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- Review of cryogenic seismic sources
- Current and emerging methods in cryoseismology
- Potential to explore structural monitoring of glacier ice

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Cryoseismology

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Abstract The last decade witnessed an explosion in yearly number of publications on passive glacier seismology. The seismic signals from a wide range of glacier-related processes fill a broad band of frequencies (from 10^{-3} to 10^2 Hz) and moment magnitudes (from $M-3$ to $M7$) providing a fresh and unprecedented view on fundamental processes in the cryosphere. New insights into basal motion, iceberg calving, glacier, iceberg, and sea ice dynamics, and precursory signs of unstable glaciers and ice structural changes are being discovered with seismological techniques. These observations offer an invaluable foundation for understanding ongoing environmental changes and for future monitoring of ice bodies worldwide. In this review we discuss seismic sources in the cryosphere as well as research challenges for the near future. The field of glacier seismology is evolving so rapidly that some parts of this review will likely soon be outdated. Nevertheless, given an overwhelming number of recent publications and rapidly growing seismic data volumes provided by modern seismic installations in polar and mountain regions, this introduction to cryosphere seismicity aims to serve as a timely and comprehensive reference for glaciologists and seismologists.

1. Introduction

The remarkable spectacle of glacier crevassing has been known to polar explorers since the end of the nineteenth century, when *Nansen* [1897] recorded the following comment on the phenomenon, "... a noise like the discharge of guns ... and the earth tremble." Nevertheless, until recently, massive ice bodies such as in Greenland and Antarctic have received relatively little interest from seismologists due to low levels of tectonic earthquake activity in the polar regions. For example, even though ice-associated tremors in Antarctic [Hatherton and Evison, 1962; Adams, 1969] were reported decades before the first tectonic events recorded in the region [Adams *et al.*, 1985; Adams, 1988], cryospheric seismicity was not an active field of research in the late 1900s. This has dramatically changed since the first report of so-called "glacial earthquakes" (henceforth GEQ) [Ekström *et al.*, 2003, 2006]. The discovery of these globally detectable seismic events associated with ice discharge to the ocean sparked multiple theoretical, experimental, and observational studies of cryogenic seismic and icequake phenomena (by "icequakes" we mean coseismic brittle fracture events within the ice). Subsequently, the number of relevant scientific investigations rose dramatically (Figure 1), leading to the establishment of new, internationally funded seismic networks and numerous conference sessions specifically focused on glacier seismology (e.g., at American Geophysical Union Fall Meeting 2010, European Geosciences Union General Assembly 2011, Japan Geoscience Union Meeting 2015 and 2016, International Union of Geodesy and Geophysics General Assembly 2013 and 2015, and European Seismological Commission General Assembly 2016). Since the first studies [Röthlisberger, 1955; Crary, 1955] more than 150 papers have been published on the subject, with a majority of articles appearing after a seminal glacial-earthquake paper by Ekström *et al.* [2003]. Their initial interpretation in terms of sudden glacier surges as the source mechanism required 10 m displacements of ice volumes on the order of 10 km^3 within a minute or less. Although this seemed unrealistically large, most surface displacement measurements on glaciers and ice streams at that point did not have the adequate temporal resolution to reject this hypothesis. Consequently, the discovery of glacial earthquakes led to the development of a wide spectrum of novel geophysical studies. This exemplifies how the introduction of new analytical tools can inspire new perspectives on glacier tectonics and dynamics. Here we provide a review of this rapidly growing body of geophysical knowledge, applicable not only across the vast variety of glaciated regions of Earth but also, possibly, on icy moons of the Solar System with active plate tectonics such as Europa and Enceladus [Nimmo and Manga, 2009; Zhan *et al.*, 2014]. Also, it is worth mentioning that many cryoseismology methods, particularly for signal detection and processing,

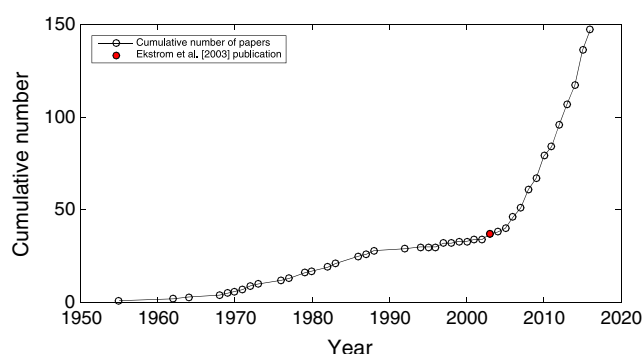


Figure 1. Cumulative number of papers on passive glacier seismology, 1950–2016.

are applicable to volcano seismology and studies of Episodic Tremor and Slip (ETS; nonvolcanic tremor), with little (if any) modification; the physics of various icequake and seismic tremor types also bear strong similarities to the physics of unusual events seen at volcanoes [Konstantinou and Schlindwein, 2003].

Our main objective is to create a topical map of this growing field, to further bridge the interdisciplinary gap between the seismology and glaciology communities. We begin this overview by introducing the principal types of seismic emissions in the cryosphere and their relevant features, followed by a discussion of processes and properties within glacier ice inferred through seismic signals. Then we offer our views on remaining challenges. Subsequently, in Appendix A we outline the basic, commonly used analytical methods, techniques of data handling, and instrumentation and describe selected long-term seismic networks in glacial regions. Our main focus is on seismic signals originating from glaciers and ice sheets; accordingly, we only briefly cover topics of sea/lake ice and do not consider snow avalanches and permafrost/frost heaving for which seismology is becoming increasingly important, too.

2. Seismic Source Processes

To avoid the possibility of this review becoming a simple catalog of literature citations, we have listed all of the main bibliographic references in Table 1, which also gives major types of cryoseismic signals and corresponding studies. The sources are ordered in Table 1 according to their “popularity” (expressed here as the number of papers focusing on a particular phenomenon). Their main categories of events are shown schematically in Figure 2.

2.1. Surface: Crevasses Formation

One of the most prominent types of glacier seismicity is associated with the extension and formation of surface crevasses (Table 1 and Figure 3). Crevasses form when tensile stresses (or, equivalently, strain rates) exceed some depth-dependent fracture criterion. Consequently, crevasses preferentially grow near the surface, where the ice-overburden pressure is low or when existing crevasses are filled with water. For a comprehensive review on these principles and the topic of crevasses in general, the reader is referred to Colgan *et al.* [2016].

Typically, surface crevasse icequakes have negative seismic magnitudes (negative values are permissible due to a logarithmic definition of a magnitude; see Table A2) and their signals occupy the frequency range between 10 and 50 Hz [Neave and Savage, 1970; Walter *et al.*, 2008; Deichmann *et al.*, 2000; Mikesell *et al.*, 2012; Roux *et al.*, 2008]. Their waveforms usually exhibit purely compressive *P* wave polarity, which reflects the dominance of tensile faulting [Walter *et al.*, 2009; Colgan *et al.*, 2016]. Compared to seismic sources at greater depth (see following sections), low-frequency Rayleigh waves dominate surface icequake waveforms [Deichmann *et al.*, 2000; Walter *et al.*, 2009; Rössli *et al.*, 2014]. Icequake Rayleigh waves travel at the glacier surface, are superpositions of *P* and *S* waves, and require a critical angle between takeoff ray path and vertical axis [Lay and Wallace, 1995]. For glacier microseismicity and local on-ice arrays, this often results in the criterion that the hypocenters’ depth does not exceed one wavelength, that is $\lambda_x = V_R / f$ [Deichmann *et al.*, 2000; Stuart *et al.*, 2005]. As a consequence, crevasse icequakes tend to have simple waveforms consisting almost exclusively of the Rayleigh phase (Figure 3c). In contrast, during large-scale fracture such as recorded during the detachment of icebergs, the coalescence of individual crevasses gives rise to sustained surface wave trains (see section 2.6).

Table 1. Main Types of Seismic Emissions/Sources in the Cryosphere (With Corresponding Studies)^a

