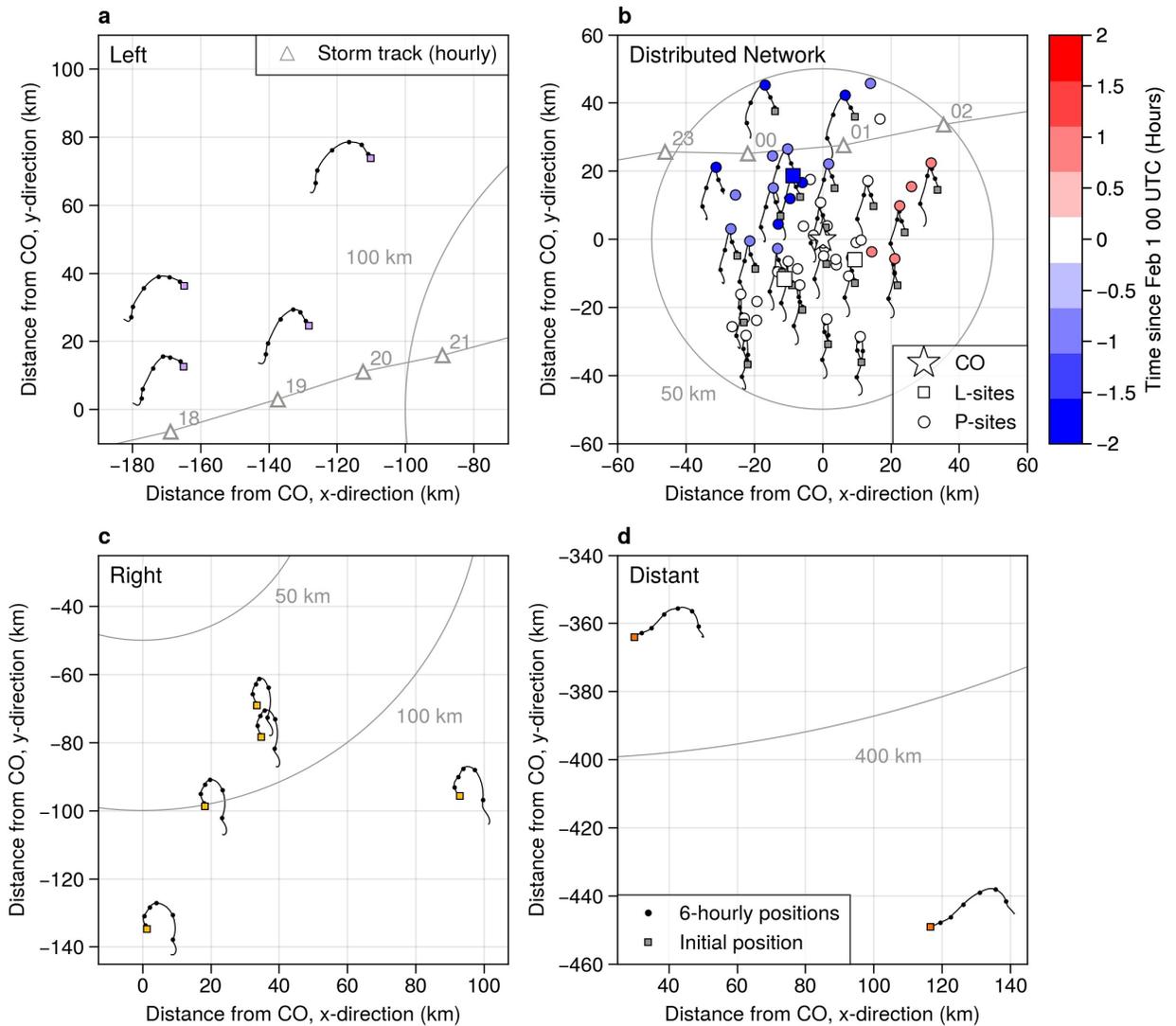


499

500 Figure 6. Top a-d: Trajectories of sites L1, L2, L3, and Met City at hourly resolution from 30
 501 January 00:00 UTC on 02 February 00:00 UTC. Black arrows indicate the ice drift direction,

502 light blue the observed wind direction, and dark blue the estimated wind from the ERA5
 503 reanalysis. Arrow length is proportional to speed; wind speeds are scaled to 2% for comparison
 504 with the drift speed. Bottom: (e) Drift speed ratio (ice speed divided by wind speed) and (f)
 505 empirical turning angle (difference between wind direction and ice drift direction) derived from
 506 10-min wind and ice drift observations at sites L1, L2, L3, and Met City.

