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Localised thickening and grounding of an Antarctic ice shelf from tidal triggering and sizing of cryoseismicity

Myrto Pirli\textsuperscript{a}, Sebastian Hainzl\textsuperscript{b}, Johannes Schweitzer\textsuperscript{c, d}, Andreas Köhler\textsuperscript{e} and Torsten Dahm\textsuperscript{b, f}

\textsuperscript{a} Tårnbyveien 370, 2013 Skjetten, Norway, e-mail: myrto.pirli@gmail.com
\textsuperscript{b} GFZ German Research Centre for Geosciences, Telegrafenberg, 14473 Potsdam, Germany, e-mail: hainzl@gfz-potsdam.de, torsten.dahm@gfz-potsdam.de
\textsuperscript{c} NORSAR, Gunnar Randers vei 15, 2007 Kjeller, Norway, e-mail: johannes.schweitzer@norsar.no
\textsuperscript{d} Centre for Earth Evolution and Dynamics (CEED), University of Oslo, P.O. Box 1028 Blindern, 0315 Oslo, Norway
\textsuperscript{e} Department of Geosciences, University of Oslo, P.O. Box 1047 Blindern, 0316 Oslo, Norway, e-mail: andreas.kohler@geo.uio.no
\textsuperscript{f} Institute of Earth and Environmental Sciences, University of Potsdam, 14476 Potsdam, Germany.

Corresponding author: Myrto Pirli, e-mail: myrto.pirli@gmail.com

Abstract

We observe remarkably periodic patterns of seismicity rates and magnitudes at the Fimbul Ice Shelf, East Antarctica, correlating with the cycles of the ocean tide. Our analysis covers 19 years of continuous seismic recordings from Antarctic broadband stations. Seismicity commences abruptly during austral summer 2011 at a location near the ocean front in a shallow water region. Dozens of highly repetitive events occur in semi-diurnal cycles, with magnitudes and rates fluctuating steadily with the tide. In contrast to the common
unpredictability of earthquake magnitudes, the event magnitudes show deterministic trends within single cycles and strong correlations with spring tides and tide height. The events occur quasi-periodically and the highly constrained event sources migrate landwards during rising tide. We show that a simple, mechanical model can explain most of the observations. Our model assumes stick-slip motion on a patch of grounded ice shelf, which is forced by the variations of the ocean-tide height and ice flow. The well fitted observations give new insights into the general process of frictional triggering of earthquakes, while providing independent evidence of variations in ice shelf thickness and grounding.

Keywords

tidally modulated cryogenic seismicity; stick-slip motion; event recurrence predictability; ice-shelf thickness; ice-shelf grounding; East Antarctica

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1. Introduction

In recent years, repetitive tide-modulated seismicity has been discovered at glaciers and ice shelves in Antarctica (e.g., Barruol et al., 2013; Hammer et al., 2015; Lombardi et al., 2016; Winberry et al., 2014; Zoet et al., 2012) and Greenland (Podolskiy et al., 2016). Typically observed near the grounding zone, it shows a wide variety of temporal patterns and correlations with the components of the ocean tide, which can be used as diagnostic tools to assess the driving mechanisms of this type of cryogenic seismicity and potential links to glacial dynamics. Main interpretations involve stick-slip motion at the ice/bedrock interface.

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1 Abbreviations
CC Cross-correlation
CFS Coulomb Failure Stress
DML Dronning Maud Land
FIS Fimbul Ice Shelf
HMM Hidden Markov Model
SNR Signal-to-noise ratio
STA Short-term average
(e.g., Barruol et al., 2013; Lombardi et al., 2016; Winberry et al., 2014; Zoet et al., 2012) and brittle deformation of the ice shelf due to tidal flexure (Barruol et al., 2013; Hammer et al., 2015; Lombardi et al., 2016), all recognizing the importance of stress and strain rate variations to the triggering of cryoseismic activity (e.g., Bindschadler et al., 2003; Hammer et al., 2015; Podolskiy et al., 2016; Winberry et al., 2014). However, relevant studies often have short observation intervals, due to the need for dedicated field deployments (e.g., Lombardi et al., 2016; Podolskiy et al., 2016), the longest ranging ones reporting one or more non-consecutive intervals of about a year (Barruol et al., 2013; Hammer et al., 2015; Winberry et al., 2014).

The study focuses on the Fimbul Ice Shelf (FIS), in Dronning Maud Land (DML), East Antarctica. The main contributor to the ice shelf is the outlet of the Jutulstraumen glacier, whose ice tongue, the Trolltunga, partly extends past the continental shelf break into the Weddell Sea (Fig. 1a). The differential flow between the fast-flowing central part of the outlet glacier and the much slower lateral parts of the shelf (e.g., Rignot et al., 2011) creates zones of shear deformation, characterised by abundant crevasses and rifts (Humbert and Steinhage, 2011). Although the central basin of the FIS cavity has a deep seabed (e.g., Nøst, 2004), near the calving front, the ice shelf becomes locally grounded on shallow bathymetric features that either divert ice flow (ice rises) or allow it to continue over them (ice rumples) (Matsuoka et al., 2015; van Oostveen et al., 2017).

We present seismic records of an almost two decade long continuous monitoring of a specific source region at the FIS, showing emergent activity and trends over several years with strikingly similar and regular seismic events near the ocean front, in an area of outcrops of bedrock. A key question is whether the distinct seismicity pattern can be explained with established, physics-based models of earthquake triggering, and whether environmental drivers can be identified and quantified. We employ a simple mechanical model based on
tidally modulated shear and normal stress and demonstrate that this activity is likely related
to stick-slip motion of the thickened ice shelf, which is newly grounded on a shallow
bathymetric feature.