Type of Seismic Source	Bibliographic References
Glacial earthquakes (GEQ)	<i>Qamar and St. Lawrence</i> [1983], <i>Ekström et al.</i> [2003], <i>Ekström et al.</i> [2006], <i>Tsai and Ekström</i> [2007], <i>Amundson et al.</i> [2008], <i>Joughin et al.</i> [2008], <i>Tsai et al.</i> [2008], <i>Nettles et al.</i> [2008], <i>Larmat et al.</i> [2008], <i>Amundson et al.</i> [2010], <i>Nettles and Ekström</i> [2010], <i>Chen et al.</i> [2011], <i>Walter et al.</i> [2012], <i>Veitch and Nettles</i> [2012], <i>Walter et al.</i> [2013b], <i>Sergeant et al.</i> [2016]
Calving of icebergs, ice mélange dynamics, seiches	<i>Hatherton and Evison</i> [1962], <i>VanWormer and Berg</i> [1973], <i>Qamar and St. Lawrence</i> [1983], <i>Wolf and Davies</i> [1986], <i>Qamar</i> [1988], <i>Gaull et al.</i> [1992], <i>Torcal et al.</i> [1999], <i>MacAyeal et al.</i> [2006], <i>O'Neel et al.</i> [2007], <i>O'Neel and Pfeffer</i> [2007], <i>O'Neel et al.</i> [2010], <i>Richardson et al.</i> [2010], <i>Walter et al.</i> [2010a], <i>de Juan et al.</i> [2010], <i>Amundson et al.</i> [2012], <i>Bartholomäus et al.</i> [2012], <i>Kanao et al.</i> [2012b], <i>Köhler et al.</i> [2012], <i>Richardson et al.</i> [2012], <i>Zoet et al.</i> [2013a], <i>Norman et al.</i> [2013], <i>Walter et al.</i> [2013b], <i>Koubova</i> [2015], <i>Bartholomäus et al.</i> [2015a], <i>Sergeant et al.</i> [2016], <i>Mei et al.</i> [2016]
Stick-slip motion, ruptures at ice-bedrock interface, basal motion	<i>VanWormer and Berg</i> [1973], <i>Weaver and Malone</i> [1979], <i>Blankenship et al.</i> [1987], <i>Anandakrishnan and Bentley</i> [1993], <i>Anandakrishnan and Alley</i> [1994, 1997b], <i>Anandakrishnan and Alley</i> [1997a], <i>Smith</i> [2006], <i>Danesi et al.</i> [2007], <i>Roux et al.</i> [2008], <i>Walter et al.</i> [2008], <i>Wiens et al.</i> [2008], <i>Walter et al.</i> [2009], <i>Winberry et al.</i> [2009a], <i>Walter et al.</i> [2010b], <i>Walter et al.</i> [2011], <i>Winberry et al.</i> [2011], <i>Zoet et al.</i> [2012], <i>Thelen et al.</i> [2013], <i>Allstadt and Malone</i> [2014], <i>Barcheck et al.</i> [2013], <i>Smith et al.</i> [2013], <i>Moore et al.</i> [2013], <i>Winberry et al.</i> [2013], <i>Walter et al.</i> [2013a], <i>Pomeroy et al.</i> [2013], <i>Pratt et al.</i> [2014], <i>Helmstetter et al.</i> [2015a], <i>Smith et al.</i> [2015], <i>Roeoesli et al.</i> [2016a]
Near surface crevassing	<i>Röthlisberger</i> [1955], <i>Lewandowska and Teisseyre</i> [1964], <i>Neave and Savage</i> [1970], <i>Röthlisberger</i> [1972], <i>von der Osten-Woldenburg</i> [1990], <i>Anandakrishnan and Alley</i> [1997b], <i>Helmstetter et al.</i> [2001], <i>Walter et al.</i> [2008, 2009], <i>Smith et al.</i> [2010], <i>Mikesell et al.</i> [2012], <i>Dalban Canassy et al.</i> [2012], <i>Dalban Canassy et al.</i> [2013], <i>Moore et al.</i> [2013], <i>Barruol et al.</i> [2013], <i>Podolskiy et al.</i> [2016]
Intermediate and deep brittle fractures within glaciers	<i>Deichmann et al.</i> [1979], <i>Deichmann et al.</i> [2000], <i>Walter et al.</i> [2008, 2009, 2010b], <i>Dalban Canassy et al.</i> [2013], <i>Röösli et al.</i> [2014], <i>Helmstetter et al.</i> [2015b], <i>Hammer et al.</i> [2015], <i>Lombardi et al.</i> [2016]
Snow and firn cracking	<i>St. Lawrence and Bradley</i> [1977], <i>Denhartog</i> [1982], <i>Nishio</i> [1983], <i>Lough et al.</i> [2013]
Glacier rumbling; tremor	<i>Eckstaller et al.</i> [2007], <i>Rial et al.</i> [2009], <i>MacAyeal et al.</i> [2008], <i>Kanao et al.</i> [2012a], <i>Winberry et al.</i> [2013], <i>Röösli et al.</i> [2014], <i>Heeszel et al.</i> [2014b], <i>Helmstetter et al.</i> [2015b], <i>Bartholomäus et al.</i> [2015b], <i>Lipovsky and Dunham</i> [2015a], <i>Roeoesli et al.</i> [2016b], <i>[Kanao, 2015]</i> , <i>Roeoesli et al.</i> [2016b]
Rifting process in ice shelves	<i>Bassis et al.</i> [2007, 2008], <i>Chen et al.</i> [2011], <i>Heeszel et al.</i> [2014a]
Ice falls and serac collapses	<i>Weaver and Malone</i> [1979], <i>Qamar and St. Lawrence</i> [1983], <i>Roux et al.</i> [2008], <i>Jónsdóttir et al.</i> [2009], <i>Dalban Canassy et al.</i> [2012]
Hydraulic transients in subglacial conduits	<i>St. Lawrence and Qamar</i> [1979], <i>Winberry et al.</i> [2009b]
Icebergs (collisions with ocean bottom, self-oscillations, breakup, etc.)	<i>Hunkins</i> [1960], <i>Kristensen et al.</i> [1982], <i>Müller et al.</i> [2005], <i>Okal and MacAyeal</i> [2006], <i>MacAyeal et al.</i> [2006], <i>Eckstaller et al.</i> [2007], <i>MacAyeal et al.</i> [2008], <i>MacAyeal et al.</i> [2009], <i>Martin et al.</i> [2010] (and references within), <i>Richardson et al.</i> [2010, 2012], <i>Norman et al.</i> [2013], <i>Dziak et al.</i> [2013], <i>MacAyeal et al.</i> [2015], <i>Pirli et al.</i> [2015]
Moulins and thaw water drainage	<i>Walter et al.</i> [2013c], <i>Röösli et al.</i> [2014], <i>Roeoesli et al.</i> [2016b], <i>Carmichael et al.</i> [2015].

Table 1. (continued)

Type of Seismic Source	Bibliographic References
Precursory icequakes before breaking off of hanging glaciers or icefalls	<i>Weaver and Malone</i> [1979], <i>Gaull et al.</i> [1992], <i>Caplan-Auerbach and Huggel</i> [2007], <i>Huggel et al.</i> [2007], <i>Faillietaz et al.</i> [2008], <i>Faillietaz et al.</i> [2011], <i>Dalban Canassy et al.</i> [2012, 2013]
Glacial lake outburst floods (GLOF), lake water discharge, hydraulic fractures	<i>Deichmann et al.</i> [1979], <i>Nye et al.</i> [1995], <i>Walter et al.</i> [2008], <i>Das et al.</i> [2008], <i>Winberry et al.</i> [2009b], <i>Roux et al.</i> [2010], <i>Walter et al.</i> [2013d], <i>Morgan et al.</i> [2013], <i>Jones et al.</i> [2013], <i>Doyle et al.</i> [2013], <i>Carmichael et al.</i> [2015].
Fluid-filled cracks or cavities	<i>St. Lawrence and Qamar</i> [1979], <i>Anandakrishnan and Alley</i> [1997b], <i>Métaxian</i> [2003], <i>Müller et al.</i> [2005], <i>Stuart et al.</i> [2005], <i>O'Neel and Pfeffer</i> [2007], <i>Winberry et al.</i> [2009b], <i>Richardson et al.</i> [2010], <i>West et al.</i> [2010], <i>Lipovsky and Dunham</i> [2015b], <i>Rössli et al.</i> [2014], <i>Helmstetter et al.</i> [2015b]
Surges of glaciers	<i>Raymond and Malone</i> [1986], <i>Stuart et al.</i> [2005], <i>Köhler et al.</i> [2015]
Ice shelf flexural waves (caused by oceanic wave forcing, upstream stick slip, etc.), ice shelf disintegration precursory seismicity?	<i>Hatherton and Evison</i> [1962], <i>Cathles et al.</i> [2009], <i>Bromirski et al.</i> [2010, 2015], <i>Kanao et al.</i> [2012b], <i>Zoet et al.</i> [2013a], <i>Wiens et al.</i> [2016]
Sea and lake ice (due to thermal expansion, flexure, ice fracturing, discharge etc.)	<i>Crary</i> [1955], <i>Hunkins</i> [1960], <i>Goto et al.</i> [1980], <i>Kaminuma</i> [1994], <i>Kanao and Kaminuma</i> [2006], <i>Ruzhich et al.</i> [2009], <i>MacAyeal et al.</i> [2009], <i>Makkonen et al.</i> [2010], <i>Tsai and McNamara</i> [2011], <i>Läderach and Schlindwein</i> [2011], <i>Kanao et al.</i> [2012a], <i>Carmichael et al.</i> [2012], <i>Marsan et al.</i> [2012]
Unidentified or nonconclusive	Alaska glaciers: <i>Rogers</i> [1976], <i>Harrison et al.</i> [1986]; Antarctica: <i>Sinadinovski et al.</i> [1999], <i>Kanao and Kaminuma</i> [2006], <i>Kanao et al.</i> [2012b]; volcano: <i>Weaver and Malone</i> [1976]
Noise and microseisms (inversion of ice shelf thickness, moulins' drainage, sea ice variability, floating state transitions, etc.)	<i>Hatherton</i> [1960], <i>Stutzmann et al.</i> [2009], <i>Cathles et al.</i> [2009], <i>Aster et al.</i> [2009], <i>Walter et al.</i> [2010a], <i>Grob et al.</i> [2011], <i>Zhan et al.</i> [2014], <i>Walter et al.</i> [2013c], <i>Anthony et al.</i> [2015], <i>Mordret et al.</i> [2016], <i>Pratt et al.</i> [2016]
Earthquake triggered glaciaseismicity	<i>Peng et al.</i> [2014]

^aNote that some sources may be not definitive and remain unclear. Some papers may appear several times due to reports on different sources. Indeed, other sources exist, like tectonic earthquakes [Adams et al., 1985], snow/ice avalanches [Vilajosana et al., 2007], permafrost and cultural noise (e.g., on-ice aircraft landing [MacAyeal et al., 2009], grooming of the snow runways [Anthony et al., 2015; Diez et al., 2016]). Multiple glacier studies were done at night in order to take advantage of the lower noise level [Neave and Savage, 1970; Blankenship et al., 1987]; however, these sources were omitted because they fall beyond the main scope of the paper.

Surface crevasse icequakes are a valuable tool for assessing strain rates and changes thereof. For high-advection ice flow, surface crevasses do not represent local stresses, because they are more quickly transported away from their birthplace than it takes them to close in regions of reduced strain [Colgan et al., 2016]. Consequently, near-surface icequakes more accurately identify regions where the fracture criterion is met than visual mapping of surface crevasses [Dalban Canassy et al., 2013]. Temporal changes in icequake activity can therefore reveal sudden reverse of extension and compression principle strains in response to changing boundary conditions such as the drainage of an ice marginal or a supraglacial lake [Roux et al., 2010; Carmichael et al., 2015]. Likewise, microseismic activity on grounded tidewater glacier tongues reflects tidally induced, semidiurnal strain rate variations [Podolskiy et al., 2016].

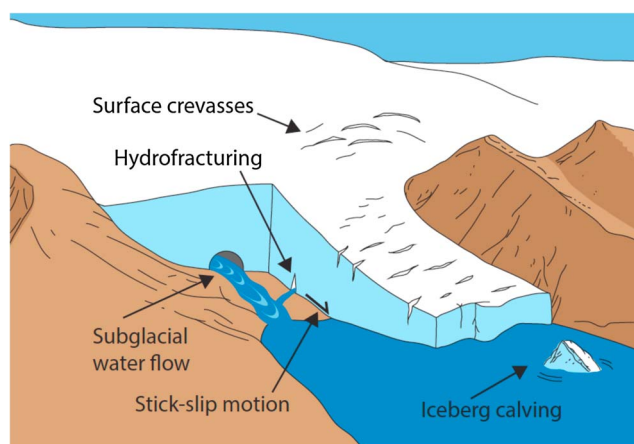


Figure 2. Cartoon illustration of prominent seismic sources in the cryosphere. Adapted from the Journal of Applied Geophysics with permission of Elsevier [Larose *et al.*, 2015].

The mere occurrence of crevasse icequakes shows that the development of surface crevasses occurs via discrete bursts rather than continuous growth. Monitoring icequake multiplets furthermore allows tracking the propagation of crevasse tips. Propagation velocities measured this way cover a broad range, at least between 30 m s^{-1} [Neave and Savage, 1970] and 0.01 m s^{-1} [Mikesell *et al.*, 2012].

