



Scale invariance in kilometer-scale sea ice deformation

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Abstract.

Large-scale modeling of sea ice dynamics assumes scale-invariance that is used to calibrate and validate current models. Validity of this assumption, particularly its lower spatial limit, remains poorly understood. Identifying when, where, and why scale-invariance does not apply is essential for linking meter-scale sea ice mechanics with large-scale sea ice dynamics and climate models. Here we address this challenge by employing unique high-resolution ship radar imagery from MOSAiC expedition in an analysis based on novel deep learning-based optical flow technique. Together these allow capturing sea ice kinematics consistently at unprecedented 20-meter spatial and 10-minute temporal resolutions over an entire winter season and into summer over a 10-kilometer spatial domain. We show that the sea ice within this domain remains largely quiescent for extended periods, with distinct events revealing a 10²-meter lower limit for scale-invariance that endures as the ice cover undergoes seasonal evolution. This threshold remains stable throughout the winter, even as deformation features become more localized and distinct, suggesting an intrinsic mechanical constraint that is invariant under varying external conditions. Once the ice transitions to a floe-dominated configuration in summer, no comparable scaling signature emerges. Our results give a limit under which continuum models fail to capture critical fine-scale processes, highlighting the need for approaches accounting for detailed description of discontinuous spatial and temporal behavior of sea ice.

1 Introduction

Since the beginning of geophysical-scale numerical modeling of sea ice, rheological models have been a topic of active discussion. Rheological models describe how sea ice moves, fragments, and deforms under external forces caused mainly by winds, currents, and waves (Leppäranta, 2011). Given the direct connection between the future of sea ice and climate, accurate rheological models are essential for reliable climate change projections. The currently available computational power allows continuum models to simulate sea ice dynamics at very high resolutions, yet it remains an open question whether the continuum models calibrated on fractal properties are valid for relatively small sea ice domains (Hutchings et al., 2024), and what is the lower limit of scale invariance in sea ice dynamics (Weiss, 2017). As the sea ice cover is becoming thinner, more fragmented, and transitions to younger and more mobile ice (e.g. Rothrock et al., 2008; Rampal et al., 2009; Kwok et al., 2013), understanding this lower limit is becoming increasingly critical. A more accurate rheological description able to account for



25 small-scale processes will be needed to predict future ice conditions with mechanical weakening and increased fracturing. Our work observes this limit and discusses how this limit connects to physical phenomena observed in sea ice.

Our work is based on novel observations of a 10 km × 10 km area of sea ice over a nine-month period. We use unique ship radar observations of ice deformation collected during the Multidisciplinary drifting Observatory for the Study of Arctic Climate (MOSAiC) expedition (Nicolaus et al., 2022). Sea ice deformation is rarely studied at these scales. Conventionally, 30 studies focus on large-scale sea ice deformation observed by using buoy or satellite data (Hibler et al., 1973, 1974; Fily and Rothrock, 1986; Rampal et al., 2008; Hutchings et al., 2011; Marsan et al., 2004; Rampal et al., 2019; Bouchat et al., 2022) or on meters-scale ice failure processes related to ice engineering applications or detailed sea ice mechanics (Sanderson, 1988; Hopkins et al., 1999). Only recent studies have begun to fully explore sub-kilometer intermediate-scale sea ice deformations (Oikkonen et al., 2017; Hutchings et al., 2024), even if detailed insight on ice behavior on this scale is essential for linking 35 meter-scale sea ice mechanics with large-scale sea ice dynamics (Feltham, 2008; Weiss, 2017; Weiss and Dansereau, 2017).

We describe the general dynamics of the studied sea ice area over one season, focus on the most active periods of ice deformation, and discuss the scale invariance in sea ice deformation. The latter is the key for discussion on the limit for continuum models for ice and often considered a key for understanding sea ice deformation across scales (Schulson, 2004). This information is essential for understanding sea ice dynamics and for model development (Leppäranta, 2011; Weiss, 2013, 2017). 40 Previous studies have indicated that, although larger deformation occur during storms, sea ice is rarely completely inactive at intermediate-scales (Oikkonen et al., 2017) and that ice deformation shows spatial and temporal localization, which holds over a wide range of scales follow power law form scaling (Marsan et al., 2004; Rampal et al., 2008; Stern and Lindsay, 2009; Marsan and Weiss, 2010). It is assumed that a rheological model, linking stresses to strains, can be developed for the range of scales where ice deformation remains scale invariant. We employ a novel deep-learning-based optical flow tool (Uusinoka et al., 45 2025), which allows full-field analysis of sea ice deformation at 20-meter resolution. The high-resolution data enables us to capture fine-scale deformation features and intermittent behaviors previously unavailable, since conventional methods lead to coarser and less accurate data. High-resolution data is crucial with the emergence of the numerical models that allow simulating hundreds-of-kilometers-scale sea ice domains with meter-scale resolution (Åström et al., 2024; Åström and Polojärvi, 2024).

We present three key findings that are all important to account for in modeling of sea ice: (1) the 10 km × 10 km sea ice area 50 we studied shows three major and several smaller deformation periods occurring during the winter; (2) the lower limit for the scale invariance for the studied area was of magnitude 10^2 m; and (3) this limit remained constant for the deformation events until the spring, after which it could not be defined due to the breakup of the ice pack. (1) differs from the observations of Oikkonen et al. (2017) as their results indicate deformation activity at the intermediate scale to be constantly active although at varying magnitudes. However, their area of measurement was closer to Marginal Ice Zone than that of MOSAiC central 55 observatory. Instead, we find that the deformation events are strongly intermittent and observe multiple periods during the winter season characterized by little activity. (2) is noteworthy, as the limit for scale invariance in sea ice deformation has been discussed for over two decades (Hibler and Schulson, 2000; Schulson, 2004), with authors hypothesizing it might apply at scales comparable to ice thickness (Weiss, 2013, 2017). While the here observed 10^2 m limit is larger than this, it is also only about one-tenth of the 10^3 m limit suggested by Hutchings et al. (2024). We also find that the 10^2 m limit is independent of the



60 temporal scale, while the magnitude of error in scale invariance increases with spatial and temporal scale, indicating a complex
 61 interplay between external forcing and ice deformation. (3) reflects the change in scale invariance due to seasonal variation in
 62 the ice field dynamics, when it transitions from pack ice to floe ice. We are able to capture the lower limit and the seasonal
 63 transition, whereas earlier work extending to an average spatial scale of 250 m did not (Oikkonen et al., 2017).