2. Data and methods

2.1 Data

We use waveform data from the permanent, international seismic network in DML (Fig. 1a).
Continuous data are used from the broadband, three-component stations SNAA and
TROLL, situated at the South African research station SANAE IV and the Norwegian station
Troll, respectively. Employed SNAA records span the time-period March 1997 to end 2015,
and TROLL data February 2012 to end 2015. In addition, we use selected records from the
Watzmann seismic array, deployed around station VNA2, both belonging to the seismic
network of the German research station Neumayer III.

Ocean tide heights for the study region, sampled at each full minute, are estimated from the
high-resolution regional inverse tide model CATS2008a_opt (Padman et al., 2002, 2008) for
the ocean around Antarctica, including the areas under the floating ice shelves. GPS
measurements (Kohler and Langley, 2016) obtained in November 2010 at a location near
the study region (~ 30 km to the SSE) are used to assess the accuracy of the predicted tidal
phase.

2.2 Compilation of cryoseismic event catalogue

2.2.1 Detection by waveform cross-correlation

The observed cryoseismic events have so similar waveforms that they can be identified by
visual inspection (Fig. 2). Therefore, we compile our event catalogue using the array-based
waveform cross-correlation (CC) detector of Gibbons and Ringdal (2006). Small magnitude
events are detected through the enhancement of signal-to-noise ratio (SNR), by stacking the
correlation traces for all individual channels of a seismic array or a single, three-component station. Master template events (Table S1 in Supplementary Material), selected to have sufficiently high SNR and to sample adequately intervals of high activity, as well as to cover the entire duration of the study, run through the continuous waveform records of TROLL and SNA in search of similar events, based on a user defined SNR threshold. The master templates employ the entire body-wave wavetrain, with a length of 50 s for TROLL and 45 s for SNA, bandpass filtered in the frequency range of maximum SNR (2 – 6 Hz). Detection thresholds (SNR, CC-coefficient) are defined through visual inspection of a large sample of results, accepting detections for SNR larger than 10, and only considering CC-coefficient values higher than 0.40.

2.2.2 Detection with a Hidden Markov Model classifier
CC-detector results are compared to those of a Hidden Markov Model (HMM) classifier (Hammer et al., 2013; references therein) (Fig. S1 in Supplementary Material), a probabilistic approach able to statistically capture the intrinsic variability of the characteristics of a class of seismic signals. The observed data are modelled by a sequence of hidden, unobserved states, each assumed to emit observations according to a particular emission probability (see Supplementary Material for details). These observations are obtained by parametrizing the seismic signal by a time series of feature vectors (here, spectral power in different frequency bands). An HMM is then trained by estimating the probability distribution of the features at each state and the transition probabilities (see Supplementary Material) between the states using a pre-selected training data set with known class labels (i.e., event type). The classifier must also include a class for the seismic background noise. In the current implementation, event states and their probabilities are estimated from seismic background noise data modelled by Gaussian mixtures, and the event HMM is initially trained using a single seismic signal (Hammer et al., 2013; references therein).
The trained classifier is applied on all available continuous TROLL data in 2013. An event is reported if the probability of the event HMM is larger than that of the noise HMM in a given data window. Post-processing includes merging detections belonging to the same event in consecutive classification windows and removing all events with too low absolute probabilities and too small probability differences compared to the noise HMM. In addition, the SNR is used to reject false positives. An extended description of the method and specifics on its implementation herein can be found in the Supplementary Material.

2.3 Cryoseismic event location

The source of the cryoseismic events is located using the seismic event location routine HYPOSAT (Schweitzer, 2001, 2018). To obtain an estimate as accurate as possible and considering the small magnitude of the events, we read seismic phase onsets on the summation trace (beam) for the Watzmann array (e.g., Schweitzer et al., 2012) and the summed traces of events with CC-coefficients larger than 0.90 (multiplets) for stations TROLL and SNAA. In addition to absolute onset times, travel-time differences between the employed seismic phase onsets (Sn - Pn) are used. The apparent velocity and backazimuth of the seismic phases are determined through array-data processing (e.g., Schweitzer et al., 2012) for the array and three-component analysis (Schweitzer, 2013) for the single stations. The seismic velocity model (Fig. S2 in Supplementary Material) that provides the best fit to the observations describes the crust and upper mantle in the region based on information from local (Nøst, 2004) and regional studies (Bayer et al., 2009; Torsvik, 2015). Focal depth is kept fixed at the surface (ice – bedrock interface), since the recording network does not provide adequate resolution to invert for it (e.g., Havskov et al., 2012).

The small magnitude of the events (largest ~ 1.2, see section 2.4) and the distance to the seismic stations do not facilitate the location of the volume of detected events. Instead, we determine the size of the activated area based on fluctuations in the time difference of
seismic onsets at TROLL and SNAA that are translatable into corresponding differences in source location. We estimate the corresponding difference in travel-time, add this change (0.060 s for TROLL and 0.099 s for SNAA) to the S-P difference of our absolute location estimate, and obtain a new solution that expresses the spatial extent of the activated area. Larger time differences between TROLL and SNAA than those of the event used for the absolute location imply that the corresponding sources are farther away from the stations than the reference location.

**2.4 Event magnitude estimates**

For our dataset, magnitude is defined as the logarithmic maximum amplitude, $A_{\text{max}}$, between the vertical, radial and transverse component for each of the stations TROLL and SNAA:

$$M = \log_{10} A_{\text{max}}$$

(1)

For this purpose, we construct short-term averaged (STA) traces of waveforms filtered between 2 and 4 Hz, using a window length of 1.4 s, and measure the maximum amplitude within a 20 s long window that contains the S-wave. We correct for the instrument response of each station, to obtain amplitudes in nm/s. The result is a self-consistent magnitude scale for this specific population of seismic events, where it is size relations between events that are important and not absolute values.