Surface crevasse icequakes present a considerable challenge to studies targeting other types of glacier short-duration ($\leq 1 \text{ s}$) microseismic events. For example, the identification of sporadic basal seismic events beneath temperate Alpine glaciers is far more laborious during summer months when melt-enhanced sliding increases strain rates and thus surface crevasse seismicity [Dalban Canassy *et al.*, 2016]. In these cases, high-pass filtering ($> 100 \text{ Hz}$), rectilinearity calculation, and ellipticity measurements can help automatically identify or eliminate near-surface events with dominant Rayleigh phases, whose particle motions describe retrograde

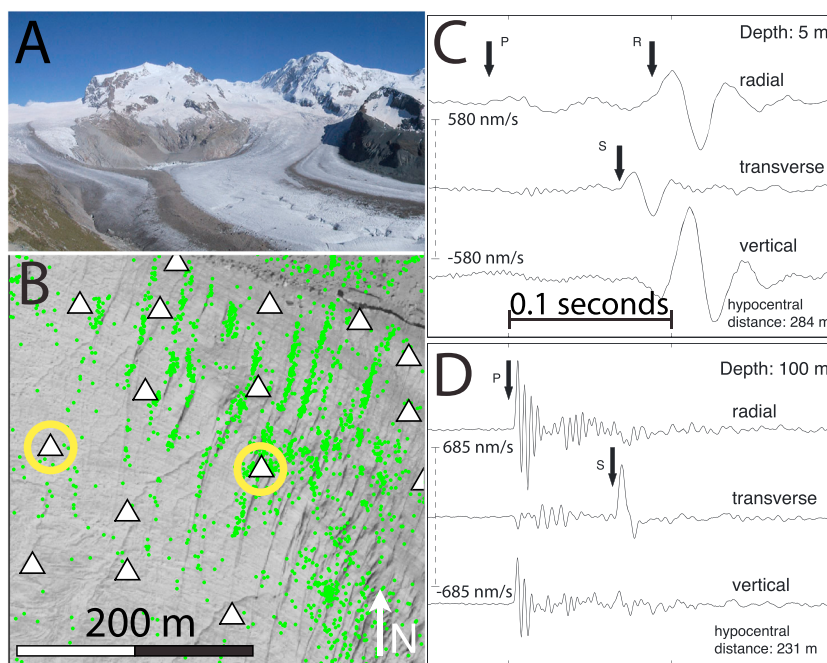


Figure 3. Alpine icequakes on Gornergletscher, Switzerland. (a) The tongue of Gornergletscher (archive of VAW, ETH Zürich). (b) Near-surface icequakes aligning with surface crevasses shown on the orthophoto in the background (icequake location uncertainty 5–10 m). (c) Waveforms of a near-surface icequake showing a dominant Rayleigh wave and (d) an intermediate-depth icequake showing dominant *P* and *S* waves. Modified from the Bulletin of Seismological Society of America [Walter *et al.*, 2009, ©Seismological Society of America].

ellipses [Pomeroy *et al.*, 2013; Walter *et al.*, 2008; Deichmann *et al.*, 2000; Helmstetter *et al.*, 2015a; Dalban Canassy *et al.*, 2013; Métaixian, 2003; Carmichael *et al.*, 2015].

The dominance of surface crevasse seismicity is site dependent. In high-melt areas or steep terrain with high strain rates, near-surface events constitute some 95–99% of short-duration seismic detections [e.g., Métaixian, 2003; Walter *et al.*, 2009; Röösli *et al.*, 2014; Carmichael *et al.*, 2015]. In contrast, among 120 icequakes detected and located during a 3 week monitoring period on Triftgletscher, Switzerland, Dalban Canassy *et al.* [2013] classified 33 and 87 icequakes as deep and shallow, respectively. The low number of detections and high proportion of basal events most likely reflect local strain rates, basal conditions and microseismic noise. Finally, basal stick-slip microseismicity seems to be dominant in some parts of Antarctic, where high subglacial water pressures favor rapid basal motion and flow resistance is confined to so-called sticky spots [Anandakrishnan and Alley, 1994; Smith, 2006; Smith *et al.*, 2015].

Though often a nuisance to passive seismic investigations on glaciers and ice sheets, surface crevasse seismicity can serve as a strain gauge and is therefore valuable to studies of steep and potentially unstable glaciers [Faillettaz *et al.*, 2008; Dalban Canassy *et al.*, 2013] and tidewater termini [Podolskiy *et al.*, 2016]. In addition, seismology may be the only tool to study different englacial ice failure modes (i.e., tensile, shear, and mixed tensile-shear) directly in the field, since these are hard to monitor due to difficult access and limitations of temporal resolution. On a different note, the attenuating and slowing effect, which surface crevasses have on the propagation of surface waves, which traverse them [Walter *et al.*, 2009; Stuart *et al.*, 2005], could be leveraged in future studies to investigate changes in englacial fracturing. Finally, near-surface seismicity associated with fractures within firn has received relatively little scientific attention [St. Lawrence and Bradley, 1977; Denhartog, 1982; Nishio, 1983; Lough *et al.*, 2013]. In principle, however, this opens the possibility to investigate how different types of snow are affected by fracture growth, which plays a key role in avalanche triggering [Schweizer *et al.*, 2003].

2.2. Basal Shear Sources

Basal motion (including basal sliding and sediment deformation) plays a central role in ice dynamics and has a significant influence on how glaciers and ice sheets behave in a changing climate [e.g., Zwally *et al.*, 2002; Schoof, 2010; Tsai *et al.*, 2015; Ritz *et al.*, 2015]. Initial theories of glacier sliding were based on viscous deformation and regelation (i.e., pressure melting and refreezing) processes at the ice-bed interface [Weertman, 1957]. Moreover, evidence from field measurements and numerical simulations suggests that cavity formation on the lee side of bed undulations promotes basal sliding [Lliboutry, 1968; Iken, 1981; Anderson *et al.*, 1982; Fountain and Walder, 1998]. Therefore, a number of mathematical formulae relating basal sliding to subglacial conditions (often called “sliding laws”) have incorporated the presence of pressurized water in basal cavities [see Zoet and Iverson, 2015, and references therein].

In recent decades, various observations worldwide have suggested that glacier basal motion is not always a smooth process but also may take place in the form of sudden stick-slip events. Theoretical considerations are only beginning to include stick-slip events into quantitative descriptions of glacier sliding [Goldberg *et al.*, 2014]. If stick-slip motion were a widespread form of glacier sliding, traditional sliding theories would have to be reevaluated. A thorough understanding of glacier sliding is therefore of pivotal importance to understanding ice dynamics. Furthermore, basal bedrock erosion is known to increase with sliding velocity, and stick-slip behavior can suddenly decrease basal water pressures to drastically enhance the plucking mechanism (i.e., glacial erosion by plucking of large blocks from the bed) [Zoet *et al.*, 2013b].

Seismology has played a central role in modern studies of basal motion, because, analogous to earthquakes on tectonic faults, stick-slip motion across “glacial faults” emits seismic energy. Furthermore, in the 1990s, it was recognized that fast ice stream sliding is made possible by water-saturated, easily deformable subglacial till [e.g., Anandakrishnan *et al.*, 1998; Alley *et al.*, 1986; Iverson *et al.*, 1995; Iverson, 2010]. This corresponds to tectonic fault gouges [Morrow *et al.*, 2000]. In the following subsections, we summarize seismic observations of glacial stick-slip motion, beginning with the best studied case of Whillans Ice Stream, Antarctic.

2.2.1. Stick-Slip Beneath Whillans Ice Stream

Whillans Ice Stream (WIS; formerly known as Ice Stream B) flows from the Transantarctic Mountains to the Ross Ice Shelf in West Antarctica (Figure 4). Near the Ross Ice Shelf grounding line (i.e., the zone where the ice starts to float), the ice stream flows at around 300 m a^{-1} and forms a region with little surface topography known as the Whillans Ice Plain (WIP) [Whillans *et al.*, 2013].

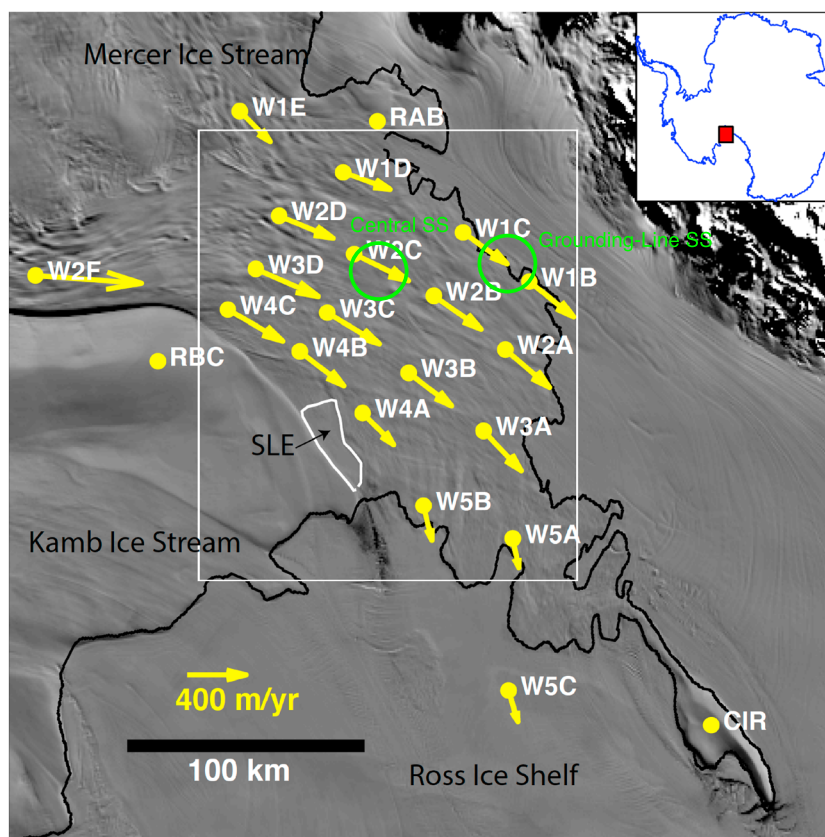


Figure 4. Whillans ice plane and surrounding areas. Black line indicates the grounding line of the Ross Ice Shelf. Global Positioning System (GPS) installations and surface flow vectors are shown in yellow. The white polygon outlines Subglacial Lake Engelhardt (SLE). On the Whillans ice plane (white box), motion is stick-slip dominated. Upflow of the WIP (e.g., at stations W1E and W2F) ice stream flow is smooth. The two green circles denote approximate locations of the central and grounding-line nucleation points (CSS and GLSS, respectively). Adapted from the Earth and Planetary Science Letters with permission of Elsevier [Winberry *et al.*, 2011].

Whereas the upstream areas of WIS and the neighboring Mercer Ice Stream flow smoothly, the entire WIP is characterized by stick-slip motion: most or all of the 300 m a^{-1} ice stream flow occurs during short, 20–30 min motion bursts followed by 6–25 h periods of stagnation. During slip events with total displacements of 0.2–0.5 m, ice flow acceleration to 1 m h^{-1} takes place within approximately 1 min [Bindenschadler *et al.*, 2003], leading to the emission of seismic waves [Wiens *et al.*, 2008].

Slip phases nucleate in two regions, termed the central and grounding-line sticky spots, or CSS and GLSS, respectively (Figure 4). These terms were originally proposed because these regions were located on hydropotential ridges, implying that low pore pressures give rise to stiffer till and thus increased flow resistance relative to surrounding areas [Winberry *et al.*, 2011]. However, as discussed below, recent laboratory work suggests that inhomogeneities in shear stress loading, rather than basal properties, determine the locations of slip nucleation [Walter *et al.*, 2015b].