64 The paper is structured as follows. Section 2 introduces the ship radar data and the tools used in the analysis. Section 3
 65 describes the nine-month-long deformation rate record and highlights the intermittency of the ice deformation events. Then
 66 it describes the major deformation events in detail and performs mean-based deformation analysis. Section 4 discusses the
 67 implications of our results and their interpretation before Section 5 concludes the paper.

2 Methods

2.1 Data

70 Figure 1 shows examples of the radar images used as the basis of our analysis. These were collected by using the radar signal
 71 digitizing system similar to Oikkonen et al. (2017) deployed on RV Polarstern as part of the sea ice measurement activities
 72 during MOSAiC (Nicolaus et al., 2022). The system continuously collected images of pack ice from mid-October 2019 until
 73 the end of September 2020. The original data set included radar images captured at 2.4 s intervals, covering a 20 km×20
 74 km area with a 8.33 m resolution with their orientation aligning with the bow of the ship. The quality of the backscattered
 75 radar signal considerably decreased beyond 5 km from the ship, for which we cropped the images to 10 km × 10 km to ensure
 76 consistent estimation quality over the entire image.

77 The digitized images contained interference from other ship and helicopter radars, which was removed by stacking three
 78 consecutive images and retaining the minimum pixel value for each point. Faulty images were identified and removed based
 79 on missing azimuth pulses, spurious rotation-symmetric patterns, incomplete rasterization, or narrow shadowed sectors. The
 80 filtered data set has a 1-minute temporal resolution. Despite filtering, the data still contained high-frequency noise concentrated

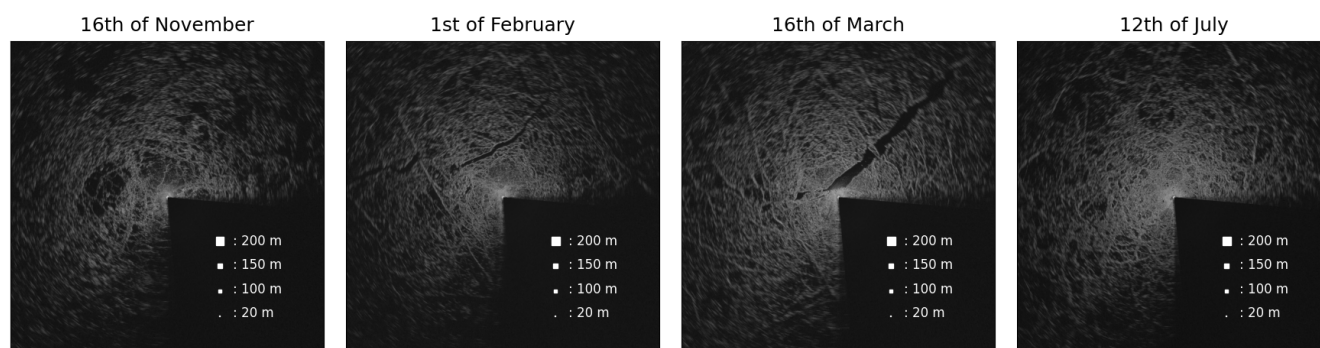


Figure 1. Radar images of the 10 km × 10 km sea ice area. Light areas are deformed ice and dark areas level ice and leads. The white squares indicate the different spatial scales, L , analyzed below.



in specific areas of the images. This noise was at least partly due to actual physical phenomena, such as minor vertical displacements caused by ocean swell, small-scale ice deformation, or radar vibrations due to strong winds. The noise manifested as abrupt, high-intensity signal variations, making it particularly challenging to extract consistent displacement estimates from the images. We performed spatial averaging of the images through bilinear interpolation to reduce the presence of artificial noise in the radar data resulting in the final spatial resolution of 10 m.

2.2 Recurrent neural network-based optical flow

The ice motion was estimated from the ship radar imagery by using a novel deep-learning-based optical flow tool described in detail by Uusinoka et al. (2025). The tool is built on Recurrent All-pairs Field Transforms (RAFT) optical flow architecture (Teed and Deng, 2020), enhanced with a temporal multi-resolution tree for increased accuracy in small pixel displacements. We found this tool to overcome the typical challenges involved in the analysis of the radar data while allowing high-resolution full-field analysis of ice motion by providing $\sim 750\,000$ trajectories for each radar image. To further ensure the robustness against the high-frequency noise in the data, described in Uusinoka et al. (2025), we chose 10-minute intervals between sequential images for sufficient displacements. We use 24-hour trajectories to avoid data loss due to artificial rotations in the radar images.

2.3 Strains and spatiotemporal scaling

We calculate deformation estimates using quadrilateral cells in accordance to Bouchat and Tremblay (2020) to avoid overestimation of deformation rates due to increased boundary-definition errors (Lindsay and Stern, 2003; Bouillon and Rampal, 2015). Instead of assuming infinitesimal strains, commonly used in sea ice dynamics, we apply Green-Lagrange strains (Appendix A). This choice was made to account for large strains and rigid body rotations; infinitesimal strains are prone to yield erroneous results when these occur (Gurtin, 1982). Strain estimates for large-scale ice drift are around 9 % over 24-hour-period and often higher at smaller scales due to localized ice deformation (Leppäranta and Hibler III, 1987; Hibler et al., 1973). We derive the estimates from finite strains to rotation-fixed infinitesimal strains, l , to relate our results to prior research (Derived in Appendix A).