**3. Observations**

**3.1 Spatiotemporal characteristics of cryoseismicity**

The recorded activity is manifested as highly similar, small magnitude events that exhibit very alike temporal patterns at stations TROLL and SNAA. One of the best recorded cryoseismic events occurred on 9 September 2013 04:40:20.3 UTC and had a station magnitude at TROLL of $M = 1.16$. It is located at 70.031°S ± 0.025°, 0.452°E ± 0.035° on the outlet of the Jutulstraumen glacier, near the front of the FIS to the ocean, at distances of 220 – 230 km from the nearest seismic stations SNAA and TROLL, and 310 km from VNA2 (Fig.
Our relative location scheme (section 2.3) shows that the events are taking place at an area with a largest length of about 470 m, trending almost N-S, with the absolute location estimate near its southern border (inlet of Fig. 1b).

The complete catalogue of events obtained using the CC-detector (section 2.2) contains altogether about 10000 cryoseismic events between March 1997 and end 2015, covering almost two decades (Fig. 1c and Fig. S1 in Supplementary Material). The earliest findings are low-level activity in 2003 and a few isolated events between 2005 and 2009. From late 2011 on, the observed level of cryoseismicity increases, varying between intervals without any events to high intensity phases with tens of events per day. In general, activity intensifies during the austral winter-spring months and decreases or pauses from summer through autumn, with the maximum levels fluctuating from year to year. This seasonal pattern has similarities to that observed along the DML coast West of Trolltunga, near the Neumayer III station in 2004 (Hammer et al., 2015), although a systematic comparison is inhibited by a lack of data during the entire austral spring season in that study.

We focus on the most intense seismic activity between August and October 2013 that is characterised by the highest waveform similarity levels (CC-coefficients in Fig. S1 in Supplementary Material). We use TROLL data to take advantage of the higher sampling rate and the better SNR conditions compared to SNAA (Schweitzer et al., 2014). Confidence in the CC-detector results is strengthened by the very similar patterns retrieved from SNAA and TROLL (Fig. S1 in Supplementary Material). The distributions at the two stations differ slightly in the number of detected events, mainly due to diverse noise conditions that affect the performance of the master event templates. However, the general activity patterns are the same at the two stations. The robustness of the results is evaluated by comparing the CC-detector results against the findings of the HMM classifier (section 2.2), which corroborates the observed temporal pattern (Fig. S1 in Supplementary Material). Although
the HMM method provides a slightly less complete catalogue for 2013 than the CC-detector, it results in a similar distribution. The HMM classifier is trained with spectral features that should allow for some degree of waveform variability within the event class. The similarity in results therefore suggests that no significant number of seismic signals with slightly different waveforms compared to the master events is missed by the CC-detector.

3.2 Tidal modulation of cryoseismicity

We observe strong correlations between the timing and magnitude of the events (blue symbols in Figs. 3 and 4) with the amplitude of the ocean tide. Within the 14-day tidal cycle, a clear correlation is found between event rates and tidal amplitude, with almost double as many events around spring tides than around neap tides (Fig. 3a,b). The tides also modulate the event magnitude distribution, with highly deterministic patterns being observed within each semi-diurnal tidal cycle. A clear magnitude trend occurs that differs depending on the proximity of the cycle to the neap or the spring tide (Fig. 3c-e). For quantification, we calculate for each cluster (defined by sequences of a minimum of 5 events with inter-event times less than 6 hours) the average magnitude difference, $\langle \Delta M \rangle$ (blue points in Fig. 3c). The result indicates negative trends ($\langle \Delta M \rangle < 0$) during spring tides and increasing magnitudes during neap tides ($\langle \Delta M \rangle > 0$). Exemplary clusters are shown in Fig. 3d,e, illustrating the increasing magnitudes during neap tides (Fig. 3d) and an almost linear decrease during spring tides (Fig. 3e). Similar trends can be also seen for the other clusters in the analysed time period (see time series in Fig. S3, Supplementary Material).

In all cases, the events occur quasi-periodically indicating that each event leads to a partial unloading of the system. All events occur with a waiting time of at least 3 min after the preceding event and most of them with an inter-event time of 5-10 min (Fig. 4a), a feature that is present in the results of both detection methods and thereby not an artifact. The observed peaked inter-event time distribution (histogram in Fig. 4a) is in contrast to random
occurrences, which would be expressed by an exponential interevent-time distribution. By
assigning a 360° phase cycle to two successive tidal maxima, we find that icequakes are
predominantly triggered during rising tides and stop after the maximum height of the tide has
been reached within each 12-hour cycle (histogram in Fig. 4b). The maximum frequency is
observed just before the tide reaches its peak, followed by an almost complete lack of
occurrences at falling tides, although we cannot exclude the occurrence of events at low
tides with magnitudes too small to record at the distance of observation (> 200 km). This
temporal pattern is quite similar to that of tidally modulated seismicity at other places in DML
(Hammer et al., 2015; Lombardi et al., 2016), although not all individual features are
common; however, Hammer et al. (2015) do not separate between different source regions,
while the results of Lombardi et al. (2016) are based only on one month of data.