No interevent motion takes place at the central sticky spot, and minor interevent motion can be measured at the grounding-line sticky spot. Analogous to earthquake faulting, the two locations represent slipping and interevent-locked asperities (i.e., high friction regions or “sticky spots”), respectively [Pratt *et al.*, 2014]. Once initiated, the ice stream slip propagates at varying velocities, with maxima up to 1.5 km s^{-1} and an average of 150 m s^{-1} [Pratt *et al.*, 2014]. Rupture velocity tends to be fastest in flow direction [Winberry *et al.*, 2011; Pratt *et al.*, 2014].

An outstanding feature of WIP stick-slip events is that they have the same temporal cadence with ocean tides at the Ross Ice Shelf [Bindenschadler *et al.*, 2003; Winberry *et al.*, 2009a]. Moreover, there exists a positive correlation between slip-magnitude variations and the length of the recurrence interval prior to the slip event.

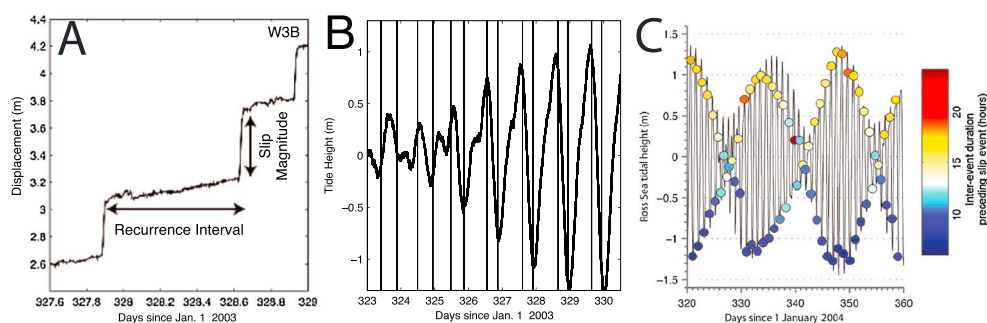


Figure 5. Dynamics and tidal pacing of large-scale stick-slip events on the Whillans Ice Stream, Antarctic. (a) Ice stream displacement during and in between stick-slip events. (b) Tide height and stick-slip event time series (stick-slip events are marked by vertical bars). Reprinted from the Journal of Geophysical Research: Earth Surface with permission of John Wiley and Sons [Winberry *et al.*, 2009a]. (c) Tide height and interevent duration to preceding slip phase (“waiting time”). During large tidal height difference (e.g., during spring tide), near-low-tide events tend to have shorter waiting times than near-high-tide events. This bimodal distribution is suppressed during neap tides. Reprinted from the Journal of Glaciology with permission of the International Glaciological Society [Winberry *et al.*, 2014].

To explain these observations, Winberry *et al.* [2009a] invoked a slider-block model in which interevent strain is stored elastically. This elastic strain is released during a subsequent slip event, which propagates until entering regions of limited interevent strain accumulation. The model assumes that an imbalance exists between upstream loading and tidally influenced resistive pressure at the Ross Ice Shelf (note that in this region the tidal period is close to diurnal rather than semidiurnal). Elastic shear stresses at the glacier base compensate this imbalance [Bindshadler *et al.*, 2003].

The slider-block model of Winberry *et al.* [2009a] shows that tidally driven stress, though small, modifies the rate at which an ice stream bed is loaded and therefore explains timing of stick-slip events (see the next section for more details). Moreover, the model determines a higher accumulated shear stress at failure for longer interevent times. This interpretation suggests that the bed’s yield stress increases over time, suggesting some kind of strengthening or healing mechanism. Various explanations for this strengthening have been proposed, such as freezing [Walter *et al.*, 2011], pore pressure diffusion, and regelation into the bed [Alley, 1993; Iverson, 2010].

Tidal Pacing. Typically, two types of stick-slip events occur on the WIP [Bindshadler *et al.*, 2003; Winberry *et al.*, 2009b, 2011; Pratt *et al.*, 2014] as follows.

1. High-tide events, initiating just after high tide, with interevent periods of 14–19 h. These events nucleate at the interevent-locked central sticky spot at higher stress levels (0.49 kPa), giving rise to higher rupture velocities.
2. Low-tide events, initiating just before low tide, with interevent periods of less than 9 h. These events nucleate at the slipping grounding-line asperity at lower stress levels (0.42 kPa).

In addition to timing, the amplitudes of ocean tides influence stick-slip behavior, particularly interevent times. Consequently, interevent times are bimodal during spring tides and nearly unimodal during neap tides, when tidal phasing approaches 12 h (Figure 5). It is also the neap tide cycle that tends to produce exceptional events that do not fit into either of the two patterns described above. Examples include skipped low-tide events and high-tide slips that nucleate from the grounding-line asperity, rather than the central asperity.

Ongoing Dynamic Changes. Whillans Ice Stream has been decelerating at a rate of 0.6%/yr² (i.e., expressed as a percentage of its mean annual speed) and may stagnate within a century [Joughin *et al.*, 2005]. Two possible reasons for this slowdown are refreezing and increasing basal strength due to decreasing availability of basal meltwater [e.g., Bougamont *et al.*, 2003]. Winberry *et al.* [2014] presented evidence that stick-slip behavior, particularly increasing numbers of skipped events, plays a central role in the slowdown of the WIS. Deceleration between the 2003–2004 and 2010–2011 field seasons near the central sticky spot (CSS) was about 3 times higher than that in nonstick-slip locations. Moreover, during the Austral summer 2010–2011, low-tide events were skipped much more often than in 2003–2004; whereas interevent periods exceeded 20 h almost every other day of the 2010–2011 field season, the same was true for only 1% of days in the 2003–2004 field season.

Of pivotal importance is the observation that a slip event after a 24 h interevent period moves the ice stream only 150% farther than a slip event after a 12 h period. This is in contrast to the 200% slip that would be expected from release of strain stored purely elastically. The missing slip can be explained with a viscoelastic rheology. In this model, during stick phases, some strain loading is accommodated by viscous deformation. As a result, only part of the strain load is released during a slip event. For a 24 h interevent period, the viscoelastic rheology predicts a 150% stress drop increase compared to the 12 h time interval, which is in agreement with the relative slip magnitudes.

The motion lost to viscous deformation on single slip-event days can thus account for the long-term slowdown of the ice stream [Winberry *et al.*, 2014].

Winberry *et al.* [2014] also reported that the relationship between sticky-spot yield stress and loading time did not change between 2004 and 2011. The resistive strength at sticky spots is therefore unlikely to be responsible for changing stick-slip behavior and ice stream deceleration. Instead, the authors found that basal strengthening, and the associated slowdown upstream of the stick-slip region, leads to a decreased stressing rate and thus delayed slip onsets. This promotes skipping of slip phases and consequently explains the reduced strain release during subsequent slip events.

These results exemplify the interaction of seismogenic ice dynamics on short, stick-slip time scales with processes active on longer time scales. Traditional ice flow models, which neglect elastic processes, may not fully capture the dynamic reactions of ice streams to changing boundary conditions, such as basal strengthening or changing resistive pressure near the grounding line. Rising sea levels, for example, may perturb ice stream forcing budgets enough to further inhibit elastic strain release during slip events, thus increasing the role of viscous deformation. The relatively regular stick-slip behavior witnessed at Whillans Ice Stream over the last decade may indeed require a specific and narrow range of boundary conditions; thus, these observations indicate a transitional state associated with ice stream slowdown [Sergienko *et al.*, 2009].

Long-Period Seismic Waves. The WIP slip events produce seismic energy with periods of 20–150 s, detectable at broadband stations up to 1000 km away [Wiens *et al.*, 2008; Walter *et al.*, 2011; Winberry *et al.*, 2011; Pratt *et al.*, 2014]. These teleseismic phases consist of far-field surface waves; up to three such arrivals are visible per event (Figure 6). Whereas the entire ice stream slip has a moment magnitude of 7.0 (the largest magnitude reported in glacier seismology literature), each individual teleseismic phase has a peak amplitude only equivalent to a $M_{3.6}$ – $M_{4.2}$ earthquake [Wiens *et al.*, 2008; Winberry *et al.*, 2013]. The observed teleseismic surface waveforms were well explained by shear faulting at shallow depths (<1 km) [Wiens *et al.*, 2008]. Pulses of compressional displacement from the WIP stick-slip events also excite longitudinal seismic waves in the floating Ross Ice Shelf as far as 700 km away [Wiens *et al.*, 2016].

The initial rupture of a slip event generates the first teleseismic phase. It is discernable only for grounding-line initiation (Figure 4), whereas the second and third phases in Figure 6 can be observed for all types of slip events [Pratt *et al.*, 2014]. These latter two phases are attributed to moment rate increases in ice stream slip, where slip acceleration occurs toward the end of an event. In contrast to changing locations of slip initiation, these later acceleration phases have stationary sources [Pratt *et al.*, 2014; Walter *et al.*, 2011]. Their respective source locations are near the downstream end of subglacial Lake Engelhardt (Figure 4) and the farthest downstream extent of the ice stream [Pratt *et al.*, 2014].

A central question concerning WIP stick-slip events and their seismic radiation patterns is as follows: What controls the temporal and spatial variations in rupture initiation and rupture speed? One explanation could be spatial variations in basal properties, in the sense that the loci of stronger ice couplings could lead to increases in rupture speed, and thus increase the seismic moment rate [Pratt *et al.*, 2014]. For rupture acceleration leading to the second and third teleseismic phases, this suggests a strongly coupled grounding zone, which is in agreement with results from independent studies [Horgan and Anandakrishnan, 2006; Alley *et al.*, 2007; Christianson *et al.*, 2013]. However, recent laboratory work suggests that basal properties may play a minor role [Walter *et al.*, 2015b]. Alternatively, details of the loading process during the ice stream stick phase may control the basal stress distribution, and thus the slip nucleation location and propagation velocity. Stress loading at the ice stream base is likely complicated, as the action of ocean tides with varying amplitudes combines with tidal flexure near the grounding zone [Vaughan, 1994] and with heterogeneous subglacial conditions [Fricker *et al.*, 2007]. However, investigating this matter is of importance beyond the cryosphere, as it may elucidate different dislocation modes on tectonic faults [Beroza and Ide, 2011].