We use total deformation rate, $\dot{E}_t$, to perform spatiotemporal scaling analysis on the ice cover. Localization and intermittency of sea ice deformation has been shown to follow the mean-based power-law forms (Marsan et al., 2004; Rampal et al., 2008)

$$\langle \dot{E}_t(L, T) \rangle \sim L^{-\beta} \quad \text{and} \quad \langle \dot{E}_t(L, T) \rangle \sim T^{-\alpha}, \quad (1)$$

where $\langle \dot{E} \rangle$ is the mean total deformation rate over the entire field, L and T are the spatial and temporal scales, respectively, and the exponents β and α represent the corresponding spatial and temporal scaling. β quantifies the degree of spatial localization, ranging from $\beta = 0$ (perfectly homogeneous viscous-like deformations) to $\beta = 2$ (deformation localized into a single point). Similarly, α quantifies the degree of temporal localization, ranging from 0 (continuous flow) to 1 (a single 'event').

Our scaling analysis uses the dispersion method (Rampal et al., 2008; Oikkonen et al., 2017; Rampal et al., 2019), where we extract velocity gradients with increasing spatial and temporal intervals corresponding to specific nominal scales. This extraction is based on the initial coordinates of the corresponding trajectories and involves estimating deformation rates over



larger scales to highlight broader patterns. The nominal spatial and temporal scales for computing deformation rates were $L = [20, 50, 100, 200, 300, 500, 650, 800]$ m and $T = [10 \text{ min}, 30 \text{ min}, 60 \text{ min}, 3 \text{ hours}, 6 \text{ hours}, 12 \text{ hours}]$, respectively. It is
 115 important that averaged spatial scales below 250 m are considered, as Oikkonen et al. (2017) demonstrate the scale-invariance extending down to this scale.

3 Results

3.1 Ice cover deformation

Figure 2 shows deformation rate records for the $10 \text{ km} \times 10 \text{ km}$ sea ice area around the MOSAiC central observatory. These
 120 records were derived using a 10-minute temporal resolution and a 20-meter spatial resolution and they cover data from November 2019 to April 2020. The time series is extended with deformation estimates from July 2020 to include summer conditions in the marginal ice zone (MIZ).

Figure 2 shows that for the most of the winter, the considered area of sea ice experienced only vanishingly small deformation rates. The recurring and extended periods of inactivity are noteworthy, especially the three-week period in February. Only
 125 minor shearing could be detected during the quiescent periods. Visual inspection of the radar imagery shows that this low-

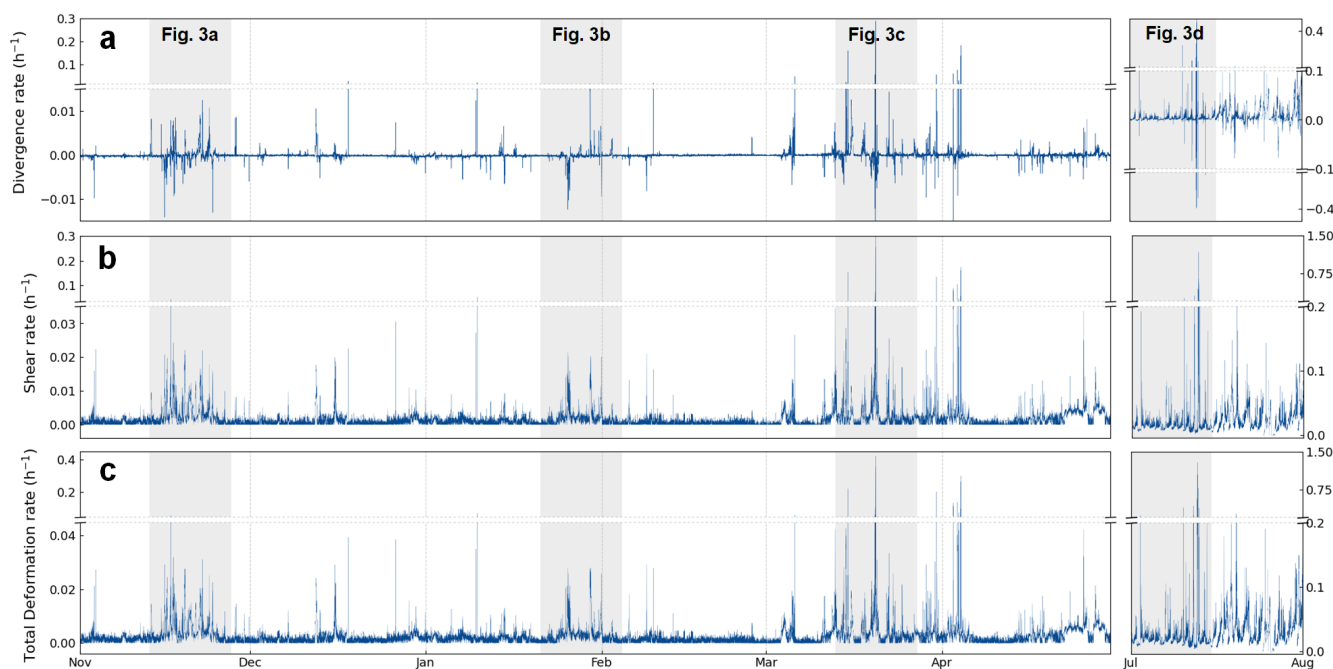


Figure 2. Time series of a) divergence rate, $\dot{E}_d$, b) shear rate, $\dot{E}_s$, and c) total deformation rate, $\dot{E}_t$, calculated with a 10-minute temporal and 20-meter spatial resolution as an average over the deformation field. Note that the y-axis limits are adjusted individually for each subplot to highlight the distinct behavior of the different deformation components and periods.