Besides the deterministic magnitude and timing patterns, we observe systematic changes in
the difference between CC-detection times at TROLL and SNAA (ΔCC), which imply a shift
in source locations (see ΔCC-time series in Fig. S4, Supplementary Material). The ΔCC-
value is found to decrease within clusters at the end of the 12-hour cycles; in particular, the
difference of the ΔCC value between event pairs is shown to be linearly correlated to the
occurrence time difference of the event pair (Fig. 4c). This result indicates a systematic
migration of the activity within the 12-hour cycles, in a landward direction opposite of the ice
flow (Fig. 1b and inlet). This is consistent with the observations of Brunt et al. (2011) and
models of tidal migration of ice-shelf grounding lines (e.g., Sayag and Worster, 2013; Tsai
and Gudmundsson, 2015) that attribute slip during high tides to increased lubrication in the
system. Finally, the event magnitudes are found to correlate with the tidal amplitudes.

Although a broad range of event magnitude values are observed for different tide heights,
Fig. 4d shows a positive trend for small tidal levels, while a reversed trend is found for larger
levels, as already seen in Fig. 3d,e.
4. Modelling

4.1 Model setup

The observed suite of temporal, magnitude and migration features provides strong constraints for modelling. Our model is based on an assumption that the observed cryogenic events result from basal stick-slip motion of the ice shelf on a localized grounding feature (ice shelf pinning point in e.g., Robel et al., 2017), as shown in the schematic plot of Fig. 5a. Loading occurs due to an increase in shear stress related to ice flow, as well as changes of the normal stress, both modulated by the tidal amplitude. Slip is initiated when the Coulomb Failure Stress \((CFS)\) exceeds a critical level in the contact area between the base of the ice shelf and the bedrock protrusion.

The main model variable effecting the stress state, the hypocentre location, and the event magnitude, is the difference \(\Delta\) between the elevation \(z_2\) of the peak of the local grounding region (bedrock protrusion) and the elevation \(z_1\) of the bottom of the ice shelf, when the latter is considered to be freely floating. This referencing is a valid approximation if the ice shelf is only slightly elevated at the bedrock protrusion (see Fig. 5a and Table 1 for a description of the model parameters). The \(\Delta\)-value varies on short-time scale due to tidal variations \(h(t)\) and on long-time scales due to ice shelf thinning or thickening. Stick-slip motion can only occur for \(\Delta > 0\) and the contact area between the grounding point and the bottom of the ice shelf increases for increasing \(\Delta\)-values. Considering that the density of glacial ice is approximately 90% of the water density and the ice shelf has a thickness \(W\), the bottom of the ice shelf is located at \(h(t) - 0.9W\) relative to the average sea-level. With parameter \(H\) defined as the average \(\Delta\)-value in our analysed time period in 2013, the temporal variation of \(\Delta(t)\) is given by \(\Delta(t) = H - h(t) + 0.9 (W - W_{2013})\), where \(W_{2013}\) is the average thickness of the ice shelf during this 80-day time interval.
In our simplified kinematic modelling, the $\Delta$-value directly determines the contact area and the stress state at any time, as well as the magnitude and the hypocentre location at times when failure occurs:

1. Contact area: The contact area increases for increasing $\Delta$-values. Although the actual area where stick-slip occurs can be smaller, for simplicity we assume that it coincides with the geometrical contact area. While actual conditions are more complex, e.g., our geometry does not account for the hinge zone and bending/flexuring of the ice (e.g., Vaughan, 1995; Hulbe et al., 2016), we simply approximate the contact area as the cut surface of the ice shelf with an elliptical topography with circular base area of maximum radius $R_{\text{max}}$, which leads to:

$$A(\Delta) = \begin{cases} 
0 & \text{for } \Delta \leq 0 \\
\pi R_{\text{max}}^2 \left[1 - \left(1 - \frac{\Delta}{H}\right)^2\right] & \text{for } 0 < \Delta < H \\
\pi R_{\text{max}}^2 & \text{for } \Delta \geq H
\end{cases}$$  

(2)

Our assumption that $A$ is fixed for $\Delta > H$ is not crucial because almost no events occur at those times. For $0 < \Delta < H$, $\Delta$ decreases during increasing tides, leading to a migration of the grounding line on the bedrock protrusion within the tidal cycles, which has been observed for ice shelves before (Brunt et al. 2011). For $\Delta < 0$, the bottom of the freely floating ice shelf is above the peak of the bedrock protrusion, and the contact is completely lost.

2. Stress: The Coulomb-Failure stress is defined as $CFS = \tau - f\sigma_n$ with the friction coefficient $f$, the shear stress $\tau$, and normal stress $\sigma_n$. The shear stressing rate is assumed to be proportional to the ice-flow velocity, which has been previously observed to be positively correlated predominantly with the 14-day, but also the semi-diurnal tidal cycles (Murray et al., 2007), a fact that is also suggested by the observed 14-day rate-cycles of the analysed activity (Fig. 3a). Although reality is more complex, we thus simply assume that the shear stressing rate is proportional to
the tidal amplitude, \( \dot{\tau}(t) = \tau_0(t)E_h(t)/\langle E_h \rangle \), where \( \tau_0 \) is the average shear stressing rate and \( E_h \) (shown in Fig. 3a as grey line) is the envelope of the tidal variations with mean \( \langle E_h \rangle \). Note that the envelope only leads to fortnightly but no diurnal variations of the shear stressing rate. For an evaluation of the impact of this assumption, the results for a constant shear stressing rate are shown in Fig. S5 in Supplementary Material. They indicate that all features besides the 14-day rate-cycle are similarly well reproduced and thus not crucially dependent on shear stressing variations.