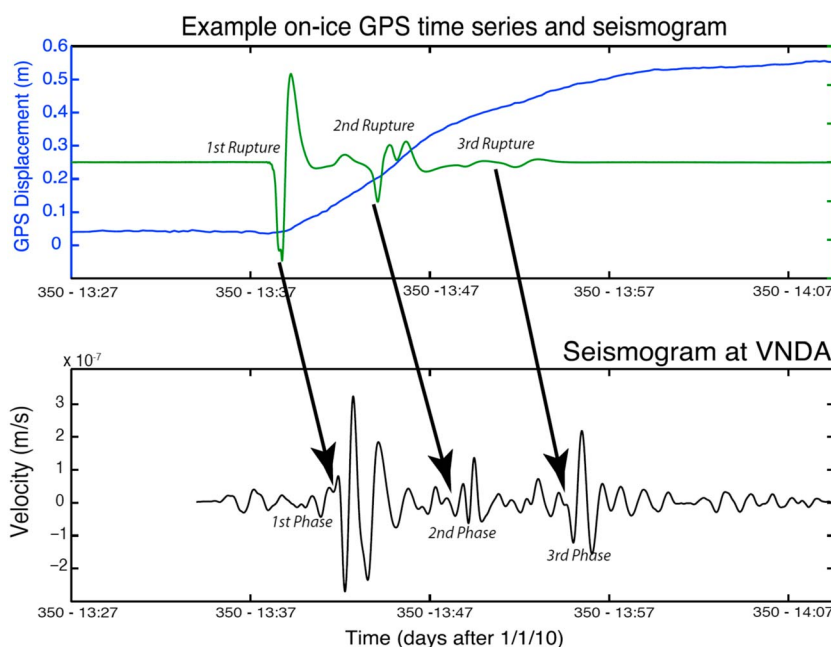


Figure 6. Synchronized GPS and seismic data showing the generation of teleseismic signals during a WIP slip event. (top) Data from a GPS (blue) and a colocated seismometer, both installed directly on the ice stream. (bottom) Seismic data from station VNDA (at almost 1000 km distance) showing the relationship between rupture phases during the slip event and the generation of teleseismic signals. Reprinted from the *Journal of Geophysical Research: Earth Surface* with permission of John Wiley and Sons [Pratt *et al.*, 2014].

High-Frequency Seismicity. Before the discovery of large-scale stick-slip events, the high-frequency signature of ice stream sliding had been observed [von der Osten-Woldenburg, 1990; Anandakrishnan and Bentley, 1993; Anandakrishnan and Alley, 1994, 1997a, 1997b]. These seismic events also exhibit tidal modulation and are indicative of elevated basal resistance. However, they have negative seismic moment magnitudes and occupy frequencies as high as 100–130 Hz [Anandakrishnan and Bentley, 1993; Smith *et al.*, 2015]. This microseismicity is therefore distinct from the previously discussed large-scale stick-slip events. Such high-frequency emissions provide indirect knowledge about differences in material properties between the bases of glaciers and ice streams [Blankenship *et al.*, 1987; Anandakrishnan and Alley, 1994; Winberry *et al.*, 2013]. Therefore, it was suggested that passive seismology could be used to map out spatial and temporal features of basal motion under the ice streams [Smith, 2006; Smith *et al.*, 2015]. Microseismic “sticky spots” (or asperities) close to the ice-bed interface were recently identified under Rutford Ice Stream, Antarctica, and found to correspond to areas of stiff, low-porosity till [Smith *et al.*, 2015].

The high-frequency component of ice stream seismicity has recently attracted new scientific attention, as local on-ice installations have provided unprecedented observations from continuous seismic records. At the onset of WIP slip events, local microseismic activity increases; it subsequently decreases near the termination of slip [Winberry *et al.*, 2013]. This microseismicity takes the form of discrete events (>10 Hz) or emergent tremor. The latter is particularly interesting, as it has gliding harmonic lines whose frequencies correlate with slip velocity (Figure 7a). Most likely, this is a manifestation of repeated shear slip over a single fault, which accumulates to produce large-scale stick-slip. This interpretation is supported by observations showing that energy is mainly concentrated on the horizontal components. The seismic signals of individual shear events overlap, and the interevent times determine the frequencies of the resultant harmonic lines [Lipovsky and Dunham, 2015a]. Such gliding harmonic tremor due to repeated slip has been observed in other cryospheric processes, such as icebergs scraping across the ocean floor [Eckstaller *et al.*, 2007; Dziak *et al.*, 2013] or past each other [MacAyeal *et al.*, 2008] (Figure 7b). In the latter case, installation of seismometers and Global Positioning System (GPS) stations on an iceberg revealed that the frequencies of harmonic tremor lines correlated with the speed at which two touching icebergs drifted past each other. At times when the drift reverses direction, and thus comes to a halt, the pause in tremor activity produces an “aseismic eye.” At the beginning and end of such

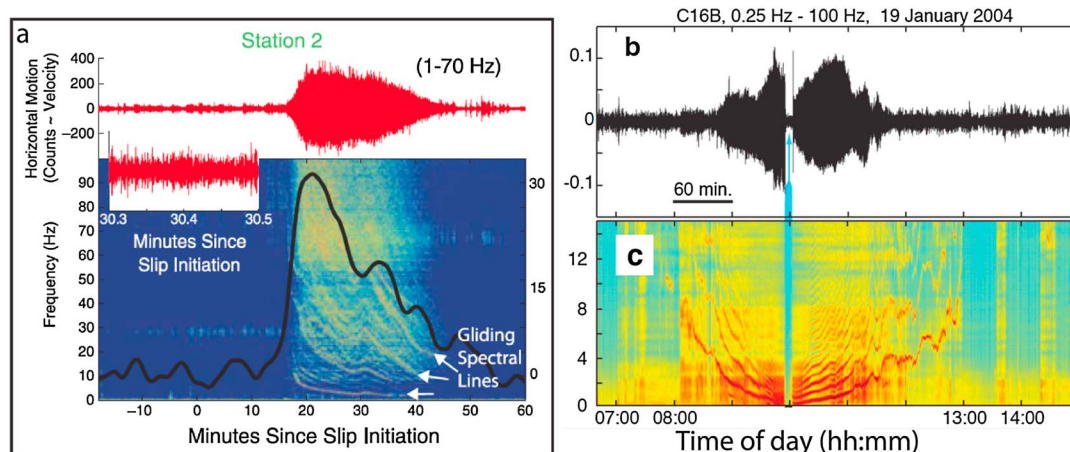


Figure 7. (a) Microseismicity and spectrograms associated with a Whillans Ice Stream slip event and (b) scraping of iceberg C16 past B15A in the Ross Sea. The black line in Figure 7a is the ice stream velocity measured with GPS at the ice surface. The seismic spectrograms show tremor with gliding spectral lines, whose resonance frequencies correlate with ice stream (Figure 7a) or (c) iceberg gliding velocity. Reprinted from the *Geophysical Research Letters* and the *Journal of Geophysical Research: Earth Surface* with permissions of John Wiley and Sons [Winberry *et al.*, 2013; MacAyeal *et al.*, 2008].

an eye, the tremor occurrence rate is low enough that individual shear faulting events clearly separate from each other.

Interestingly, tremor signals related to repeated slip and/or gliding spectral lines have been observed in volcanic regions [Dmitrieva *et al.*, 2013; Powell and Neuberg, 2003; Hotovec *et al.*, 2013]. This phenomenon may therefore be widespread in shear faulting environments. Further investigations would most likely provide new insights into fault physics and the interplay between small- and large-scale asperities during fault rupture.

2.2.2. Stick-Slip Beneath David Glacier

David Glacier is Antarctic ice stream hosting seismogenic stick-slip motion, though on a smaller scale than the WIS events [Danesi *et al.*, 2007; Zoet *et al.*, 2012, 2013b]. It is an outlet glacier in Victoria Land, East Antarctica, flowing into the Drygalski Ice Tongue in the Ross Sea.

Among more than 6000 detected David Glacier events, Danesi *et al.* [2007] analyzed and located 121 low-magnitude seismic events recorded by a broadband network during 3 months of the Austral summer (November 2003 to February 2004). These events (i) had irregular occurrence, (ii) clustered in space (one of three located seismic-event clusters, called “Downstream” or “DW”, occupied heavily crevassed and fast-flowing area of less than 2 km²), (iii) show high waveform similarity (cross-correlation coefficients between 75 DW events exceed 0.95), and (iv) have a characteristic magnitude size (an average M_w was 1.65 for all the events in the DW cluster). In contrast to Gutenberg-Richter relationships of tectonic earthquakes, local magnitudes of the DW cluster followed a Gaussian-like magnitude-frequency distribution.

Although Danesi *et al.* [2007] could not provide accurate source models for their David Glacier events, they proposed repeated rupture of a basal asperity with a radius on the order of 10–100 m, likely related to a promontory in the basement. The cumulated annual coseismic slip was estimated to be on the order of 16 m, which is negligible compared to the annual ice speed of 400–500 m yr^{−1}.

In a subsequent study, Zoet *et al.* [2012] analyzed data from the Transantarctic Mountains Seismic Experiment (TAMSEIS) array and a permanent station Vanda (VNDA) between June 2002 and March 2003. They identified more than 20,000 seismic events, which had highly similar waveforms (cross-correlation coefficient exceeding 0.9), and during a 9 month period a remarkably regular reoccurrence pattern with interevent time about 25 ± 5 min. The estimated focal mechanism was a low-angle thrust fault with a slip in the direction of ice flow. The source epicenter was calculated to lie within the 10 km uncertainty bounds of DW cluster previously studied by Danesi *et al.* [2007]. The estimated mean M_w was 1.8, and the source radius in the order of 350 m.

Similar to the Whillans Ice Stream events, Zoet *et al.* [2012] furthermore found that interevent times of their David Glacier events were modulated by ocean tide: shortest interevent times occurred about an hour after

low tide. Longest interevent times occurred after high tide. Although no surface flow velocity measurements for David Glacier were available, interevent minima and maxima presumably correspond to highest and lowest flow speeds, respectively. Furthermore, seismic source magnitude increased with longer waiting times suggesting fault healing in the interseismic period.

The period of regularly repeating events was preceded and followed by repeating but irregularly occurring events. The transition to regularly repeating earthquakes occurred gradually over about 5 days in June 2002 and the transition back took place over about 11 days in March 2003. Zoet *et al.* [2012] explain this transition by advection of debris-rich basal ice over the asperity.

The abundance of stick-slip events beneath David Glacier prompted a theoretical argument that sliding glaciers may be far more efficient at eroding their bed than previously thought [Zoet *et al.*, 2013b]: during sudden slip, subglacial water-filled cavities may open rapidly enough to locally cause drastic drops in water pressures. Given that fluctuations in subglacial water pressure are believed to promote bedrock fracturing [Cohen *et al.*, 2006], stick-slip events would substantially enhance basal erosion.

2.2.3. Non-Antarctic Stick-Slip Icequakes

Whereas ice stream stick-slip events seem to be a prominent characteristic of ice dynamics in Antarctic, it is not clear how widespread this type of seismicity is beneath other glaciated regions in the world. In fact, a number of studies report a lack of detectable seismicity beneath mountain glaciers in the Northern Hemisphere [Neave and Savage, 1970; Moore *et al.*, 2013; Pomeroy *et al.*, 2013] and the Greenland Ice Sheet or GrIS [Carmichael *et al.*, 2015]. Even if present, basal seismicity beneath alpine glaciers may not be related to sliding, but instead to opening and closing of tensile fractures near the glacier base [Walter *et al.*, 2013a]. The reason for this difference is an open question that is beyond the scope of the present paper and certainly needs further research. Pomeroy *et al.* [2013] offered possible explanations in terms of subglacial till, lower viscosity of warm ice leading to more efficient viscous dissipation of accumulated stress, and difficulty to detect basal signals in temperate glaciers due to low signal-to-noise ratio. Also, Carmichael *et al.* [2015] suggested that basal shear slip is less likely to be detected on the GrIS in typical noise environment relative to surficial icequakes.