130 magnitude shearing resulted from a mixture of subtle ice displacements, rather than significant changes in the ice cover. The high-frequency noise in the radar data may also contribute partially to the low-magnitude shear observed. Further, the figure shows several intermittent deformation events that occurred during three periods of major ice deformation. These periods are highlighted in gray in the figure and occurred in November 2019, January-February 2020, and mid-March 2020. The last one of these is noted for exceptionally high mean deformation rates approaching 0.4 h^{-1} . During these three periods, the ice deformed rapidly in various modes, including events with all deformation occurring within few-hour periods. Inactive periods are not seen

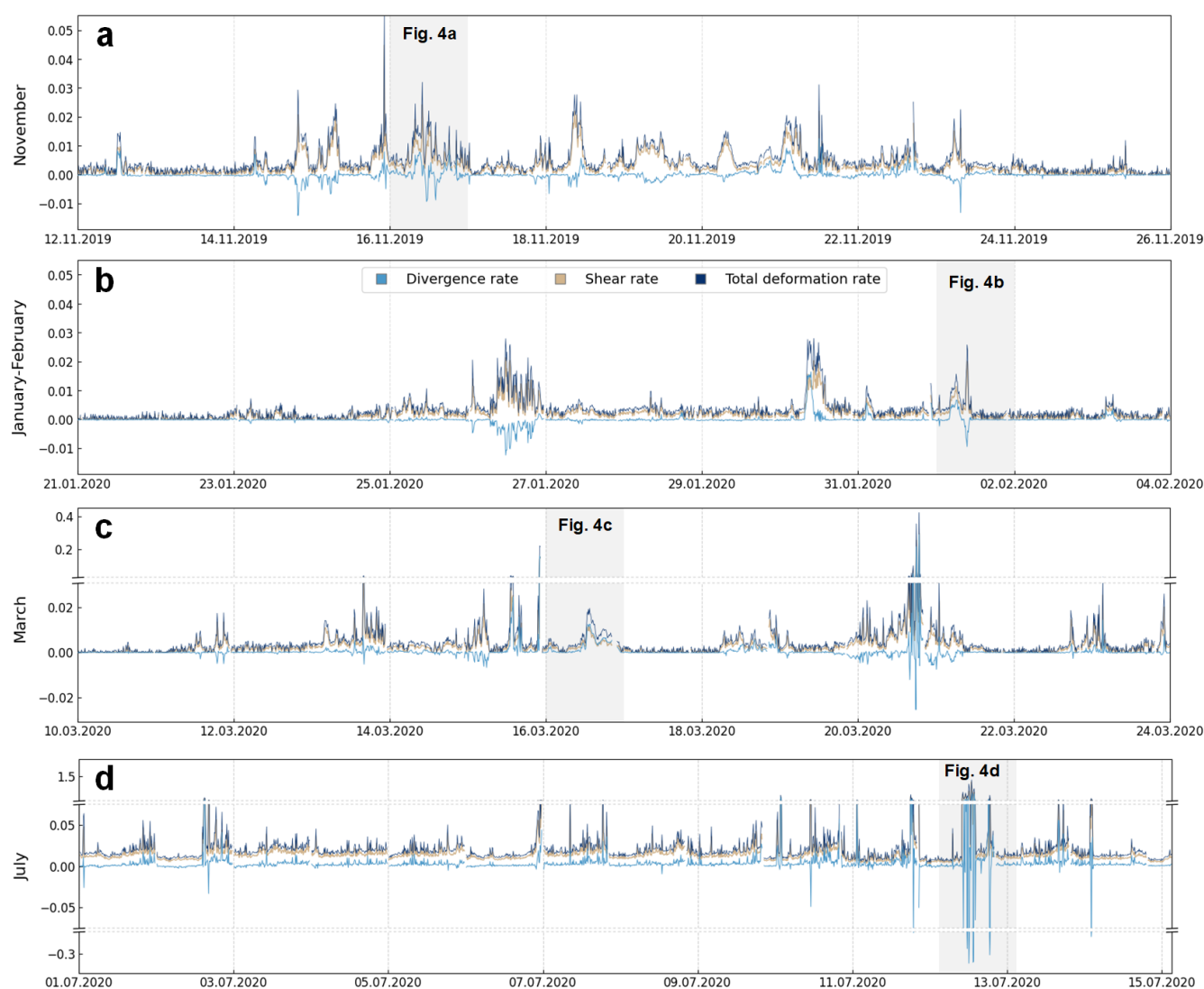


Figure 3. Time series of the three deformation rate components during the three active periods highlighted in Figure 2. Additional times series of the first two weeks of July is provided for comparison between different systems.



in July data, which describes ice deformation within the MIZ. In this case the time series describes a significantly different system with practically continuous ice deformation. The deformation rates increase by an order of magnitude compared to winter, and the peak values are much higher than those reported for similar conditions by Oikkonen et al. (2017).

Figures 3a-c provide a more detailed view of three active two-week winter time periods of Figure 2. The three time series exhibit a consistent intermittency similar to that described above for the seasonal analysis. Increased intermittency is observed during the cases of January-February and March (Figure 3b-c) in relation to the observations of Oikkonen et al. (2017), whose results indicated more continuous deformation in conditions closer to the MIZ. For November (Figure 3a), the data shows several deformation events and brief inactive periods, likely due to the low load-bearing capacity of the relatively thin ice (Itkin et al., 2023). In contrast, the active period in February (Figure 3b) is characterized by few intermittent deformation events and long inactive phases, possibly owing to the thickened ice cover. Data for March (Figure 3c) depicts a blend of patterns observed during the preceding months and it is characterized by significant deformations, which suggests a shift from the winter towards the spring. In July (Figure 3d), within the MIZ, ice undergoes constant deformations with few extreme peak deformation events, which in this case signify a uniformly intense deformation across the area studied.