The second component of the CFS-value, the normal stress \( \sigma_n \), depends on \( \Delta \) and thus is also time dependent. For a first-order approximation, we use the analytic point force solution for a deflection of \( \Delta \) at the centre of an infinite elastic plate overlying an inviscid fluid, which yields a linear relation between the point force \( F \) and \( \Delta \), \( F = c_f \Delta \), where the proportionality factor \( c_f \) depends on the elastic parameters, the thickness \( W \) of the elastic plate, and the density contrast (Brotchie and Silvester, 1969; Jha et al., 2017). Thus, the normal stress acting within the contact area \( A(t) \) can be approximated by \( \sigma_n(t) = c_f \Delta(t)/A(t) \), and the CFS becomes:

\[
\text{CFS}(t) = \text{CFS}(t_0) + \dot{\tau}_0 \int_{t_0}^{t} E_h(\tilde{t})/\langle E_h \rangle d\tilde{t} - f c_f \Delta(t)/A(\Delta(t)) 
\]

(3)

To get rid of undetermined parameters, we considered the normalized stress \( S = \text{CFS}/\overline{\text{CFS}} \), which can be expressed by substitution using Eqs. (2) and (3) as:

\[
S(t) = \begin{cases} 
0 & \text{for } \Delta(t) \leq 0 \\
S(t_0) + C \int_{t_0}^{t} E_h(\tilde{t})/\langle E_h \rangle d\tilde{t} - \frac{\Delta(t)}{H} \left[ 1 - \frac{\Delta(t)}{H} \right]^{2} & \text{for } 0 < \Delta(t) < H \\
S(t_0) + C \int_{t_0}^{t} E_h(\tilde{t})/\langle E_h \rangle d\tilde{t} - \frac{\Delta(t)}{H} & \text{for } \Delta(t) \geq H 
\end{cases}
\]

(4)

with \( \overline{\text{CFS}} = f c_f H/(\pi R^2_{\text{max}}) \) and \( C = \dot{\tau}_0/\overline{\text{CFS}} \). Each icequake is assumed to lead to a constant shear stress drop \( S_\Delta \) at origin time \( t_0 \). For simplicity we assume instantaneous healing, and the next event is triggered when the stress is reloaded, i.e., if \( S(t) - S(t_0) \geq S_\Delta \). An example of the temporal evolution of the stress
components during a 12-hour tidal cycle is shown in Fig. S6 in Supplementary Material.

3. Magnitude: We define the magnitude $M$ as the logarithm of the maximum amplitude of the seismic waves (as done for the observations) and assume that this amplitude is proportional to the seismic moment $M_0 = \mu A d$ of the event, where $\mu$, $A$, $d$ is the shear modulus, the slip area and the average slip value, respectively. Here, we assume that the whole contact area slips during an event and that $A$ is thus given by Eq. (2). The frictional properties of glacial ice are known to be strongly dependent on temperature, with a transition from slip-weakening (brittle) to slip-strengthening (creep) characteristics for increasing temperatures (McCarthy et al., 2017). Thus, we assume that seismic slip is limited by a maximum value $E_d$ of the energy released per unit area, where $E_d$ is assumed to shift the temperature from the brittle into the creep regime. If this constant is exceeded, ongoing slip is assumed to be aseismic, as a result of reduced slip velocities due to velocity-strengthening. Because the released energy is equal to force times slip, the seismic slip is thus assumed to be inversely related to the normal stress, i.e., $E_d = \sigma_n d$. Consequently, the magnitude depends on the $\Delta$-value at the time of the event and is equal to:

$$M = \log_{10} \frac{\mu A E_d}{\sigma_n} = M_1 + \log_{10} \frac{A^2}{\Delta}$$

(5)

where $M_1$ is a constant and $A$ depends on $\Delta$ according to Eq. (2).

4. Hypocentre: The boundary conditions of the ice shelf relative to the source region are asymmetric with a freely floating ice shelf on the ocean side, while the ice is fixed on land. During high tides, the normal stress will be therefore smaller on the ocean side of the local grounding feature and minimum at the edge of the contact area. According to the Coulomb-Failure criterion (King et al., 1994), we assume that the event will always nucleate at the ocean-side edge of the contact area, where CFS is
largest, which leads to the hypocentre location \( x = \sqrt{A/\pi} \) measured relative to the peak location (with \( x \)-axis directed towards the ocean and \( A \) given from Eq. (2)).

Model simulations are driven by the ocean tides, calculated for the given location by the CATS2008a_opt model (Padman et al., 2002, 2008) during the analysed 80 days between August and October 2013. The simulations depend only on three model parameters that change the occurrence and magnitude patterns, namely (i) \( C \), which determines the relative strength of shear stress changes versus normal stress changes during tidal cycles; (ii) \( H \), which is the average height of the topographic peak of the shallow bedrock feature relative to the bottom of the ice shelf; and (iii) the stress drop, \( S_\Delta \) (see Table 1). The latter is fixed by the condition that the total number of simulated events equals the observed ones, leading to \( S_\Delta = 0.03 \), and \( H \) is set to the maximum tidal amplitude \( h \) which triggered an event in this time period, namely \( H = 0.8 \) m. Thus, the only remaining free parameter is \( C \), which effects the phase distribution of the triggered events. The best model fit to the observed phase distribution (see Fig. 4d) yields a value of \( C = 0.7 \). For smaller (larger) \( C \)-values, the model predicts a later (earlier) onset of activity during the tidal cycles. All other patterns of the simulated activity (Figs. 3a,c and 4a,c,d) are a direct outcome, without additional parameter adjustments.