507

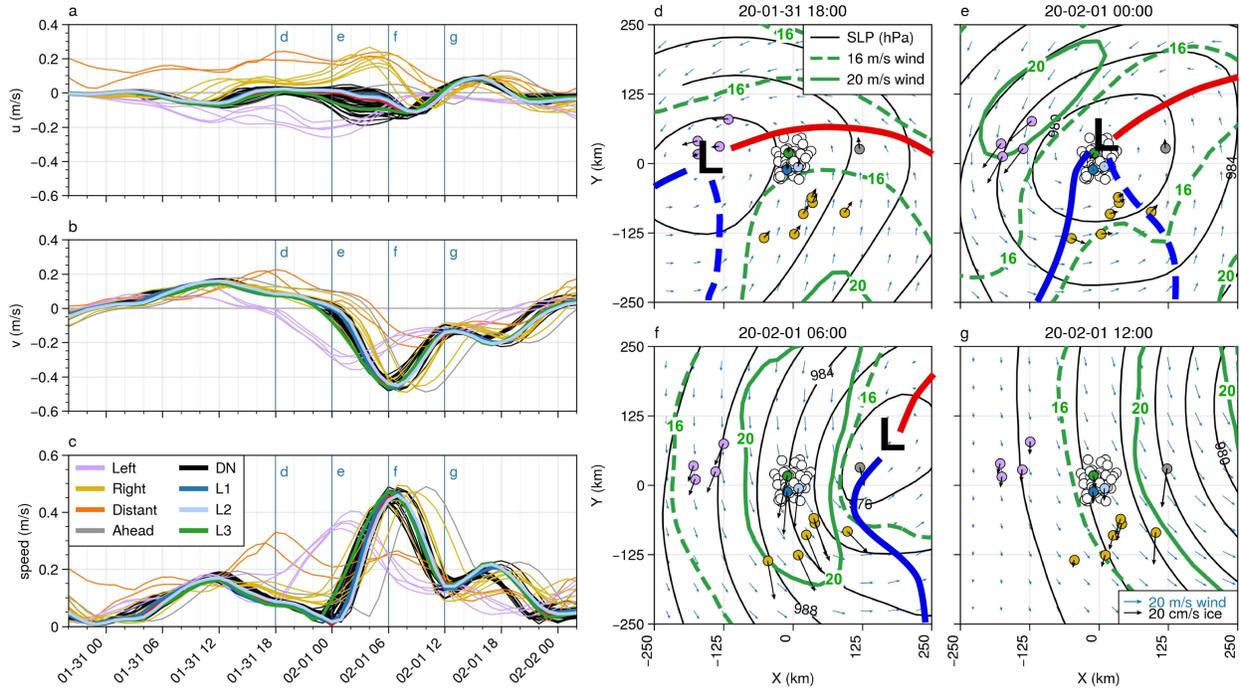


508

509 **Figure 7.** Buoy trajectories from 31 January 00:00 UTC to 2 February 00:00 UTC. Small
 510 squares indicate the beginning of the time series. Black lines show the 30-min resolution drift
 511 tracks, and black circles show the position every 6 hours. Distance from the CO is indicated with
 512 the axis units and radii at 50 km, 100 km, and 350 km. In panel b, the time of the drift speed

513 minimum relative to 1 February 00:00 UTC is indicated with color. Trajectories of a subset of
 514 DN buoys are shown for clarity. L-sites are marked with large squares, while the CO is marked
 515 by a star. In panels a and b, the position of the sea level pressure low from the ERA5 reanalysis
 516 is marked at hourly intervals with triangles and labeled with the corresponding hour of day.

517



518

519 **Figure 8.** Left: Buoy velocity components (a, b) and magnitude (c) for the period from 30
 520 January 20:00 UTC to 2 February 02:00 UTC. The top and middle panels show the u and v
 521 velocity components relative to the north polar stereographic projection, thus corresponding to
 522 the x and y axis, respectively, in the panels on the right. For the period shown here, the positive y
 523 direction is approximately northward. Right: Snapshots of buoy motion (thick black arrows)
 524 superimposed on the ERA5 sea level pressure isobars (black contours, 4 hPa spacing), near-
 525 surface (10 m) wind fields (blue arrows), and 16 and 20 m s⁻¹ isotachs of the 950 hPa winds
 526 (green contours) at times corresponding to vertical lines in the velocity time series to the left.

527 The position of the SLP minimum is marked with “L”. The cold front is marked in blue and the
528 warm front is marked in red. Solid fronts are surface level and dashed are elevated.

529 The clearest sign of the storm’s impact on the ice velocity is through the effect of the LLJ as it
530 develops and moves across the MOSAiC array. High ice drift speeds indicate efficient downward
531 mixing of momentum through the atmospheric boundary layer. Since buoy velocities were not
532 assimilated in ERA5, coincident ice velocity and 950 hPa wind speed maxima (Figure 8d-g)
533 serve as an independent confirmation that the location of the LLJ in ERA5 is approximately
534 correct. During the second cyclone, the cold sector LLJ core first passes over the left buoy group
535 (Figure 8e), where drift speeds reach an average speed of 37 cm s^{-1} between 01:00 and 02:00
536 UTC on 1 February (Figure 8c) under the LLJ core of $>20 \text{ m s}^{-1}$. The DN buoys come nearly to a
537 full stop before reversing direction and being accelerated by the cold sector LLJ (Figure 8c).
538 Maximum speeds of $42\text{-}49 \text{ cm s}^{-1}$ occur between 05:00 and 08:00 UTC on 1 February (Figure
539 8c) as the LLJ core passes overhead (Figure 8f). There is a larger spread in velocity between the
540 DN buoys during this time, implying deformation. The DN buoys and the right group reach their
541 maximum speeds at approximately the same time (Figure 8c, f) yet due to the wind curvature
542 within the LLJ core, the direction of ice motion is different.