The deformation fields displayed similar seasonal characteristics, as shown by Figure 4, which visualizes the mean total deformation rate over the 24-hour periods highlighted in Figure 3. The November case (Figure 4a) has the deformation distributed across several larger scale fractures with numerous smaller deformation features emerging from them and indicating localized activity around them. The February case (Figure 4b) has the deformation concentrated along a line-like lead, with smaller perpendicular deformation features appearing, indicating secondary failure processes. By March (Figure 4c), nearly all activity becomes localized along a single quasi-linear fracture, representing a large lead. This main feature is flanked by smaller regions where ice fragments collide and form a complex interaction network. Deformation field for July (Figure 4d)

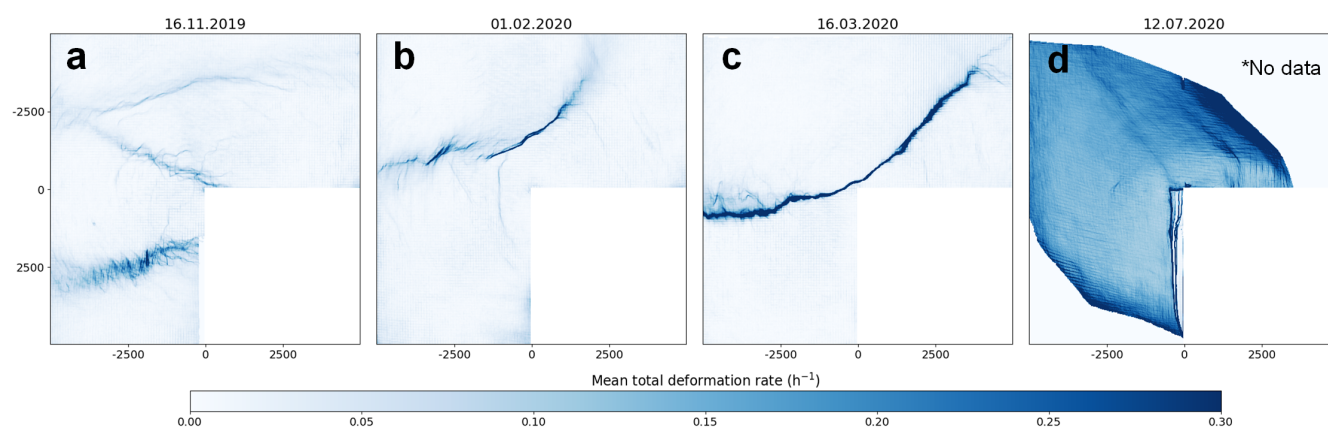


Figure 4. Total deformation (Eq. A5) fields for the 24-hour periods highlighted in Figure 3. Each field show mean deformation for 20 m $\times$ 20 m cells during the period considered. Features exiting the radar images have been neglected and no filtering has been used. In d) all trajectories exiting the radar image during the 24-hour period have been removed.

is markedly different from the rest, as the entire area studied undergoes viscous-like deformation as it is located in the MIZ during summer conditions.

3.2 Spatiotemporal scaling and limit of scale in variance

155 We conducted a spatiotemporal power-law analysis using the mean total deformation rates of the four 2-week periods highlighted in Figure 2. Figure 5a depicts how the scaling exponent β behaved in the case of observations made over various spatial