**4.2 Modelling results**

Most of the characteristics of the observed cryoseismicity can be well explained by the proposed model (results shown by red symbols in Figs. 3 and 4). The model replicates well the 14-day periodicity of the event rates (Fig. 3a). It also reproduces the general magnitude trends, although it under-predicts the strength of the magnitude increase during cycles at neap tides (Fig. 3b). The assumed unloading-loading mechanism also explains the quasi-periodic recurrences (Fig. 4a), as well as the correlation of event frequency with the semi-diurnal tide (Fig. 4b). In the latter case, however, the simulations show a narrower
distribution with an earlier maximum compared to observations. A possible 30 min delay of
the real tides relative to the calculated ones would better fit to our model results (blue line in
Fig. 4b). In the model, epicentres are assumed to occur at the ocean-side edge because of
smallest normal stress. As a result, epicentre locations are found to migrate landwards
during cycles with a constant speed (Fig. 4c). This is in agreement with the observed linear
trend of the change of the CC-time differences. Finally, the observed positive and negative
correlations between tidal heights and event magnitudes are reproduced (Fig. 4d), although
the model fails to reproduce the observed variability.

Based on the successful model fit of the activity between August and October 2013, we use
the model to understand the long-term trends of the observed seismicity. In the period
between 2012 and end 2015, the timing of the events within tidal cycles shows a systematic
trend that is strongly correlated to the trend of event magnitudes. During periods with larger
magnitudes, the events are also triggered at peak levels of the tide, whereas no events are
triggered at high tidal levels during phases with lower magnitudes. This is shown in the left
plot of Fig. 5b, where the maximum tidal value that triggered an event is compared to the
maximum tidal amplitude at this time and the event magnitudes. In terms of our model, this
correlation can be explained by a variation of the ice shelf thickness. The underlying physical
process is illustrated in Fig. 5c. A certain thickness of the ice shelf is necessary that it
grounds on an underlying bedrock shoal. When the shelf is thinner, no grounding and hence
no stick-slip seismic events occur. This might explain the absence of recorded activity before
2011 (Fig. 1c). The behaviour of a grounded shelf still depends on its thickness. A medium
thickness, much like the concept of ephemeral grounding (Schmeltz et al., 2001), will result
in seismic events at tidal amplitudes low enough that the coupling between ice and bedrock
is not lost, while the ice goes afloat, and no stick-slip events occur at high tidal levels. A
thicker shelf will have on average a larger contact area, which implies that slip along it will
produce events of larger magnitudes compared to a thinner shelf, while allowing for event
triggering to continue also for higher tidal amplitudes than previously. Since the installation of station TROLL in February 2012, the largest event magnitudes have been observed during our modelling interval in 2013, so we can assume that the ice shelf was thickest then. To test the influence of different ice thickness $W$ at other time periods, simulations were repeated for different $W$ values, keeping all other parameters constant. As a result, we find an almost linear trend between ice thinning and the maximum tidal amplitude that triggers events (right plot in Fig. 5b). This model result can be used for the calibration of the observed variations during 2012 and 2015, which suggests variations of the ice shelf thickness in the range of 0.5 m in this period.

5. Discussion and conclusions

A primary assumption when designing our model is that the ice shelf is locally grounded. In the source region, the ice shelf is assumed to be generally afloat in shallow water (< 200 m), but the available underwater topography is only based on a sparsely interpolated bathymetry because of the inaccessible nature of the specific site (Nøst, 2004). The proximity to the Kupol Moskovskij ice rise to the East and various ice rumples at the eastern side of Trolltunga (Van Oostveen et al., 2017) suggests it is likely that the ice shelf is locally resting on a shallow bathymetric feature. This is in agreement with the well-developed SH phases of the recorded seismic waveforms (Fig. 2) that can only be generated with mechanical coupling to the bedrock (e.g., Müller, 2007).

Although our mechanical model is based on simplifications of the complex nature of ice shelf dynamics, it successfully explains most of the various distinctive features observed for the recorded seismicity. These characteristics include strong determinism of the timing and sizing of events. This contrasts with earthquakes, where the magnitudes and occurrence times of subsequent earthquakes are usually uncorrelated and described by random distributions, namely the Gutenberg-Richter distribution (Gutenberg and Richter, 1944) and
the Poisson model (e.g., Turcotte, 1992), respectively. This randomness is likely related to earthquake interactions in complex crustal fault networks. The observed cryogenic seismicity might act as a prototype for an isolated fault, which leads to a much higher predictability of
the system.

The time, magnitude, and migration patterns are well reproduced with only two model parameters. The occurrence of stick-slip motion is predicted when Coulomb stress exceeds a critical threshold value in the grounding area of the ice shelf. The assumed proportionality of shear stressing rates to tidal amplitudes reproduces successfully the observed fortnightly event rates. The daily event timing results from the competing effects of stress drops related to the cryoseismic events and Coulomb stress loading by normal stress reduction during increasing tides and continuous shear stress loading. The latter is essentially of constant rate during the time scale of the triggered diurnal stick-slip sequences (Fig. S6 in Supplementary Material). Event magnitudes depend on the rupture area; while the tide is rising, the contact area gradually decreases, accounting for the decreasing magnitude trend observed during semi-diurnal spring-tide cycles.

The only deviations between the model and observations concern the 30 min (~ 18°) delay between the tide that would produce the best fit and the CATS2008a_opt tidal model, and the failure of the model to reproduce the observed magnitude variability. The latter is likely related to our simplified modelling of magnitudes that is not based on more appropriate dynamic rupture modelling, which is outside the scope of the present study. The former may too be related to neglected dynamic effects or healing properties, but might also be a possible effect of actual bathymetry deviations from the geometry reflected in the tidal model (e.g., Padman et al., 2018). Although GPS measurements at a distance of ~ 30 km (Kohler and Langley, 2016), closer to the Kupol Moskovskij ice rise, show a good fit to the tidal model (~ 10 – 15 min phase deviations), discrepancies from actual bathymetry might be
more pronounced at the source region. Robel et al. (2017) consider phase lags between measured horizontal ice-displacement variations and tidal heights as indicators of stress variations, in settings where ice shelf/sheet grounding occurs. Small lags suggest that strain and the modulation of ice flow are dominated by back-stress variations rather than hydrostatic stresses. In our source region, known and uncharted local grounding points (i.e., ice rises and rumples) introduce resistance to ice flow (e.g., Schmeltz et al., 2001; Favier et al., 2016). Our model is very similar in concept to that of Robel et al. (2017), exploring the same processes from a different scope.