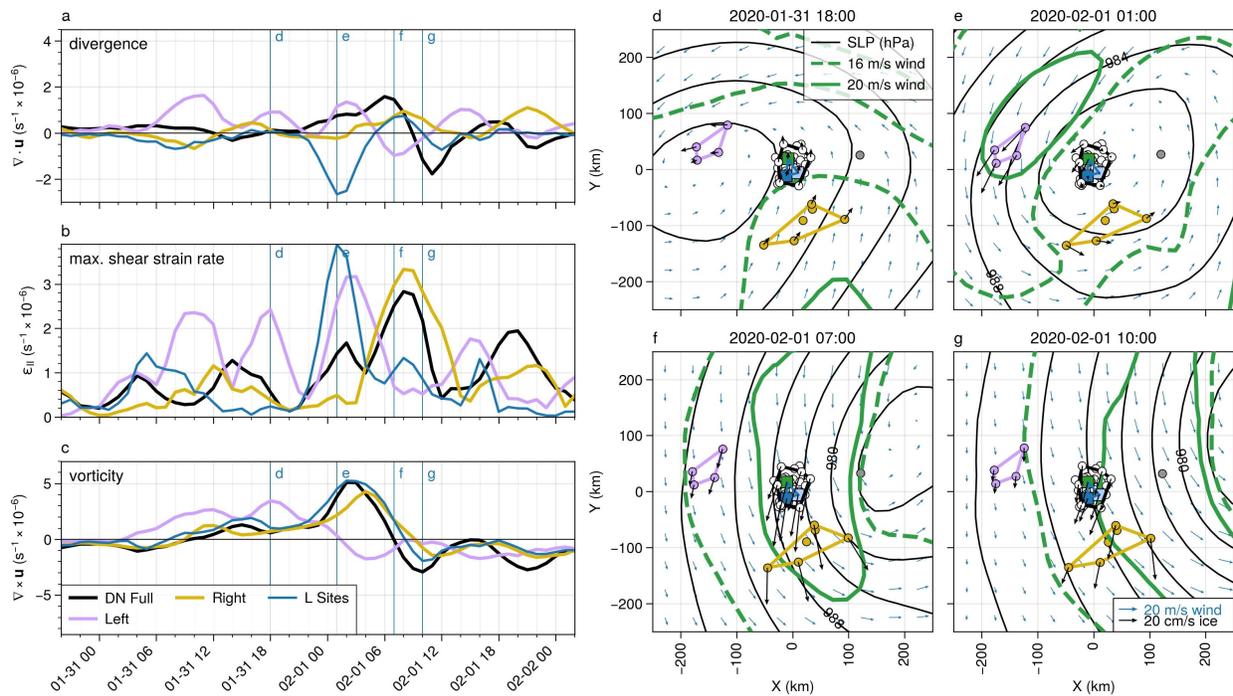
543 4.2 Sea ice deformation

544 Differential motion across the buoy array implies deformation. We measure this deformation by
545 monitoring the changes in polygons formed by subsets of the buoy array (Figures 9, 10) and
546 through examination of ice radar imagery (Figure 11). At moderate-to-large scales (purple,
547 yellow, and black lines and polygons in Figure 9), the largest signals in strain rates can be
548 understood as responses to the cyclone-scale wind gradients and the positions of the LLJ cores.
549 As a band of high wind speeds is advected over the ice, the ice experiences changes in vorticity,
550 divergence, and shear strain rate (Haller et al., 2014; Lindsay, 2002). For the LLJ behind the
551 storm, the vorticity pattern is first cyclonic, accompanied by gradually increasing divergence
552 (opening), then as the wind speed slows, the sense of rotation reverses, and the ice closes. This is
553 seen on 1 February both for the purple (00 to 08 UTC) and black (04-11 UTC) polygons, but at
554 slightly different times. Significant variability exists in the strain rates, particularly maximum
555 shear strain rate, likely due to the complex interaction of the geometry of ice fractures and the
556 varying wind forcing. The vorticity signal is broadly coherent (Figures 9c, 10c), with a clear

557 peak in positive vorticity at 2-3 UTC on 1 February and a trough of strongly negative vorticity
 558 between 9-11 UTC on 1 February in all except the purple buoy group, which has the same
 559 positive/negative vorticity couplet except 6-8 h earlier. The cold-sector trajectory and different
 560 timing of the position of this buoy group relative to the LLJ core seen in Figure 9d-g likely
 561 explain this time difference. This coherent positive ice-vorticity signal should be expected from
 562 the presence of the narrow axisymmetric atmospheric LLJ annulus surrounding the cyclone. The
 563 positive vorticity signal as the storm approaches is damped because the LLJs developing in the
 564 warm and cold sectors of the storm had not yet joined, and the winds ahead of the low center
 565 were weaker than the winds behind it (cf. Figure 3).

566

567



568

569 **Figure 9.** Time series of ice deformation components (a-c) and fields of 10-m winds, MSLP
 570 isobars, and 950 hPa isotachs (d-g) of the DN and Extended DN for the February 1 cyclone.

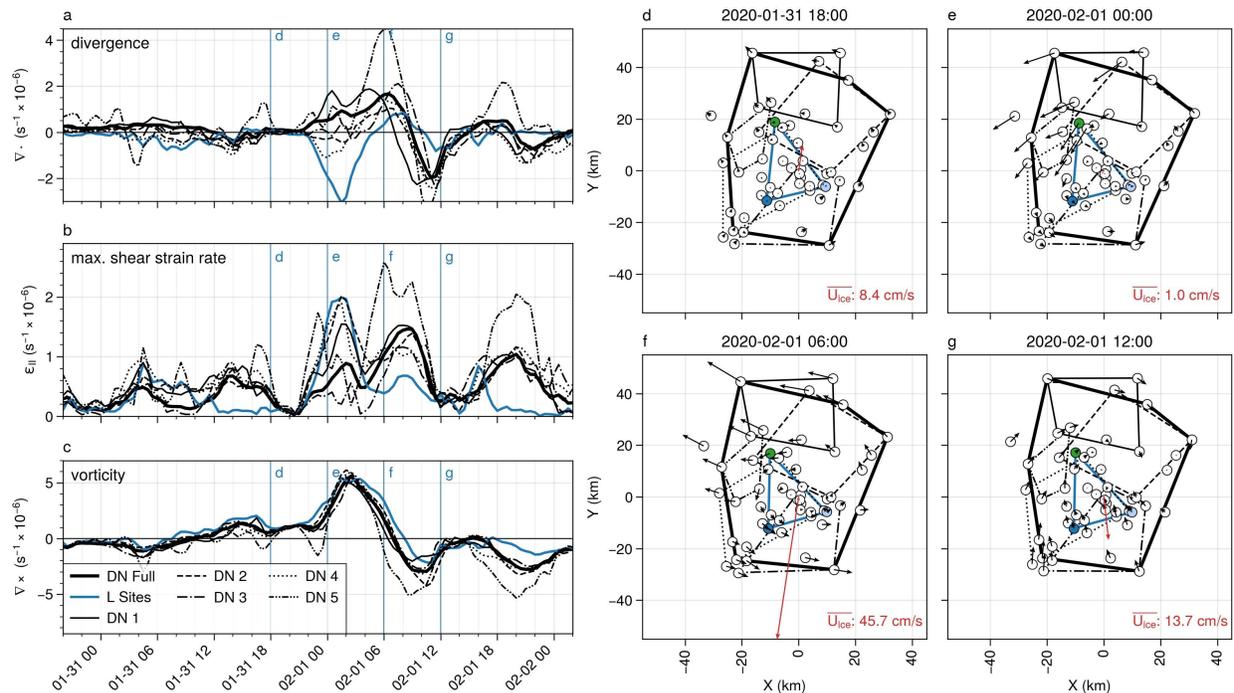
571 Polygons used for the deformation calculations are shown in panels d-g. Polygons were selected

572 manually. Length scales are 57 km, 40 km, and 66 km for the DN Full, Left (purple), and Right
573 (yellow) arrays, respectively, and 18 km for the L-site array (blue).