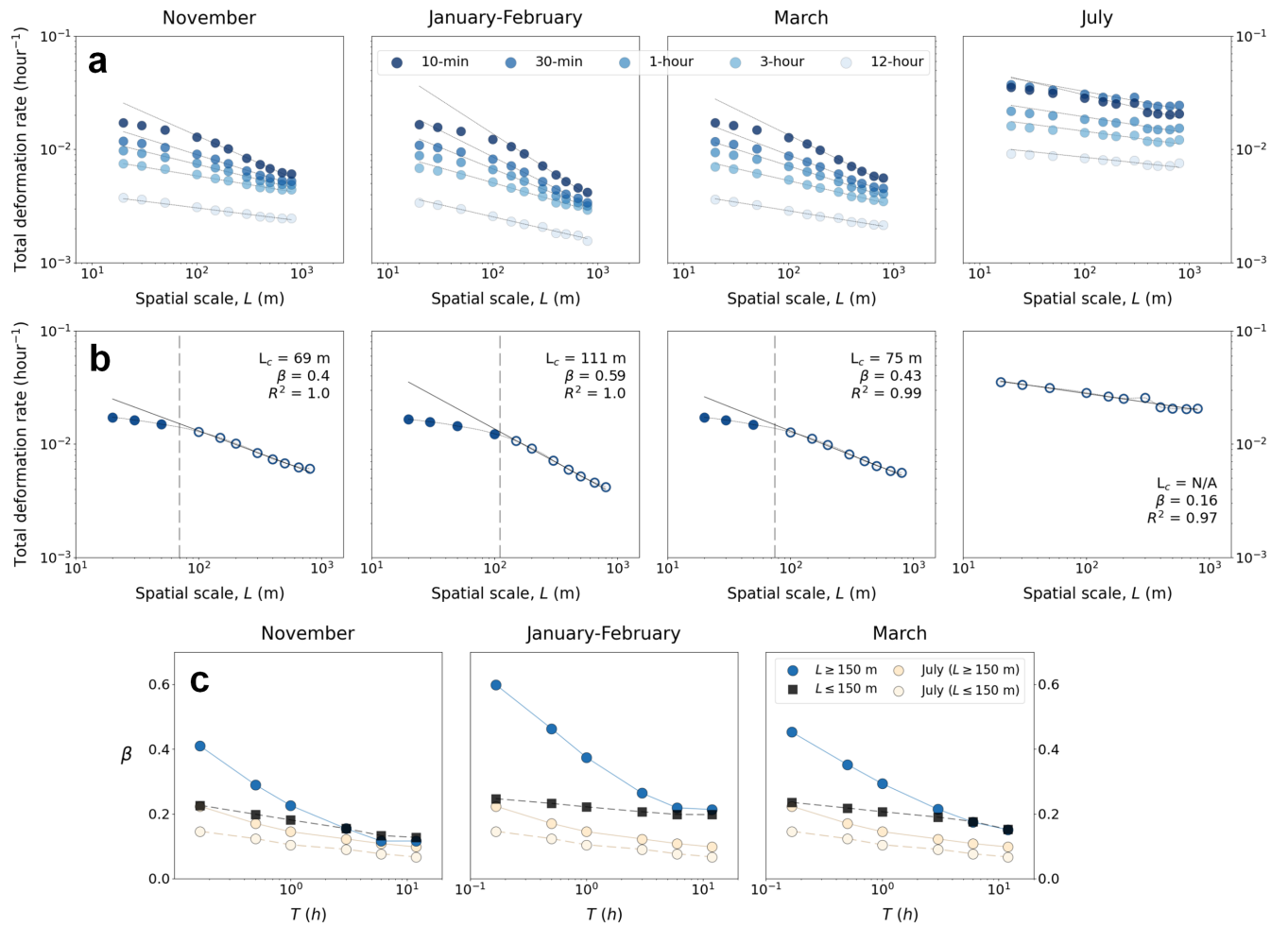


Figure 5. a) Mean based spatial scaling properties of total deformation rate (Eq. A5) during the four periods highlighted in Figure 2. The linear fits are for the data with length scale $L \geq 150$ m. b) Limit for the scale invariance for the four two-week periods of Figure 3. Coefficient of determination R^2 and scaling coefficient β are for the fit applied on the data points that follow the mean-based scaling (marked with white dots). c) The development of the spatial scaling exponents, β , over different temporal scales. The development of β is considered individually for both $L \geq 150$ m and $L \leq 150$ m in comparison with the July case.



and temporal scales, L and T , respectively. Figure 5c shows β as function of T for data with $L \leq 150$ m and $L \geq 150$ m in comparison to the July case. In Figure 6, similar analysis is performed for the temporal scaling exponent, α , as a function of L to explore similar limitations of scale invariance in the temporal domain. In the analysis, we use the July case as a reference point for when the continuum approach and fractal analysis are appropriate, given its relatively uniform and diffuse deformation field (Figure 4d). The July case helps in illustrating how the fractal properties diminish when a continuum perspective begins to break down.

Figure 5 shows that β varies between 0.4 and 0.6 for the winter cases when $L \geq 150$ m and $T \geq 10$ min. The β values align well with the observations of increasing β towards smaller spatial scales in Hutchings et al. (2024), although they are about $\sim 25\%$ smaller than those observed by Oikkonen et al. (2017). Further, the β values decrease with increasing T in all winter cases as expected. For data $L \geq 150$ m, the two-week period in January showed the highest values of β , even if this period showed the lowest peak values for total deformation rates (Figures 3). Additionally, the November and March periods exhibit nearly identical scaling behavior, even if their deformation time series and deformation fields are not alike (Figures 3 and 4). Values of β for July data are low, which is expected based on the deformation field in Figure 4d.

For scales $L \leq 150$ m, a notable shift in system behavior occurs, with the values of β being about constant for all T used in the analysis. At the smaller spatial scales, mean strain rates are similar across different time scales, suggesting a dominance

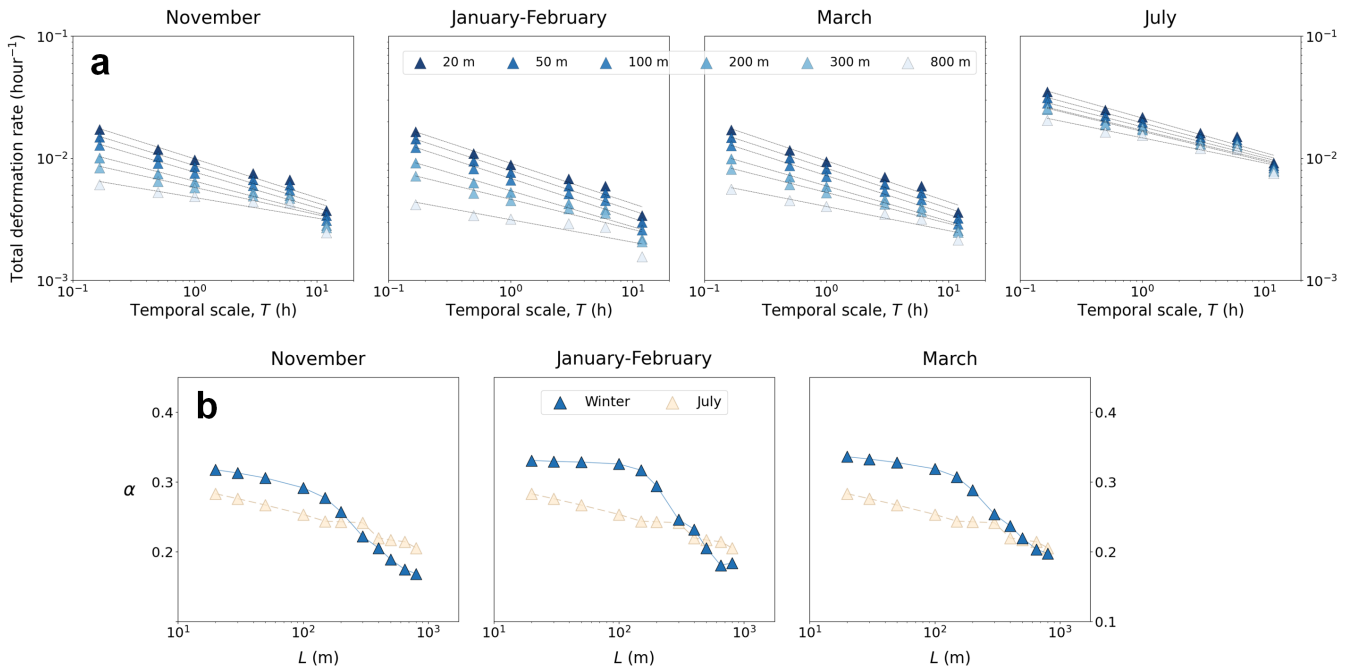


Figure 6. a) Mean based temporal scaling properties of total deformation rate similarly to Figure 5. The linear fits over for all the presented temporal scales. b) The development of the temporal scaling exponents, α , over different spatial scales. The development of α is presented in comparison with the July case.



of the largest deformation features in the overall deformation field. Figure 1 illustrates scales $L \leq 150$ m, which are often less than the typical width of deformation features such as leads. At larger scales of $L \geq 150$ m, β stabilizes around 0.2 for all winter cases around $T \approx 1 \dots 3$ h, which is the value of β for scales for the scales $L \leq 150$ m for all T . Data for July does not show a scale-dependent shift as $\beta \approx 0.1$ to 0.2 for all T for both $L \leq 150$ m and $L \geq 150$ m. For $L \leq 150$ m, the values of β in the winter and the July case exhibit comparable magnitudes and similar behavior, suggesting that at sufficiently small spatial scales (or, as seen in Fig. 5a, at longer averaging times) with the given spatial domain size of $10 \text{ km} \times 10 \text{ km}$, the deformation processes tend to loose the typically observed fractal properties.