Based on the interpretation above, we propose that the lack of similar cryoseismic activity in our study region prior to late 2011 is connected to a reduced thickness of the ice shelf that keeps it afloat. A gradual increase of about 0.5 m takes place from late 2011 through 2013, resulting in the intense seismicity levels between August and October 2013. This phase is followed by some minor thinning of about 0.3 m, to account for the smaller event magnitudes observed in 2015. Recent studies of Antarctic ice discharge (Gardner et al., 2018; Shen et al., 2018) find no significant ice thickness variations for DML over the past 7 years, while studies going back to 2003 find slight variations (both positive and negative) at various locations (Horwath et al., 2012). However, relevant uncertainties in those studies are far larger than the thickness variations discussed herein. Our observations concern a very local scale and results from dedicated studies are required for a proper comparison. The origins of the inferred variation in ice thickness (e.g., advection, changes in accumulation or melting) are outside the scope of this paper. It is noteworthy that the seasonal fluctuations (i.e., more seismic activity winter to spring, less summer to autumn) coincide fairly well with the seasonal melt rates beneath the FIS (Abrahamsen, 2012) that control ice shelf thickness, maxima near the ocean front occurring during the austral summer.
Our model explains both the short-term oscillatory seismicity patterns and the long-term trend of the observations, with the fluctuation of seismic activity levels, including the lack thereof. It thus provides independent evidence of localised variations in ice-shelf thickness and a means to identify ephemeral grounding points.

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resolution (450 m), digital mosaic of ice motion in Antarctica (Mouginot et al., 2012; Rignot et al., 2011, 2017), acquisition 1996-2016, downloaded from https://nsidc.org/data/nsidc-0484.

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**Author contributions**

M.P. compiled the dataset, carried out seismic waveform analysis and collected non-seismological data. M.P. and J.S. performed CC event detection and location. A.K. performed HMM event detection. S.H. and T.D. developed the model. S.H. implemented the model and ran the simulations. All authors contributed to the interpretation of the results and to writing the manuscript.

**References**


(dataset) Rignot, E., Mouginot, J. and Scheuchl, B. (2017). MEaSUREs InSAR-based Antarctica ice velocity map, Version 2. NASA National Snow and Ice Data Center Distributed Active Archive Center, Boulder, Colorado USA, doi:10.5067/D7GK8F5J8M8R.


Table 1

Model parameters with their meaning and values.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$z_0$</td>
<td>Average elevation of the sea level</td>
<td>n.s.(^a)</td>
</tr>
<tr>
<td>$h(t)$</td>
<td>Tidal height relative to $z_0$</td>
<td>CATS2008a_opt</td>
</tr>
<tr>
<td>$W, W_{2013}$</td>
<td>Thickness of the ice shelf ($W_{2013}$ value in 2013)</td>
<td>n.s.</td>
</tr>
<tr>
<td>$z_1(t) = z_0 + h(t) - 0.9 W$, $&lt;z_1&gt;_{2013}$</td>
<td>Elevation of bottom of a freely floating ice shelf ($&lt;z_1&gt;_{2013}$: average in 2013)</td>
<td>n.s.</td>
</tr>
<tr>
<td>$z_2$</td>
<td>Peak elevation of the bedrock protrusion</td>
<td>n.s.</td>
</tr>
<tr>
<td>$\Delta(t) = z_2 - z_1(t)$</td>
<td>Relative elevation of the bedrock peak to the ice shelf bottom</td>
<td></td>
</tr>
<tr>
<td>$H = &lt;\Delta&gt;<em>{2013} = z_2 - &lt;z_1&gt;</em>{2013}$</td>
<td>Height of the pinning point above $&lt;z_1&gt;_{2013}$: average value</td>
<td>0.8 m</td>
</tr>
<tr>
<td>$A(t)$</td>
<td>Area of the contact zone between ice and bedrock</td>
<td>Eq. (2)</td>
</tr>
<tr>
<td>$R_{\text{max}}$</td>
<td>Radius of the maximum contact area</td>
<td>n.s.</td>
</tr>
<tr>
<td>$\tau_0$</td>
<td>Average shear stressing rate</td>
<td>n.s.</td>
</tr>
<tr>
<td>$E_h(t), &lt;E_h&gt;$</td>
<td>Amplitude of the tidal variations ($&lt;E_h&gt;$: average value) calculated from $h(t)$, see Fig. 3a</td>
<td></td>
</tr>
<tr>
<td>$\tau(t) = \tau(t_0) + \int_{t_0}^{t} E_h(\tilde{t})/(&lt;E_h&gt;) d\tilde{t}$</td>
<td>Shear stress (after last event at time $t_0$)</td>
<td></td>
</tr>
<tr>
<td>$\sigma_n(t) = \sigma \Delta(t) / A(t)$</td>
<td>Normal stress ($\sigma$: proportionality factor)</td>
<td></td>
</tr>
<tr>
<td>$f$</td>
<td>Friction coefficient</td>
<td>n.s.</td>
</tr>
<tr>
<td>$CFS(t) = \tau(t) - f \sigma_n(t)$</td>
<td>Coulomb Failure stress</td>
<td>Eq. (3)</td>
</tr>
<tr>
<td>$\bar{CFS} = f c_f H / (\pi R_{\text{max}}^2)$</td>
<td>Normalization factor</td>
<td>n.s.</td>
</tr>
<tr>
<td>$S(t) = CFS(t) / \bar{CFS}$</td>
<td>Normalized $CFS$</td>
<td>Eq. (4)</td>
</tr>
<tr>
<td>$C = \tau_0 / \bar{CFS}$</td>
<td>Normalized (average) shear stressing rate</td>
<td>0.7</td>
</tr>
<tr>
<td>$S_\Delta = \Delta \tau / \bar{CFS}$</td>
<td>Normalized value of the stress drop associated with slip</td>
<td>0.03</td>
</tr>
<tr>
<td>$M = M_i + \log_{10}(A^2/\Delta)$</td>
<td>Event magnitude (constant $M_i$ is set to 0)</td>
<td>Eq. (5)</td>
</tr>
</tbody>
</table>