574 Within the DN, the small-scale polygons generally show ice deformation consistent with the DN
575 Full polygon, with initial positive divergence as the LLJ core approaches the polygon. The
576 importance of local fracture network structure in floe-floe interactions is demonstrated in the
577 broad range of divergence values, and in the outlying behavior of the L-site triangle. Considering
578 the wind field shown in Figure 9d-g and 10d-g, and the tendency of ice to move to the right of
579 the wind, we expect opening (positive divergence) while the low is centered over the DN. In
580 pack ice, individual floe motion is limited by interaction with neighboring floes, described as
581 “multifloe” (~2-10 km) and “aggregate” (10-75 km) motions in the hierarchy proposed by
582 McNutt and Overland (2003). The buoy velocity anomalies show that a region of at least 30 by
583 60 km is moving approximately coherently within the DN (Figure 10d-g). As the wind direction
584 changes, the geometry of the interlocked floes results in different regions moving as aggregates.
585 The passage of the cold sector LLJ, as indicated by the rise and fall of sea ice velocity, occurs
586 within approximately 12 hours (00-12 UTC on 1 February). Differences in the ice motion due to
587 the storm structure are visible at ~100 km (larger than “aggregate”) length scales, while
588 significant deformation is occurring at ~10 km (“multifloe”) length scales. Remote sensing
589 observations of ice drift are only rarely available at higher than 1 day resolution, and most
590 products have spatial resolution between 25-75 km; typically, higher spatial resolution comes at
591 a cost of smaller scenes and longer times between repeat observations (Sandven et al., 2023;
592 Wang et al., 2022). Global-scale coupled model experiments have primarily been run on 0.25° or
593 coarser grids (e.g., Long et al., 2021; Taylor et al., 2012). Thus, the strongest impact of the storm
594 on the ice velocity and, especially, deformation is occurring at time and space scales shorter and
595 smaller than many satellite ice motion observations and coupled model resolutions can resolve.

596 Within a consolidated ice cover, there is considerable resistance to ice opening, though some
597 leads do open. As the winds recede, the newly opened leads offer little resistance to convergent
598 motion. Thus, there is considerable spread in positive divergence across the DN polygons from
599 31 January 22 UTC to approximately 9 UTC on 1 February, while convergence after that time is
600 faster and more cohesive across the set of polygons (Figure 10a). The position of shear zones can
601 lead to ambiguity in area-average strain rates, as discussed in Lindsay (2002) (see also Bouillon
602 & Rampal, 2015; Lindsay & Stern, 2003; Thorndike, 1986). The anomalous convergence shown

603 in Figure 9a and 10a for the L-site triangle is an artifact of the triangle orientation relative to a
 604 shear zone that cuts through it. The shearing motion is a discontinuity in the ice velocity that
 605 leads to the triangle area to not be representative of the deformation, and compression is
 606 overestimated. The presence of this shear zone is clearly visible in the velocity anomaly map of
 607 Figure 10e. Higher confidence can be placed in the estimate of deformation from the DN Full
 608 array due to the larger number of buoys (vertices) used and larger area relative to shear zones.
 609 Over the 10-day period from 26 January 2020 to 5 February 2020, the area of the DN Full
 610 polygon changed from $3.17 \times 10^3 \text{ km}^2$ to $3.21 \times 10^3 \text{ km}^2$, a change of just over 1%. Rapid area
 611 increase (i.e., positive divergence) occurred on 1 February due to the passage of the cold sector
 612 LLJ from 00-09 UTC (Figure 10a), such that the area of the polygon increased by 3.5% in a 9-
 613 hour period, likely producing leads. Given that the surface air temperature was $-10 \text{ }^\circ\text{C}$ or below
 614 during this period, any leads would have quickly begun freezing over. The resulting net increase
 615 of area over the 10-day period represents both thermodynamic ice growth in leads and
 616 mechanical redistribution of ice thickness in subsequent convergence.

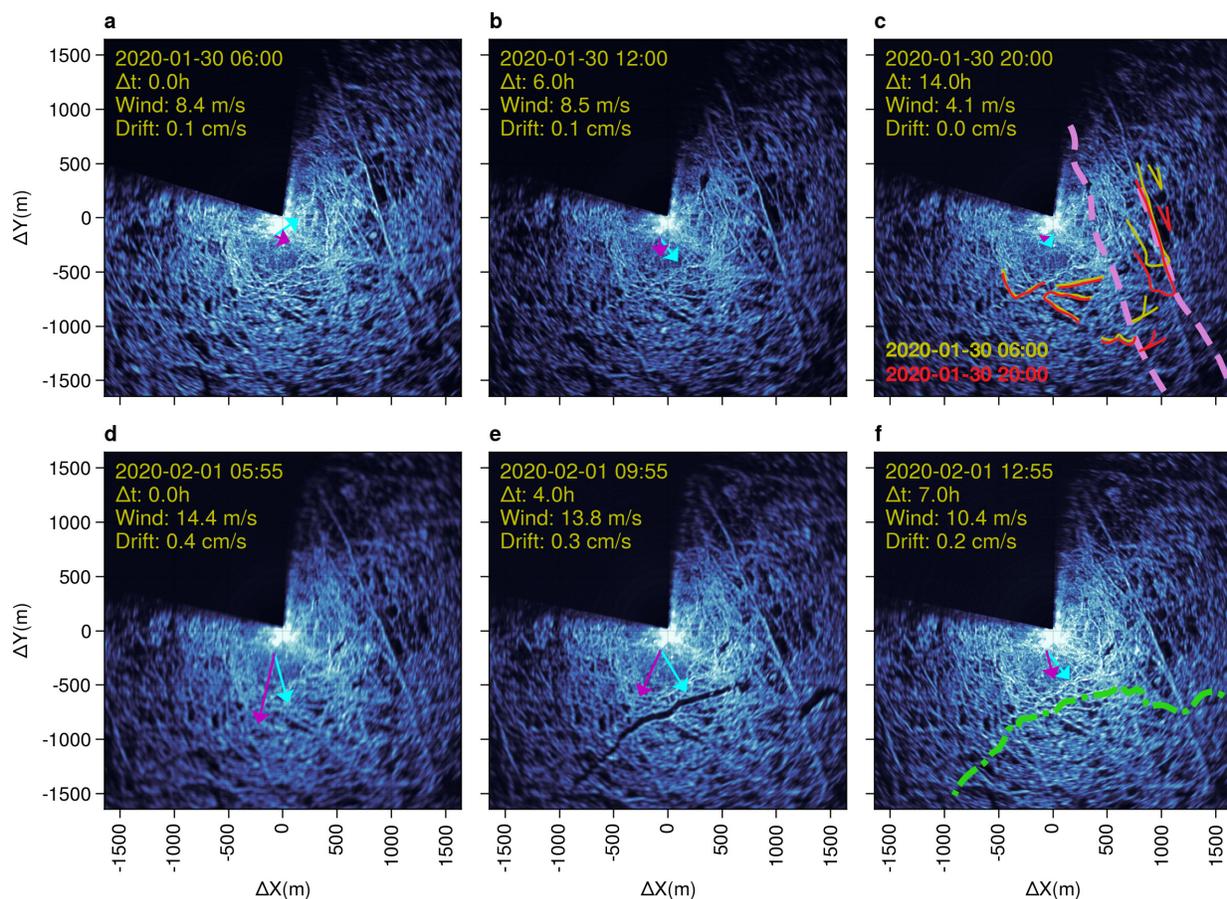


617

618 **Figure 10.** Time series of ice deformation components (a-c) and fields of buoy velocity
 619 anomalies (d-g) in the DN for the February 1 cyclone. Polygons used for the deformation
 620 calculations are shown in panels d-g; the polygons were selected manually. Note that the buoy in