From the above, it is clear that the mean-based scaling does not apply for all L , as data for $L \leq 150$ m behave differently than those for the larger scales. Therefore we looked for a critical length scale, L_c , for the scale invariance, using the data for the case in Figure 5a with a 10-minute temporal resolution. Figure 5b illustrates how this was done. We performed a least-squares linear fit to the data points $\{(L_j, \dot{\epsilon}_j)\}_{j=1}^n$ on a log-log scale. If the coefficient of determination was $R^2 \leq 0.95$, we excluded the data points with the smallest L , then repeated the fitting with the remaining $n - i$ data points, where i is the number of excluded points. This was continued until $R^2 \geq 0.95$. Then L_c was defined as L where the relative difference between the fit and the data exceeded 10 %.

Figure 5b shows that the magnitude of $L_c \sim 10^2$ m throughout the winter season, with only minor variations observed between the three winter time periods of active deformation. Based on the deformation patterns of Figure 4, L_c in November appears to be related to multiple minor deformation features across the whole $10 \text{ km} \times 10 \text{ km}$ sea ice area, whereas later during the winter, it appears to be related to few major features. We also tested that this estimate is consistent for other deformation events occurring during the wintertime. However, we could not find L_c for July. This was expected due to the diffuse nature of summer time ice deformation, already seen in the analysis of deformation time series and deformation fields above (Figures 2 and 4).

Figure 6a show the behavior the system over temporal scales, T , and Figure 6b compares the derived values of α with the summer conditions. Values of α derived for our data, describing the intermittency of the deformation events, also showed a spatial-scale dependent change. For scales $L \leq 150$ m, α was about constant for all winter time cases as Figure 6b shows. This stabilization of α suggests that to maintain the typically observed scaling behavior, one must consider spatial scales significantly larger than 150 m. For observations with $L \geq 150$ m, on the other hand, α decreased with increasing L . July data does not show this feature. Overall, the alpha α values were about 30% lower than those reported by Oikkonen et al. (2017), despite our time series exhibiting stronger intermittency. This is likely because of our analysis considering only active deformation periods, where the average deformation rates even over longer temporal scales stay consistently larger.

4 Discussion

Our study presents new insights into sea ice dynamics by extending the scaling analysis towards smaller scales. We find that the limit for scale invariance is approximately $L_c \sim 10^2$ m for winter pack ice. For $L_c > 10^2$ m, our results align with previous observations of fractal properties at similar scales (Oikkonen et al., 2016, 2017). Scale invariance at intermediate scales has



205 been thoroughly investigated only by Oikkonen et al. (2017), who analyzed a $15 \text{ km} \times 15 \text{ km}$ with $L \geq 250 \text{ m}$ and $T \geq 10 \text{ min}$,
 with the results considered evidence for scale invariance extending at least down to $L = 250 \text{ m}$ (Weiss, 2017; Rampal et al.,
 2019). Our findings now provide observations on the lower limit for the scale invariance. Since our analysis was conducted
 on the same $10 \text{ km} \times 10 \text{ km}$ area over the winter season, it can be argued a different region might have produced a different
 magnitude for L_c . With this in mind, it is intriguing that $L_c \sim 10^2 \text{ m}$ remained constant, even if the ice cover properties went
 210 through changes over the winter (Itkin et al., 2023).

It is well-established that large-scale sea ice deformation fields are multifractal in both spatial and temporal dimensions
 (Marsan et al., 2004; Weiss, 2013), indicating that different statistical moments could reveal further complexities. A multifractal
 analysis could further describe the limits of the fractal properties in the small domains and at high resolutions. The early
 winter deformation fields, characterized by more continuous spatial and temporal deformation, is assumed to exhibit different
 215 multifractal properties compared to late-winter fields dominated by only a few large deformation features. The objective of
 this paper, however, has been to perform an analysis of the high-resolution MOSAiC data using methodologies akin to those
 employed by other works with similar datasets.

The variation of the spatial scaling exponent β with increasing spatial scale was recently observed by Hutchings et al.
 (2024), suggesting that the sea ice deformation rate does not actually exhibit scale invariance. Their analysis considered buoy
 220 data down to spatial scales of 1 km , below which they did not have sufficient data points to support claims about smaller scales.
 By extending these observations to spatial scales smaller than 1 km , we similarly observe a distinct change in the behavior of
 β at approximately $L \sim 10^2 \text{ m}$, thereby complementing the observations of Hutchings et al. (2024). However, in contrast to
 the buoy data, we observe a clear stagnation in the values of β for $L \leq 10^2 \text{ m}$. This is likely due to the dominance of larger,
 distinct deformation features clearly observable with coarser spatial resolutions—mainly leads and shear zones—over smaller
 225 deformation features. We also observe that this relationship depends on the temporal scale of the observations. Unlike the buoy
 data, we have approximately 4×10^8 data points over a two-week period at spatial scales of $L = 20 \text{ m}$ and temporal scales of
 $T = 10 \text{ min}$. Our results complement the analysis of the buoy data by revealing system regime changes with varying spatial
 and temporal scales.