\(^a\) n.s. refers to values which do not need to be specified.
Fig. 1. Geographic and temporal aspects of observed cryoseismicity. (a) Map of western DML. Grounded ice is in white, floating ice shelf in light blue, ocean in blue and bedrock outcrops in brown. Thin contour lines are topography in 200 m increments. The locations of permanent seismic stations in the region are noted with black squares. IR marks the Kupol Moskovskij ice rise. The red rectangle encloses the source region of the cryoseismic activity shown in panel (b). (b) The cryoseismicity source region. The red circle notes the absolute location result for the source of the large cryoseismic event that occurred on 9 September 2013. The overall uncertainty of this estimate is expressed by the 95% confidence-level error ellipse, shown in red. The colour scale describes ice flow velocity in m/yr and the vectors show direction of flow (Mouginot et al., 2012; Rignot et al., 2011, 2017), scaled to a maximum of 850 m/yr. The white rectangle encloses the area in the inlet, that shows the absolute location epicentre in red and the relative location estimate in black, the arrow noting the direction of epicentre migration. (c) Number of cryoseismic events per day from 2003 to
end 2015 (no detections between 1997 and 2003), based on the results of the CC-detector at SNA (blue) and TROLL (red).

Fig. 2. Examples of highly similar events (multiplets), as recorded at TROLL. Waveforms are bandpass filtered between 2 and 6 Hz. The vertical component is shown for waveforms in black, normalized to maximum amplitude. Waveforms in grey show the vertical and rotated horizontal components of the located event on 9 September 2013, scaled to the maximum amplitude. Note the well-developed S-wave onset on the transverse component (SH).
Fig. 3. Time series of observations (blue) in comparison to simulations (red). (a) Histogram of daily numbers of events at station TROLL in comparison to the envelope amplitude $E_n$ of the tidal signal (grey line), as predicted by the CATS2008a_opt tidal model (Padman et al., 2002, 2008) for the cryoseismicity source region, between 5 August and 24 October 2013. (b) Event magnitudes versus occurrence times, where the ocean tides are shown by grey line. (c) The average magnitude difference, $\langle \Delta M \rangle$, between successive events within temporal clusters as function of time, indicating negative trends during spring tides and increasing magnitudes during neap tides. Each point refers to the result of one cluster, which is defined by a sequence of minimum 5 events with inter-event times less than 6 hours. (d),(e) Two examples of the magnitude trend within individual clusters, each cluster represented by one point in (c).
Fig. 4. Observed (blue) and modelled (red) characteristics. (a) Distribution of inter-event times between successive events. (b) Histogram of the relative occurrence times of events between two successive maxima of the tides (phase between 0° and 360°). The blue line indicates the result if a delay of 30 min is assumed for the true tides with respect to the tides calculated by the CATS2008a_opt model (Padman et al., 2002, 2008). (c) The average change of the delay time, ΔCC, between the detections at stations TROLL and SNAA, as function of the time difference between events. The error bars refer to the same result plus/minus one standard deviation for the randomly reshuffled ΔCC-times. The observed trend indicates migration which is in agreement with the linear trend in the simulations, where Δx is the difference of the epicentre position and $R_{max}$ the maximum radius of the contact area. (d) Magnitude of the events as function of the tidal height.
Fig. 5. Model setup and ice thickening as explanation for the long-term trend. (a) Illustration of the assumed stick-slip motion at a bedrock protrusion (pinning point), where the ice shelf becomes locally grounded. The pinning point is depicted in a generic form, with exaggerated topography to facilitate recognition of model parameters. 90% of the shelf is below the average sea level (a.s.l.). Tidal variations ($h$) change the normal and shear stress, as well as the contact area between bedrock and shelf ice. $H$ defines the height of the shallow bathymetric feature relative to the average position of the bottom of the ice shelf in the period between August and October 2013. Here, increasing tides are assumed ($dh/dt > 0$),
so the contact area decreases due to the decreasing difference $\Delta = H - h$ (see section 2.5), leading to a migration of the hypocentres (stars) and reduced rupture areas (black/red lines).

(b) Event magnitudes versus occurrence times recorded between 2012 and end 2015 (grey dots). The magnitude trend is found to correlate with the timing of the events within the tidal cycles. The solid, blue curve shows the maximum tidal values that triggered events, while the dashed, blue line indicates the maximum tidal amplitude at this time (scale on right: calculated for time bins of 30 days with a minimum of 50 events). The plot on the right shows the maximum triggering tide as function of ice-shelf thickness decrease in the model simulations (without changing any other parameters). (c) Illustration summarising the general effects of varying ice thickness in different time periods.