621 the upper left was not included in the DN Full array due to periods of missing data. Velocity
 622 anomalies in panels d-g were computed relative to the ensemble average velocity, which is
 623 shown as the red arrow at the center of each panel with magnitude shown in the lower right. The
 624 length scales of the polygons (square root of the average polygon area) are 28, 33, 28, 30, and 15
 625 km for DN sub-arrays 1-5 respectively, 18 km for the L-site triangle, and 57 km for the DN Full
 626 array.

627



628

629 **Figure 11.** Backscatter from the ice radar on board the R/V Polarstern show small-scale
 630 deformation near the CO. The dark sector is blocked from the radar by the ship superstructure.
 631 Image date and time, wind speed, and drift speed are indicated in the upper left of each panel.
 632 Elapsed time (Δt) since the first image in each row is also indicated. Magenta and cyan arrows
 633 show the ice drift and wind directions at the Met City tower, respectively. Arrow length is

634 proportional to speed; wind speed has been scaled by 2% for comparison with drift speed.
635 Annotations in panel c show the relative movement of manually identified ice features at 30
636 January 06 UTC (yellow) and at 20 UTC (red), revealing two prominent shear zones marked
637 with the dashed pink lines. The green dashed line in panel f shows the location of a fracture that
638 began opening at 1 February 07 UTC and closed by 13 UTC.

639 Ice radar images provide details of the ice deformation near the MOSAiC CO. Figure 11 depicts
640 radar backscatter intensity which is related to sea ice roughness features. Dark areas in radar
641 images are interpreted as undeformed level ice or leads. High backscatter (bright areas) arises
642 from ridges and edges of leads. Relative motion of these features or their
643 appearance/disappearance between images indicate ice shearing, the formation of leads or ridges,
644 or the closing of leads. Motion is readily apparent in the 15-minute resolution animations of the
645 radar images from 25 January at 00:00 UTC and 5 February 0:00 UTC provided in the
646 Supplement, but in some cases can be discerned in side-by-side comparisons as in Figure 11.
647 Note that the radar is located on the roof of the bridge of the *R/V Polarstern*, located at the apex
648 of the unsampled dark area towards the stern of the ship at the center of each image. All depicted
649 ice motion is relative to the radar.

650 The first row of images illustrates the shearing that occurred between 06:00 and 20:00 UTC on
651 30 January, as cyclone C₁ passed north of the DN. During this time, the group of highlighted
652 bright features to the right of the dashed line moved southward relative to the *R/V Polarstern*
653 (Figure 11c). Most of the shear was concentrated in two regions indicated with pink dashed lines.
654 Both shear zones had been activated at least once in the week prior. During this event, shearing
655 began at the right-most shear zone at 06:00 UTC, then along the left shear zone at 09:50 UTC.
656 This shear zone activated again between 22:00 UTC on 31 January and 00:00 UTC on 1
657 February, corresponding to the peak in shear near 23:00 UTC on 31 January in Figure 10b, and
658 corresponding to the approximate time that the ice motion at the CO and L-sites reversed
659 (Figures 6a-d). These local details corroborate the deformation measured with the DN buoys
660 (Figure S1); the southward motion anomaly is coherent across a region of at least 30 by 30 km.
661 The ice divergence maximum occurring near 06:00 UTC on 1 February in Figure 10a is also
662 apparent in the ice radar data (Figure 11d-f). Starting at 07:00 UTC on 1 February, a fracture
663 activates 1.5 km to the south of the *R/V Polarstern* (green dashed line in Figure 11f), reaching a

664 maximum opening near 10 UTC. Two leads are formed with maximum width of 100-200 m,
665 separated by a shear zone (black patches in Figure 11e). The leads are open only briefly, closing
666 by 12:55 UTC. This time period corresponds to the period of large atmospheric stress (Figure 5f)
667 and the atmospheric horizontal roll vortices (Figure 4h-i). Note that Figure 10a shows ice
668 convergence occurring between 09:00 UTC and 13:00 UTC on 1 February, in excellent
669 agreement with the ice radar observations. These abrupt ice motions are more easily seen in the
670 animation of images at 15 minute resolution between 25 January at 00:00 UTC and 5 February
671 0:00 UTC presented in the Supplement.

672 Together, the drifting buoy data and ice radar data together show a consistent and
673 complementary picture of sea ice response to synoptic and mesoscale features of the atmospheric
674 wind structure. The passage of mesoscale features in the wind field (fronts and LLJs) exerts
675 stress on the sea ice, resulting in deformation. The spatial coherence of the sea ice response is
676 determined both by the scale of the wind forcing and by the geometry of local fracture networks.
677 We therefore can expect that the significant transfer of momentum from the atmosphere into the
678 sea ice results in a strong momentum flux into the upper ocean.