Nevertheless, various passive seismic studies have presented evidence for glacier motion in the form of stick-slip events at the base of mountain and tidewater glaciers [VanWormer and Berg, 1973; Weaver and Malone, 1979]. Subglacial observations on the Argentière Glacier support these claims [see Weaver and Malone, 1979, and references therein]. One example is the seismicity on Cascade volcanoes in North America [Weaver and Malone, 1979; Thelen *et al.*, 2013; Allstadt and Malone, 2014]. These seismic events have peak frequencies between ~ 1 and 5 Hz, and magnitudes below 1. While location uncertainties are too large to distinguish between hypocenters near glacier surfaces and beds, the events do originate from glaciated portions of volcano massifs. Moreover, the tendency of these icequakes to form clusters of thousands of events, the regular interevent times (as low as a few minutes), and the high waveform similarities all suggest repeated shear rupture across a glacier bed (Figure 8). *P* wave polarities are mixed, as expected for shear dislocations when sources migrate at speeds expected for glacier flow. Interestingly, the activity of these events increases during snow precipitation, suggesting that additional loading due to snow cover perturbs the glaciers from smooth sliding to stick-slip regime [Allstadt and Malone, 2014].

A recent study on the Glacier d'Argentière, a typical alpine glacier in the French Alps, reported clusters of repetitive seismic events occurring every few minutes [Helmstetter *et al.*, 2015a]. Limited station coverage yielded large location uncertainties of these $\sim M-3.2$ to $M-2.2$ events. However, *P-S* times and *P* wave particle motions suggest hypocenters near the glacier bed. Moreover, the polarities and amplitudes of *P* and *S* waves are more compatible with shear dislocations than tensile faulting, suggesting that these icequakes are indeed stick-slip events. However, to rigorously constrain both hypocentral locations and faulting mechanisms, a better station coverage is needed, which reduces the azimuthal gap and thereby provides detailed records of the *P* and *S* waves radiation patterns which are required to determine a fault plane solution [Stein and Wysession, 2003].

Roesli *et al.* [2016a] recently reported the first seismogenic stick-slip events beneath the Greenland Ice Sheet. The seismic monitoring network was located on a relatively slow-flowing (0.27 m d^{-1}) part of the ice sheet. Beside the dominant near-surface-crevassing seismicity, around 12,000 stick-slip icequakes were detected. These events have negative magnitudes and form clusters of up to 10^3 individual icequakes. More than 100 clusters were detected over a total area of around 10 km^2 .

The Greenland events exhibit first-motion polarities consistent with bed-parallel slip. However, slip direction for some clusters differs substantially from the direction of surface ice flow. Local bed topography or a

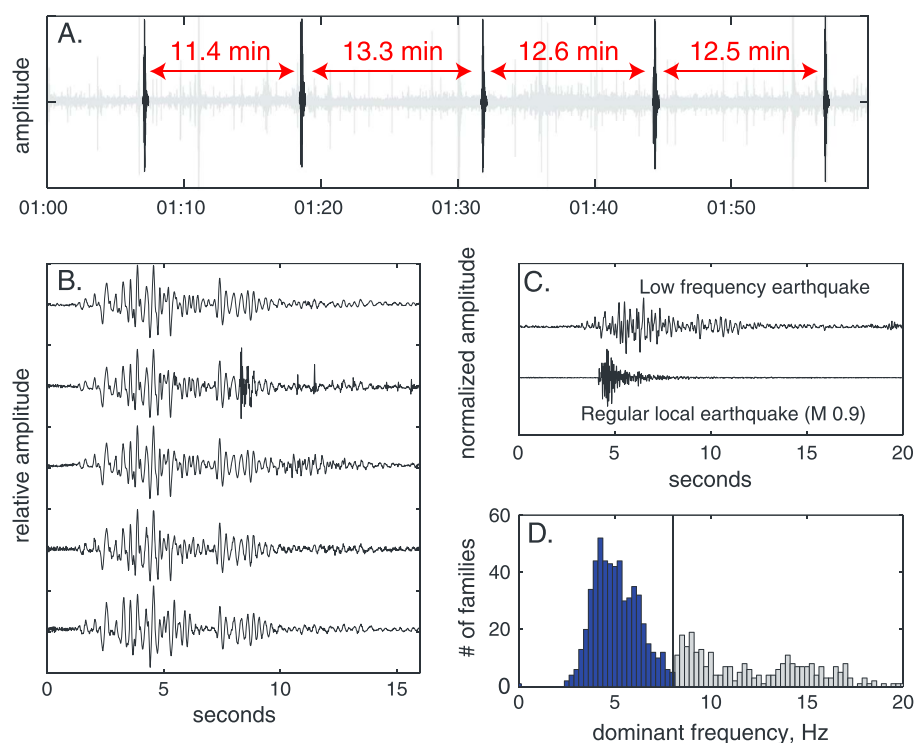


Figure 8. Glacier-related repeating earthquakes on Mount Rainier, USA. (a, b) These events have highly repetitive waveforms and (c, d) exhibit lower frequency content than regular tectonic earthquakes of comparable size. Reprinted from the *Journal of Geophysical Research: Earth Surface* with permission of John Wiley and Sons [Allstadt and Malone, 2014].

complicated basal stress field may be responsible for this discrepancy. An outstanding feature of certain clusters is that the icequake seismic moments show a clear anticorrelation with basal water pressure measured in an efficient subglacial conduit. This can be explained by shear failure within a till layer, whose shear strength depends on pore pressure, which in turn is modulated by diurnal fluctuations in subglacial pressure [Walter *et al.*, 2014]. Although the idea of till weakening due to increasing water pressures is not new, the Greenland stick-slip icequakes arguably provide the first direct observations of subglacial water pressure variations interacting with basal motion.

2.2.4. Basal Nonshear Seismicity

Basal seismic events are rather rare, in some studies comprising less than 0.5–8% of all detectable events [Stuart *et al.*, 2005; Walter *et al.*, 2008; Pomeroy *et al.*, 2013]. So far, three main mechanisms for basal nonshear events have been proposed: (i) basal tensile faulting, (ii) hydraulically induced basal seismicity [Winberry *et al.*, 2009b], and (iii) collapse of an evolving cavity. These processes may coexist as a result of sustained basal fracture networks for water transport [Dalban Canassy *et al.*, 2016]. Walter *et al.* [2013a] provide a detailed discussion of possible processes at the bases of Alpine (temperate) glaciers.

Nonshear basal seismicity most likely requires the presence of pressurized water near the glacier bed, so that tensile faulting occurs in the presence of high ice-overburden pressures [Van der Veen, 1998]. A possible relationship between water pressure oscillation and glaciogenic seismic activity was already noted by Röthlisberger [1972]. More recent studies showed that low water pressure at the base of Gornergletscher, Switzerland, promote basal fracture opening and closing icequakes. This reinforces the idea that such icequakes depend on diurnal changes in subglacial hydraulics, which also alter sliding across the glacier bed [Walter *et al.*, 2008].

2.3. Iceberg Calving

Similar to basal motion, iceberg calving is a key process of dynamic discharge, i.e., the way through which ice sheets transfer ice mass to the ocean. In general, iceberg calving is the loss of ice mass to a proglacial water body, either a lake or the tidewater (see Benn *et al.* [2007] for a review). Together with submarine melt [Truffer and Motyka, 2016] it contributes to frontal retreat, which is responsible for half of Greenland's and

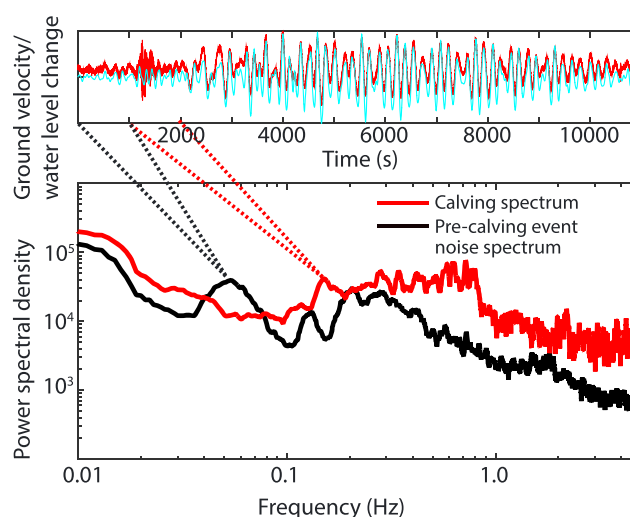


Figure 9. Seismic signature of a calving event at Kangerdlugssup Sermerssua (Greenland), recorded on the broadband station NUUG of the Greenland Ice Sheet Monitoring Network (GLISN). (top) Horizontal north-south component of seismic time series (red) and nearby measured ocean water level (cyan). Both time series clearly show the hourlong water level oscillation (“fjord seiche”) triggered by the iceberg calving event. (bottom) Spectrum of the calving event’s onset (red) and preevent noise (black). Adapted from the *Journal of Glaciology* with permission of the International Glaciological Society [Walter *et al.*, 2013b].

nearly all of Antarctic’s mass loss [Rignot and Kanagaratnam, 2006; Rignot *et al.*, 2008a; van den Broeke *et al.*, 2009; Enderlin *et al.*, 2014; Rignot, 2006; Rignot *et al.*, 2008b]. For grounded tidewater glaciers, the volume loss during individual calving events ranges from the order of ice crystal size to km^3 [O’Neel *et al.*, 2010; Amundson *et al.*, 2008; Walter *et al.*, 2012].

Concerning floating ice fronts, entire Antarctic ice shelves covering over 1000 km^2 may collapse in a single event lasting a few days only [Scambos *et al.*, 2003]. At the same time, intact ice shelves can produce icebergs of similar sizes several times a year [Lazzara *et al.*, 1999]. These events occur after decades of pervasive fracture propagation within the floating ice (“rifting”), which eventually isolates the iceberg [Bassis *et al.*, 2005; Joughin and MacAyeal, 2005; Fricker *et al.*, 2005].

Iceberg calving is difficult to monitor. Direct observations are expensive and often provide only estimates on calving volume, whereas satellite images are subject to temporal and spatial resolution constraints. Seismology can potentially monitor the activity of calving fronts remotely and at an unrivaled temporal resolution. For decades, seismological investigations have therefore targeted the challenge to extract information of calving activity from glacial seismograms [Hatherton and Evison, 1962; Gaull *et al.*, 1992; Torcal *et al.*, 1999; Köhler *et al.*, 2015] (Table 1).