Our results have several implications, particularly for modeling sea ice using the continuum approach. Rheological models
 230 applied on spatial resolutions larger than L_c may not capture ice behavior below L_c , where the continuum assumptions may
 break down altogether for the ice pack during winter. We see that at the scales studied here, the largest discontinuities dom-
 inate the overall deformation field (Figure 4). The size of these features is approximately L_c (Figure 1). This is particularly
 pronounced during mid- and late winter, where the ice field akin to rigid bodies separated by large deformation features. Early
 winter conditions exhibit simultaneous deformation at multiple locations, resulting in less distinct deformation features. In con-
 235 trast, the summer ice pack behaved drastically different with no L_c found, highlighting the need for general rheological models
 to account for seasonal variation. Moreover, our observations suggest that ice deformation in winter is driven by short-lived
 (hour-scale) and spatially localized events that generate features on the order of 100 m .

How can we model sea ice cover if there is no continuum description at small scales? The discrete element method (DEM)
 has been used in ice mechanics and dynamics studies on various scales for decades (Hopkins et al., 1991; Hopkins and Hibler,



1991; Hopkins, 1994, 1996, 2004; Hopkins and Thorndike, 2006) and seen as a tool for future studies as well (Blockley, 2020; Hunke et al., 2020). Current DEM developments enable modeling multifracture of sea ice, three-dimensional deformation processes, and interactions of ice features over large sea ice domains, without requiring a continuum model (Manucharyan and Montemuro, 2022; Åström et al., 2024; Åström and Polojärvi, 2024; Muchow and Polojärvi, 2024; Tsarau et al., 2024); apart from that for intact level ice, high resolution DEM tools do not need a rheological model.

5 Conclusions

This paper reveals a 10^2 -m lower bound to the scale-invariance of sea ice deformation. The results challenge conventional assumptions that scale invariance might reach even down to scales comparable to ice thickness and underscores the need for further consideration on the applicable models at smaller scales. We show that sea ice deformation below the 10^2 -m scale is dominated by large features such as leads and fracture zones. These features disrupt the traditional continuum assumption that underlies most high-resolution sea ice models. Seasonal differences further reinforce the complexity of these mechanics: what holds true for winter pack ice breaks down as the ice transitions towards the marginal ice zone in summer. During winter, intermittency and spatial localization in deformation persist, while in summer deformation becomes more granular, defying the applicability of the fractal analysis. The seasonal transition highlights the influence of ice properties, thickness, and external forcing on deformation patterns. Our study was based on unique field observations from ship radar data gathered during MOSAiC expedition and on their analysis by using novel deep learning-based tools; our results demonstrate that such high-resolution empirical data can guide the development of hybrid modeling frameworks that combine continuum modeling with models explicitly accounting for the discontinuities in ice. These approaches will help in understanding of scale dependency in sea ice deformation and in improving predictions of ice-covered seas in future and estimates related to climate warming.



Appendix A: Strain measures

Any strain tensor can be described in terms of deformation gradient $\mathbf{F} = \mathbf{I} + \partial \mathbf{u} / \partial \mathbf{X}$, where $\mathbf{u} = \mathbf{u}(\mathbf{X}, t)$ is a time-dependent displacement vector of a material point and the partial derivative taken with respect to its position vector $\mathbf{X}$ in a reference configuration. By polar decomposition, $\mathbf{F} = \mathbf{R}\mathbf{U}$, where $\mathbf{R}$ and $\mathbf{U}$ are the rotation matrix and stretch tensor, respectively, describing the rigid body rotation and the actual deformation. Green-Lagrange strain tensor is then defined as (Gurtin, 1982)

$$\mathbf{E} = \frac{1}{2} (\mathbf{F}^T \cdot \mathbf{F} - \mathbf{I}) = \frac{1}{2} (\mathbf{U}^T \cdot \mathbf{R}^T \cdot \mathbf{R} \cdot \mathbf{U} - \mathbf{I}) = \frac{1}{2} (\mathbf{U}^T \cdot \mathbf{U} - \mathbf{I}), \quad (\text{A1})$$

since $\mathbf{R}^T \cdot \mathbf{R} = \mathbf{I}$ for rotations. The last form of the this definition shows that $\mathbf{E}$ is independent of rotations. The components of $\mathbf{E}$ are given by

$$E_{ij} = \frac{1}{2} \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} + \frac{\partial u_k}{\partial x_i} \frac{\partial u_k}{\partial x_j} \right) \quad (\text{A2})$$

and differ from the components of the infinitesimal strain tensor, ϵ , by the inclusion of quadratic terms, negligible for small deformations. Above we compare our results to previous large-scale estimates, that have typically used ϵ . For this it is practical

to introduce unit extensions

$$l_{ii} = \sqrt{1 + 2E_{ii}} - 1 \quad \text{and} \quad l_{xy} = \sin^{-1} \left(\frac{2E_{xy}}{\sqrt{(1 + 2E_{xx})(1 + 2E_{yy})}} \right) \quad (\text{A3})$$

which in the case of small deformations or deformations leading to normal strains only reduce to components ϵ_{ii} of ϵ . The principal values of $\mathbf{l}$ are given by

$$E_{1,2} = \frac{l_{xx} + l_{yy}}{2} \pm \sqrt{\left(\frac{l_{xx} - l_{yy}}{2} \right)^2 + l_{xy}^2} \quad (\text{A4})$$

and can be used to express the divergence, maximum shear, and the total deformation as

$$E_d = E_1 + E_2, \quad E_s = \frac{1}{2}(E_1 - E_2) \quad \text{and} \quad E_t = \sqrt{E_d^2 + E_s^2}. \quad (\text{A5})$$

Similar operations can be applied in strain-rate analysis to derive $\dot{E}_t$, since the strain-rate tensor is derivable from the two strain tensors defined for two consecutive time steps within analyzed data.



Code and data availability. The source code of the deformation detection method is archived in Zenodo (Uusinoka, 2024). The ship radar
280 raw data is available in PANGAEA (Krumpen et al., 2021). The time series and mean-based scaling data can be requested from Matias
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