679 **5 Upper ocean response to sea ice motion**

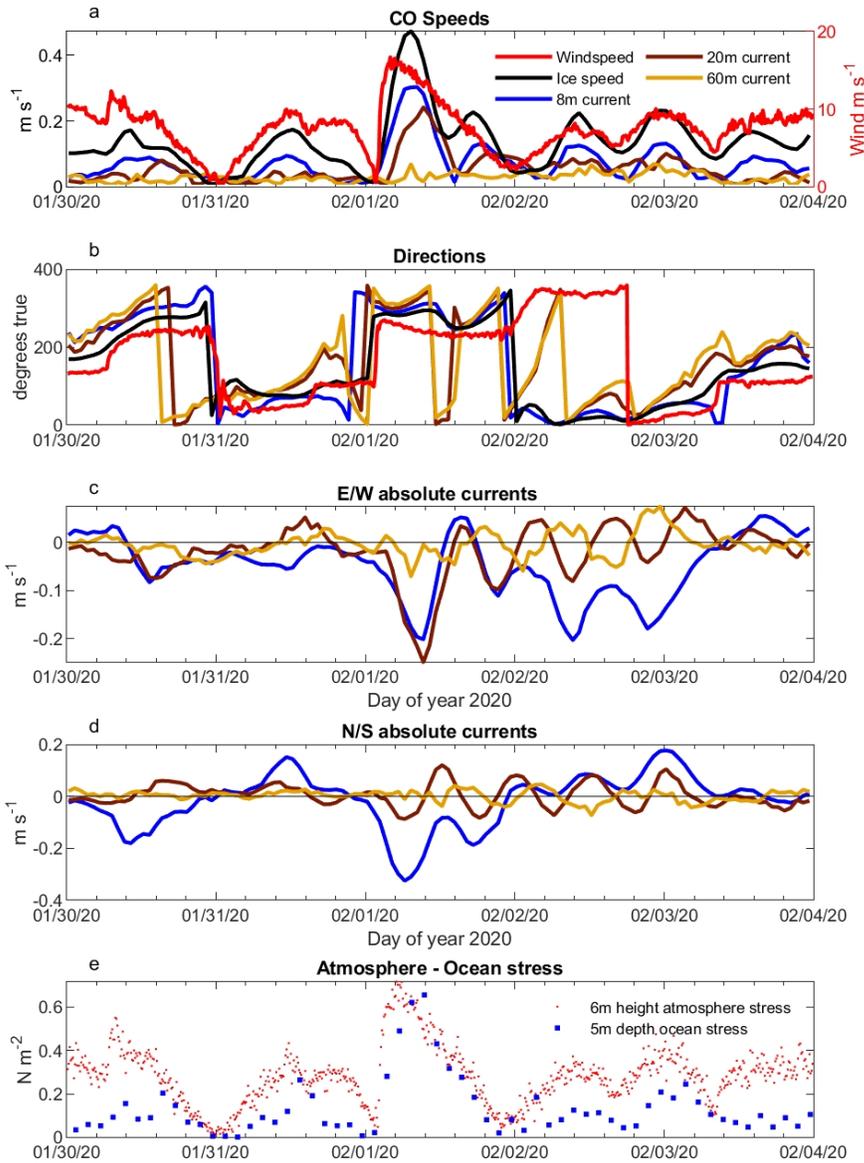
680 Comparisons between the wind, ice and earth-reference current speeds at 8, 20 and 60 m depths
681 (Figure 12a) summarize the transfer of momentum from the atmosphere, to the ice, and then to
682 the ocean. This timeseries is dominated by distinct wind events on 30 and 31 January, and the
683 strong transient event early on 1 February (Figure 12a; see also Figure 5). Each wind event
684 accelerates the ice, which in turn accelerates the ocean layer below the ice as the turbulent ocean
685 Ekman boundary layer forms. This can most clearly be seen in the 1 February event when wind
686 magnitude dropping to near zero, within the annulus of the atmospheric LLJ, followed by an
687 increase to 16 m s^{-1} in the following few hours. A local maximum ice velocity of 0.5 m s^{-1} lags
688 the wind speed peak by 3 hours, while a 0.27 m s^{-1} current speed maximum at 8 m depth lags the
689 ice speed maximum by ~ 1 hour compared to ~ 2 hours at 20m depth. These temporal lags are a
690 result of the inertia of first the surface wind stress accelerating the ice, and then the depth-
691 dependent acceleration of the upper ocean as the ice-ocean turbulent boundary layer deepens in
692 response to changes in direction and magnitude of the ice motion.

693 Rapid changes in ice speed and direction during this event also force significant levels of circular
694 inertial motion in the coupled ice-ocean system. This can be seen in the dampened oscillatory
695 current components in the north-south and east-west velocity timeseries in Figures 12c-d,
696 starting near 1 February at 06:00 UTC and continuing for over 2 days, with the inertial ringing
697 decaying over time. The observed ocean currents represent a superposition of inertial ringing and
698 the evolving boundary layer currents forced by the 1 February 02:00 UTC wind event and
699 subsequent smaller wind maxima at 12:00 UTC on 2 February and 00:00 UTC on 3 February.
700 The inertial ringing is a resonant response to the combination of sharp transient lateral
701 accelerations of the ice/upper ocean coupled with the orthogonal Coriolis acceleration. They are
702 widely observed in the Arctic, with higher magnitudes seen in high open water fraction
703 conditions where ice mobility is enhanced (for example, Brenner et al., 2023). For this event, the
704 8 m depth east-west currents track the ice motion very closely (Figure 12a) with a small phase
705 lag and reduced current component magnitude at 20 m, while the north-south component shows
706 an inertial response from the ice down to at least 20 m but with a larger mean boundary layer
707 current superimposed during 1 February. As expected, there is little direct coupling of this
708 inertial motion at depths below the seasonal (~ 40 m deep) mixed layer as seen in the 60 m depth
709 time series (Figure 12a); the strong density jump at the base of the seasonal mixed layer greatly
710 reduces mixing and hence momentum transfer to greater depths.

711 Comparison between atmospheric surface stress and 4 m ocean stress during this period (Figure
712 12e) shows a deficit on the ocean side of the ice. There are two primary reasons for this
713 difference. The first is the ability of the ice pack to remove surface-imposed momentum through
714 a combination of internal ice stresses and ice deformation. The second is the important role of
715 form drag from the MOSAiC ice pack. The momentum transferred by ice keels and floe edge
716 features is not captured by the friction velocity u_o^* which arises from upstream, small scale
717 roughness features across the ensemble of ice floes and generates the turbulent ocean boundary
718 layer. Lags between the peaks of atmospheric stress and ocean stress, most clearly seen in the
719 strong 1 February event, arises from the inertial lag of the ice pack to surface wind stress (Figure
720 12e).