The first evidence indicating strong seismic activity possibly near (or within) the terminus of the Jakobshavn Isbræ ice stream, west Greenland, was reported three decades ago: 42 low-frequency (1–2 Hz) icequakes, with durations up to 26 min, were recorded during the June–July 1980 field campaign by Qamar and St. Lawrence [1983]. The mean dominant frequency was $1.3 \pm 0.33 \text{ Hz}$, with an “equivalent” local magnitude M_L of 2.5–3.1 for the largest events. The authors noted that a precise magnitude-scaling relationship could not be established, because the low dominant frequency was “not typical of tectonic earthquakes for which the magnitude scale applies” [Qamar and St. Lawrence, 1983]. Furthermore, they suggested a seasonal variation in low-frequency icequakes based on the observation of a small number of events in February 1980 compared to summer season.

Since the pioneering measurements of Qamar and St. Lawrence [1983], technology has evolved and now allows for reliable and long-term installations of broadband seismometers in remote regions of the Earth near calving termini. These data have provided detailed insights into calving seismicity, which can occupy low and high frequencies, as further explained below. Figure 9 shows an example of a large-scale calving event (23 August 2010, recorded at $\sim 70 \text{ km}$ distance) at Kangerdlugssup Sermerssua, a tidewater glacier in west Greenland’s Ummannaq region [Walter *et al.*, 2013b]. The figure shows the exceptionally broad spectrum

of calving-generated seismicity. A prominent hour-long monochromatic oscillation with periods of several minutes is a manifestation of the “fjord seiche,” consisting of normal oscillation of fjord waters in response to the iceberg detachment [Amundson *et al.*, 2012]. The actual iceberg detachment lasts on the order of minutes, only, and occupies the spectrum at frequencies between 0.01 and 5 Hz. These signals are generated by detaching icebergs impacting the ocean surface [Bartholomaeus *et al.*, 2012], fracturing and avalanching of debris [O’Neel and Pfeffer, 2007; Amundson *et al.*, 2008], shifting of the fjord’s debris cover (i.e., ice mélange [Amundson *et al.*, 2012; Sergeant *et al.*, 2016]) and interaction between the detaching iceberg and the glacier terminus and/or fjord bottom (“glacial earthquakes” [Murray *et al.*, 2015]). The following sections will focus on these mechanisms.

2.4. Glacial Earthquakes (GEQs)

The term “glacial earthquakes” describes long-period (30–150 s) [Ekström *et al.*, 2003] surface wave transients (Figure 10) generated during large-scale calving events in Greenland and Antarctic [Amundson *et al.*, 2008; Nettles and Ekström, 2010; Chen *et al.*, 2011]. With equivalent moment magnitudes of $M_{550}=4.6\text{--}5.1$, glacial earthquakes can be detected thousands of kilometers away from calving fronts. At such distances, glacial earthquakes lack the high-frequency content of tectonic earthquakes of comparable sizes, which explains why they went undetected for decades [Ekström *et al.*, 2003].

In contrast to shear dislocation mechanisms of tectonic earthquakes, glacial earthquakes have single force sources [Ekström *et al.*, 2003; Tsai *et al.*, 2008; Larmat *et al.*, 2008]. This led to the initial interpretation that glacial earthquakes are large-scale sliding events of glacier termini [Ekström *et al.*, 2003; Tsai and Ekström, 2007]. Consequently, glacial earthquakes were modeled with the centroid single force model, which describes the time history of single force pairs generated during release and frictional slow down of a mass, such as a landslide [Kawakatsu, 1989]. However, full-waveform inversions of glacial earthquake seismograms determined a product of sliding mass and sliding distance, which had not been observed on glaciers over the tens of seconds duration of glacial earthquake sources [Ekström *et al.*, 2003; Tsai *et al.*, 2008].

Direct observations of calving events at Jakobshavn Isbræ unequivocally identified the detachment of large-scale icebergs as the source of glacial earthquakes [Amundson *et al.*, 2008; Joughin *et al.*, 2008; Walter *et al.*, 2012]. Capsizing of icebergs seems to be a necessary condition for a calving event to generate glacial earthquakes [Nettles and Ekström, 2010]; however, the details of the seismogenic source processes are still debated. A recent study argues that single forces are generated as capsizing icebergs create a low-pressure zone in the water volume adjacent to the ice cliff, which temporarily reduces the load on the bedrock [Murray *et al.*, 2015]. This view is supported by laboratory experiments with polyethylene blocks representing icebergs and by measurements of the terminus’ elastic compression at the beginning of iceberg capsizing. This causes a temporary flow reversal of the terminus and together with the water pressure-induced downward deflection determines the single force mechanism of glacial earthquakes. To date, alternative source mechanisms cannot be fully ruled out, because inversion of low-frequency surface waves is insensitive to details of the source time function of the single force mechanism [Tsai *et al.*, 2008; Walter *et al.*, 2012]. Secondary effects of calving events, such as the reaction of the fjord’s ice debris cover (mélange) also contribute to the single force [Sergeant *et al.*, 2016].

2.4.1. Calving Monitoring With Glacial Earthquakes

Detection and waveform modeling of glacial earthquakes offer the possibility to remotely monitor dynamic changes at large tidewater glaciers. This is likely the reason why glacial earthquakes and calving are the most intensely discussed and researched aspect of glacier seismology (Table 1). Seasonal and temporal variations of GEQ in Greenland were reported by Ekström *et al.* [2006]. A general increase in GEQs with time was demonstrated through 2006, but apparently has remained flat since then, due to differences in temporal activity patterns in western and eastern Greenland [Nettles and Ekström, 2010; Veitch and Nettles, 2012]. However, an anticipated northward migration of GEQ activity [Dahl-Jensen *et al.*, 2010] was confirmed with a recent analysis by Veitch and Nettles [2012] who found that glacial earthquake locations expanded into northwestern Greenland. Here remote sensing studies have shown widespread accelerated ice thinning [Pritchard *et al.*, 2009] in response to warming ocean water penetrating into Greenland’s fjords and compromising terminus and mélange stability [Holland *et al.*, 2008; Murray *et al.*, 2010].

Whereas the number of glacial earthquakes per unit time agrees well with the dynamic behavior of Greenland’s outlet glaciers, it is difficult to arrive at more quantitative information. Surface wave magnitudes range only between $M_{550}=4.6$ and 5.1 [Nettles and Ekström, 2010]. Smaller events may evade detection,

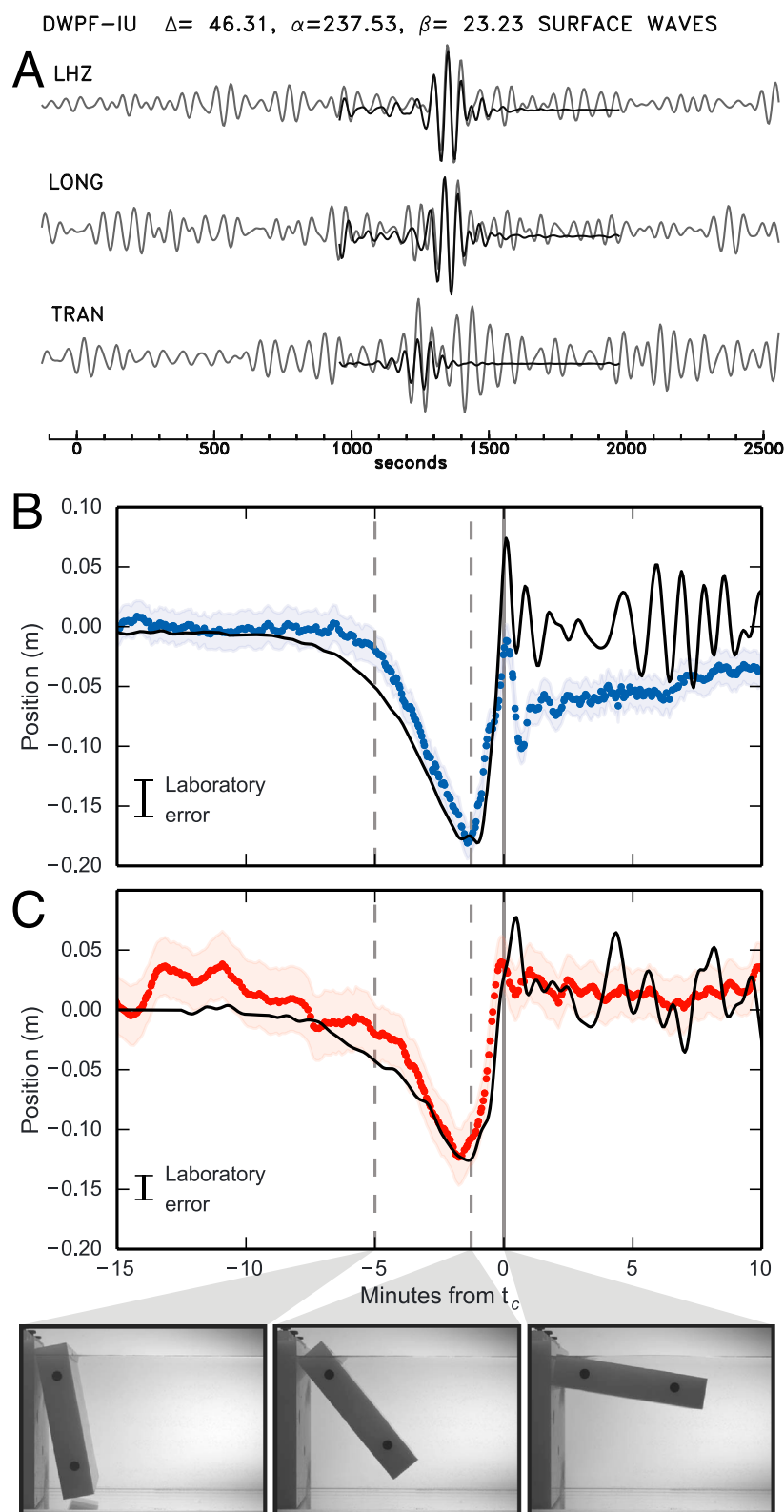


Figure 10. Glacial earthquake signals and source model: (a) Full-waveform fits of surface waves emitted by a Greenland calving event recorded in Florida, USA. Modified from Tsai and Ekström [2007]. (b) horizontal and (c) vertical displacements of the calving terminus in response to a large-scale calving event (colored lines) and a scaled laboratory test (black lines). Bottom panel also shows iceberg detachment and subsequent rotation in the laboratory. Reprinted from Science with permission of the American Association for the Advancement of Science [Murray et al., 2015].

but the lack of larger events to be expected in analogy to tectonic earthquake size scaling suggests that $M_{550}=5.1$ constitutes an upper bound. This may reflect the general scale of those Greenland fjords, which host large-scale calving events. On the other hand, single force magnitudes calculated with full-waveform inversions in the 10–20 s range were found to be a poor representation of calving iceberg volume [Walter *et al.*, 2012]. This makes calving volume estimations based on glacial earthquake signals an ongoing challenge [Sergeant *et al.*, 2016].