721 The vertical structure of upper ocean currents in response to this wind event (Figure 13)
722 illustrates the fairly complex interaction of the ice/ocean boundary layer with weakly stratified
723 mesoscale ocean features within the seasonal mixed layer, which were seen during much of the

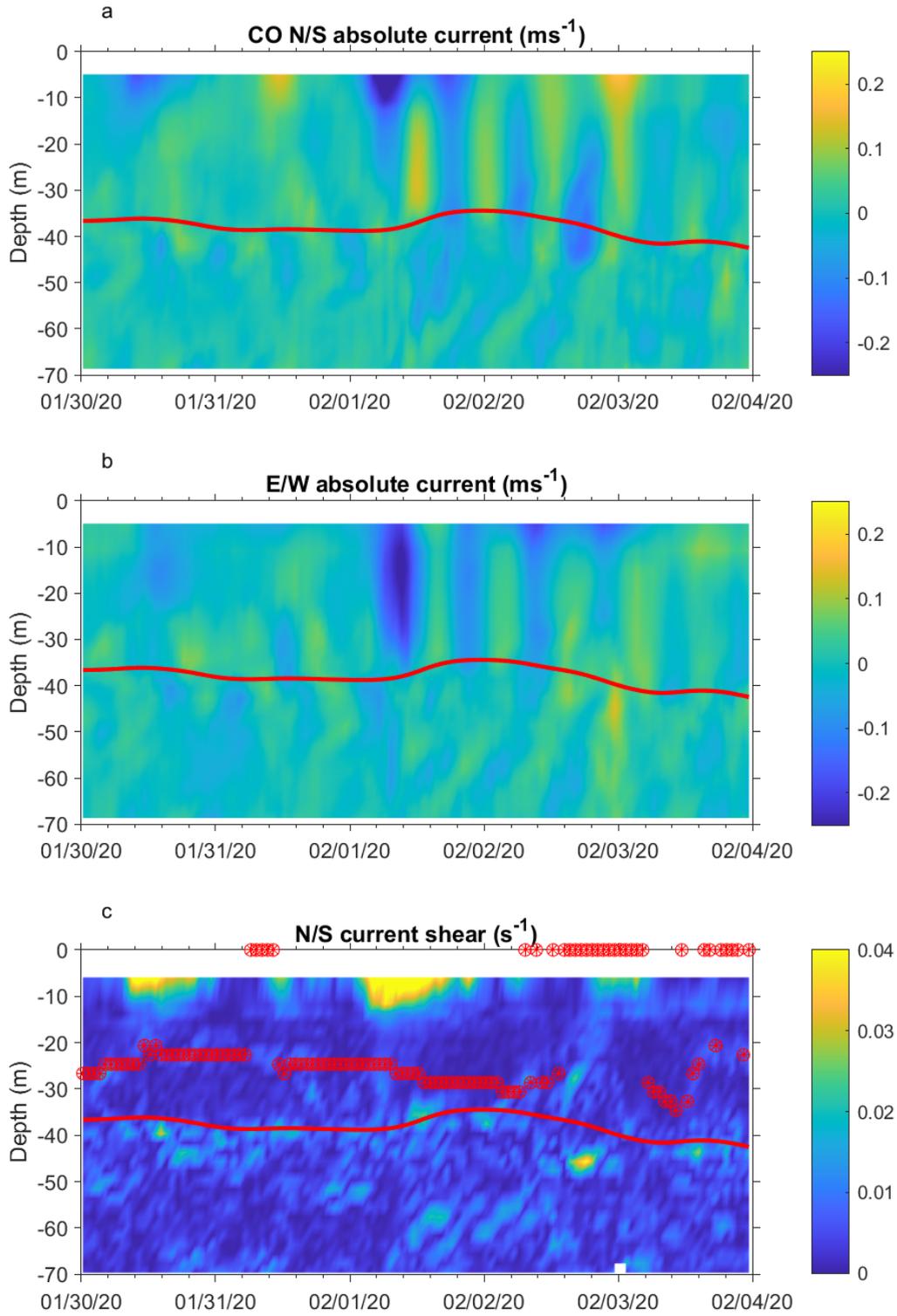
724 MOSAiC transpolar drift. High temporal resolution vertical shear of the N/S current component
725 sampled every 15 minutes by the AOFB current profiler at the CO (Figure 13c) provides some
726 insight into the complex structure of the active mixing layer. CTD profiles were limited by wind
727 conditions and as such are sometimes only available once per day. Ideally, CTD profiles at a
728 comparable temporal resolution to the AOFB current profiler would show the evolution of
729 stratification within the mixed layer, which frequently contained weak mesoscale density
730 structures limiting the depth of mixing during wind events. However, the much higher resolution
731 shear profile time series in Figure 13c reveal both the development of strong near-surface shear
732 as the sub-ice Ekman layer forms, and the development of regions of higher shear within the
733 ~40m deep seasonal mixed layer. These shear layers indicate the lower extent of the mixing layer
734 where even weak density gradients inhibit turbulent mixing deeper within the seasonal mixed
735 layer. Measurements of these weak stratification layers are estimated from the depths where
736 there is a density increase of 0.01 kg m^{-3} from surface values for each CTD profile, and are
737 plotted as filled red circles in Figure 13c. These sparse-in-time observations coincide with the
738 layers of increased shear measured in the current profiles. The red mixed layer depth timeseries
739 in the Figure 13 panels represent coarse interpolated estimates of the depth of the top of the
740 halocline.



741

742 **Figure 12.** From top to bottom: (a) Timeseries of windspeed (red), ice speed (black), 5 m (blue)
 743 20 m (green) and 60 m (orange) depth absolute current magnitude; (b) Corresponding current
 744 and wind directions in degrees true; (c) Timeseries of 5 m (blue), 20 m (brown) and 60 m (gold)
 745 north-south current components; d) east-west current components; e) 4 m depth ocean kinematic

746 stress from the CO site Autonomous Ocean Flux Buoy (blue dots) and atmospheric stress (red
747 dots) for this study period.



749 **Figure 13.** North-south current profile timeseries from the L1 site Autonomous Ocean Flux
750 Buoy acoustic Doppler profiler. b) Corresponding East-West current component profiles. c)
751 North/South current shear profiles with a clipped color scale to emphasize shear layers within the
752 ocean mixed layer and upper part of the salinity-stratified pycnocline. Near surface shear reaches
753 0.07 s^{-1} during the 1 February wind event. The continuous red line represents an estimate of the
754 depth of the top of the halocline. The sloping orange line highlights the rapid penetration of
755 mixing in response to this wind event. The four black sloping lines identify shear associated with
756 inertial internal waves within the strongly stratified pycnocline forced by the strong inertial
757 motions within the ocean mixed layer.

758 The highest vertical shear levels of the north/south current component (Figure 13c) are seen in
759 the upper 15 m during the strong 1 February wind event. However, active mixing extends down
760 through the seasonal mixed layer to the halocline seen most clearly in the E/W current profile
761 (Figure 13b) and the elimination of shear at the weak stratification interface between 20 m depth
762 and the halocline (marked by the orange line in Figure 13c). An example of reduced mixing
763 depth by a mesoscale feature at the base of the mixed layer is seen starting at 12:00 UTC on 1
764 February despite the continued strong surface forcing. The CO drifts over another weak
765 stratification feature that extends up from the pycnocline as winds reduce to 3 m s^{-1} at 00:00
766 UTC on 2 February. The interplay between surface mixing and these frequent mesoscale features
767 with a wide range of density gradient strengths observed during the MOSAiC drift complicates a
768 1D view of wind-forced turbulent momentum transfer into the ocean. Analysis of these
769 mesoscale features is beyond the scope of the present paper, and will be explored in subsequent
770 publications.

771 Strong inertial-period motions in the ocean mixed layer are capable of generating internal
772 inertial-period waves within the pycnocline after the mixing layer inertial currents contact the
773 strongly salinity-stratified pycnocline. In the current component profiles (Figures 13a and 13b)
774 this can be seen as slanted bands of enhanced current shear with inertial periods starting around
775 45 m depth after the 1 February wind event. These regions of enhanced shear are also shown as
776 black slanting lines in Figure 13c. These inertial waves are an important source of shear that can
777 induce mixing in the otherwise very quiescent and non-diffusive Arctic pycnocline.

778 **6 Discussion and conclusions**

779 We presented a detailed description of an observed, strong, mid-winter, central Arctic cyclone
780 which passed over the MOSAiC observatory from 31 January to 1 February 2020, closely
781 following the passage of a weaker cyclone. This cyclone produced air-ice-ocean changes,
782 including leading to the development and passage of a strong quasi-axisymmetric low-level jet
783 (LLJ) in the lower atmosphere, producing widespread sea ice deformation, and propagating
784 momentum flux into the upper ocean. The comprehensive suite of MOSAiC instruments together
785 provides unique observations of the coupled air-ice-ocean system during an evolving cyclone
786 with unprecedented detail and spatial resolution.

787 The developing atmospheric LLJ, which eventually appears as an annulus of ~ 140 km radius
788 around the low-pressure center, is the key atmospheric feature of this cyclone impacting the
789 momentum transfer to the sea ice. A smaller jet core within this LLJ is identified in the cold
790 sector of the ERA5 reanalyses between 00 and 12 UTC on 1 February, and is linked to observed
791 faster ice motion as well as shearing and divergence of the sea ice. The stage of storm
792 development and the spatial structure of the LLJ strongly impacted the timing and location of sea
793 ice deformation. The elevated surface wind speeds ahead of the cyclone produced an increase in
794 drift speed and resulted in ice shear. The developed LLJ behind the cold front produced strong
795 deformation in the ice, with divergence ahead of the jet core and convergence behind. This
796 produced opening and closing of leads, respectively. Local sea-ice trajectories are a function of
797 distance to the storm track and the side of the track. The sudden change in wind and ice-drift
798 direction and the rapid increase in sea-ice velocity with the arrival of the cold-sector LLJ and its
799 core produced a jump in the air-ice and ice-ocean stresses. The local destabilization of the lower
800 atmosphere behind the cold front contributed to the former, while the latter initiated an inertial
801 oscillation in the sea ice and upper ocean. The observations also showed that the change in ice-
802 drift direction occurred locally in the DN about 2 h prior to the change in wind direction with the
803 cold front, suggesting that wind forcing of the ice behind the cold front propagated ahead of the
804 front through the internal ice stress. Hence, wind forcing of ice acceleration may not always
805 occur locally.

806 The initiation of the inertial oscillation in the ocean extended the impacts of the storm beyond the
807 time taken for the atmospheric depression to fully cross the observatory. A second increase in sea

808 ice strain rates 12 hours after the arrival of the LLJ occurred due to the differing timescales
809 between the atmosphere and the coupled ice-ocean boundary layer during the inertial oscillation
810 and the gradual change in the wind direction. The ice and near-surface ocean returned to
811 following the wind after approximately 24 hours, while at depth, the effects of the inertial
812 oscillation were visible for at least 3 days.

813 Because of the apparent importance of the LLJ and the LLJ core for air-ice interactions, it must
814 be noted that there is some uncertainty in its spatial and temporal structure. Since it was only
815 directly observed by the 1 February 06:00 UTC sounding, and temporally and spatially spread by
816 the ERA5 data assimilation, there could have been other LLJ cores or this core could have been
817 present before 1 February 00:00 UTC. However, no atmospheric or ice observations suggest this
818 to be the case. Furthermore, the structure and strength of the LLJ in the warm sector is also not
819 well observed, as the 31 January 18:00 UTC sounding only captures the inner edge of the LLJ
820 annulus at a time when the axisymmetric characteristic has not yet developed (Figure 3a). Hence,
821 to describe the LLJ structure we use ERA5 to fill time and space between observations. Finally,
822 LLJs have not been a part of the classical conceptual models of Arctic cyclones (e.g., Aizawa &
823 Tanaka, 2016), likely due to these studies relying on reanalyses with a resolution incapable of
824 resolving this mesoscale feature. More recent Arctic cyclone structure studies using ERA5 (e.g.,
825 Vessey et al., 2022), have mentioned the presence of strong low-level winds in the warm sector,
826 however.

827 The breadth of observation types available through the MOSAiC observatory provides
828 opportunity for numerical model evaluation and development, enabling examination of multi-
829 scale, strongly coupled processes. While numerous case studies of cyclones exist, most focus on
830 the summer and the marginal ice zone. Few observations are available for the central Arctic in
831 full pack ice during mid-winter. We have identified key processes for the transfer of energy from
832 atmosphere to sea ice to the upper ocean. A companion study will examine the representation of
833 these processes in modern coupled air-ice-ocean numerical weather forecast models.

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849

850 **Open Research**

851 Atmospheric and ice drift data used in this paper are archived at the Arctic Data Center (Sea ice
852 buoys: Bliss et al., 2022; atmospheric data: Cox et al., (2023a-d) and in the Alfred Wegner
853 Institute PANGAEA archive (Maturilli et al., 2022). Atmospheric Ka-band radar is archived at
854 the Department of Energy Atmospheric Radiation Measurement User Facility (Bharadwaj et al.,
855 2019; Hardin et al., 2019). Ice radar data is archived in the Alfred Wegner Institute PANGAEA
856 archive (Krumpfen et al., 2021a). Data from the Autonomous Ocean Flux Buoy is archived at the
857 Arctic Data Center (Stanton & Shaw, 2023). Code supporting the data analysis and visualization
858 is archived at Zenodo (<https://doi.org/10.5281/zenodo.10698905>).

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