2.5. Calving Seiches

The potential of wind to excite normal modes (“seiches”) in lakes and harbors [Miles, 1974] has long been known. However, the advent of broadband seismometers opened up new possibilities of seiche detection via long-period tilt signals [McNamara *et al.*, 2011; Wielandt and Forbriger, 1999]. Such long-period (>100 s) tilt signals result from changing load on the Earth’s crust as the water sloshes back and forth. Residing primarily on the horizontal seismogram components [Wielandt and Forbriger, 1999], the seiche signal period is a direct measure of normal mode periods of the resonating water body.

Calving events disturb the water level in a fjord, generating seiches [Sergeant *et al.*, 2016] as well as tsunamis [Amundson *et al.*, 2012; Lüthi and Vieli, 2015]; see video at <https://youtu.be/U3F6kv3To3Y>. The seismic signature of this loading has been used to detect calving events [Amundson *et al.*, 2012; Bartholomaus *et al.*, 2012]. Similar to glacial earthquake monitoring, automatic calving seiche detection is, in principle, possible [Walter *et al.*, 2013b]. However, changes in terminus position and mélange cover influence fjord resonance frequencies and thus complicate calving volume estimations based on seiche signals [Walter *et al.*, 2013b; MacAyeal *et al.*, 2012].

2.6. High-Frequency Calving Seismicity

In contrast to glacial earthquake signals, high-frequency seismicity (>1 Hz) is too complex to allow for full-waveform modeling. During ongoing iceberg detachment, fracturing, slipping, and avalanching of ice debris [O’Neel and Pfeffer, 2007; Amundson *et al.*, 2008; Köhler *et al.*, 2012] as well as interaction between iceberg and water surface [Bartholomaus *et al.*, 2012] produce a high-frequency “chaos” of seismic records. Maximum seismic energy in the 1–5 Hz band is a prominent feature of high-frequency calving seismicity and a manifestation of detaching icebergs’ impact on the water surface and subsequent deceleration and air cavitation within the water [Bartholomaus *et al.*, 2012] (Figure 11). Alternative explanations of the 1–5 Hz calving band seismicity are englacial water resonances [O’Neel and Pfeffer, 2007; Walter *et al.*, 2010a].

High-frequency calving seismicity is prominent even on regional seismic networks at source station distances of tens or hundreds of kilometers [O’Neel *et al.*, 2010; Köhler *et al.*, 2015] and thus well suited for monitoring purposes. In contrast to tectonic earthquakes, the frequency content of long- and short-duration calving icequakes (7 min versus 20 s) are independent of discharged volume [O’Neel *et al.*, 2007]. Similarly, these signals showed no clear relationship between seismic amplitudes and discharge mass [O’Neel *et al.*, 2007; Walter *et al.*, 2012].

In contrast, Qamar [1988] found that the duration T of calving seismic signals at Columbia glacier (Alaska, USA) was a promising indicator of volumetric change V . He proposed a relationship between duration T (in s) and calved volume, for seismic signals recorded within 10 km of the calving terminus. This relationship is strictly empirical, because signal attenuation as well as site-related signal-to-noise ratio and local amplification effects do not allow application to other calving fronts and different source station distances. A similar relationship between T and V was later confirmed by observations on the same glacier by O’Neel *et al.* [2007], on Yaktse Glacier (Alaska, USA) by Bartholomaus *et al.* [2015a] and on the Arctic archipelago of Svalbard (Norway) (A. Köhler and C. Nuth, personal communication, 2016).

Seismic monitoring revealed that calving mass loss depends on ocean tides and terminus geometry [O’Neel *et al.*, 2007; Walter *et al.*, 2010a; Bartholomaus *et al.*, 2015a; Koubova, 2015]. Recently, Köhler *et al.* [2015] analyzed 14 year period (2000–2013) of seismicity on Spitsbergen and observed a high number of icequakes (1–8 Hz) associated with tidewater calving glaciers. A higher number of events during the ablation season, often delayed by 1 to 2 months from the melt peak, suggested a control on calving by sea surface temperatures (presumably, through frontal ablation, which is undercutting the glacier terminus).

Unfortunately, monitoring of high-frequency seismicity has not matured toward a global, stand-alone tool for calving studies. Detection algorithms require site-specific tuning [O’Neel *et al.*, 2007; Bartholomaus *et al.*, 2015a]. Iceberg disintegration [Richardson *et al.*, 2010, 2012] and collision with obstacles [MacAyeal *et al.*, 2008;

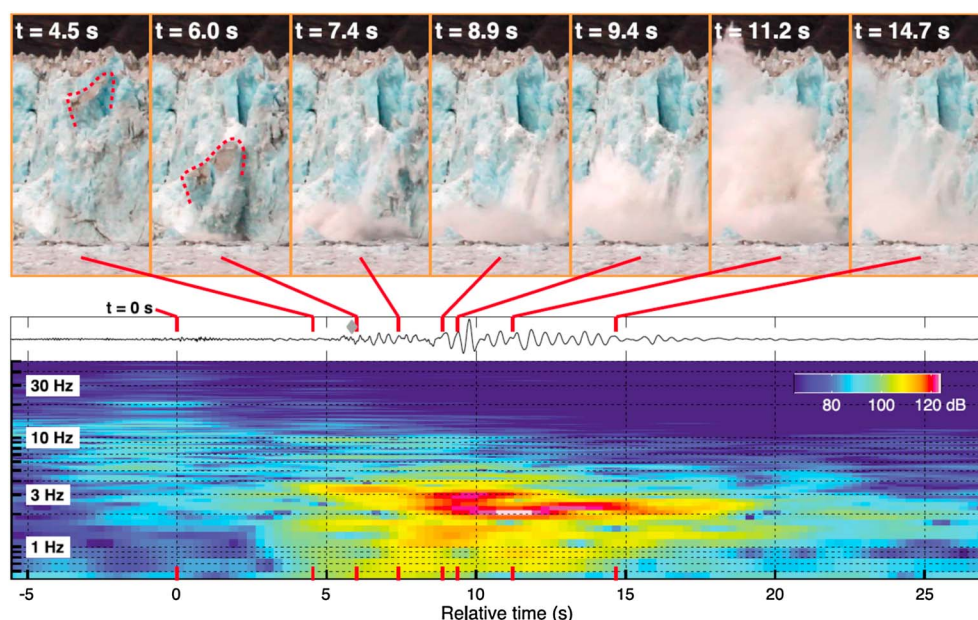


Figure 11. Video stills of calving event at the terminus of Yaktse Glacier, unfiltered seismic waveform and spectrogram of waveform. Terminal cliff is approximately 60 m tall. In the first two panels from video, the top of the major detached block is outlined with a red, dashed line. Ice associated with the calving event is observed to begin falling at 7 September 2010, 22:13:50.3 UTC ($t=0$). The time of each video panel is identified in seconds relative to $t=0$ and marked on the seismic data by vertical red ticks at the top of the waveform and bottom of the spectrogram. Seismic data have been shifted forward 0.95 s to correct for seismic wave travel time. A “step” in the icequake amplitude is identified with a gray diamond. At 8.8 s, a Worthington jet emerges from the fjord. The spectrogram presents the velocity of the sensor (in dB) as a function of frequencies between 0.5 and 50 Hz, as a function of time. Reprinted from the *Journal of Geophysical Research: Earth Surface* with permission of John Wiley and Sons [Bartholomaus *et al.*, 2012].

[Dziak *et al.*, 2013] produce signals within a seismic spectrum, which includes 1–5 Hz calving band. Furthermore, the generation of seismic signals within the proglacial water is highly sensitive to the iceberg’s free-fall height [Bartholomaus *et al.*, 2012]. The state of the ocean surface which can be covered by sea ice, ice mélange, or be ice-free depending on the season [e.g., Sergeant *et al.*, 2016] adds another degree of complexity. Improvement of calving monitoring with the help of high-frequency seismicity could therefore come from more sophisticated signal classification approaches [Köhler *et al.*, 2012; Hammer *et al.*, 2015] and techniques for source location suitable for emergent signals [Richardson *et al.*, 2010; Koubova, 2015; Mei *et al.*, 2016].

2.6.1. Ice Shelf Calving

Compared to grounded tidewater termini, seismic studies of calving on floating ice shelves are sparse. One reason may be that ice shelves lie in high-latitude polar regions (in particular Antarctic), where access is difficult. On the Amery Ice Shelf (AIS), a combination of GPS and local seismic networks (sizes on kilometer scale) revealed seismogenic rift propagation episodes, which occur once every several weeks and last for a few hours [Bassis *et al.*, 2005, 2007, 2008; Heeszel *et al.*, 2014a]. The rift propagation velocity seems unrelated to external forcing such as wind and tides but instead depends on material properties of the ice surrounding the rift tip [Bassis *et al.*, 2007, 2008].

Rift propagation through a floating ice shelf eventually produces large tabular icebergs, and times between major calving events may reach decades or even centuries [Lazzara *et al.*, 1999]. Such tabular icebergs do not capsize or disintegrate upon detachment, which contrasts with the nearly continuous calving style on grounded termini. This may be a reason why glacial earthquake detection in Antarctic is relatively sparse [Nettles and Ekström, 2010; Chen *et al.*, 2011]. An alternative explanation is that the transmission of seismic waves between a floating ice shelf and the Earth’s crust is less efficient than for grounded termini.

Seismic observations at Columbia Glacier, Alaska, provide important insights into these issues. Untypical for nonpolar temperate termini, the glacier temporarily had a floating calving front during its rapid retreat, which initiated in the 1980s. The flotation episode coincided with a change in calving style from nearly continuous

but small events to rift formation and episodic release of large icebergs. At the same time, seismic emission in the 1–5 Hz calving band decreased [Walter *et al.*, 2010a]. Whether this decrease is a result of changes in seismic transmission properties [Bartholomäus *et al.*, 2015b] or englacial and subglacial water pressures [Walter *et al.*, 2010a] is subject to debate. The answer would also explain why Columbia Glacier produced seismic signals detectable at regional distances (tens to hundreds of kilometers) when its floating calving front retreated toward the grounding line (S. O'Neel, personal communication, 2014).

2.7. Dry Calving

Catastrophic break-off events of steep glacier tongues or high-altitude glaciers frozen to their beds ("hanging glaciers") can result in devastating gravitational flows, with volumes up to several million cubic meters and run-out distances up to 30 km (e.g., Kolka and Huascarán) [Cuffey and Paterson, 2010; Faillettaz *et al.*, 2015]. Many mountain communities are threatened by these rare but highly destructive "dry" calving events. Experiments by Weaver and Malone [1979] and Gaull *et al.* [1992] indicate that seismic monitoring is helpful in predicting ice falls, due to observations of higher seismicity just before ice avalanches. For example, Weaver and Malone [1979] showed a tenfold increase in the number of seismic events before an ice avalanche that involved the entire glacier's thickness.

More recently, Faillettaz *et al.* [2008] reported an order-of-magnitude increase in seismic activity before catastrophic ice rupture of a hanging glacier; at the same time, they noted a decrease in low-amplitude icequakes and an increase in stronger events. Faillettaz *et al.* [2015] studied the waiting time between successive icequakes and proposed a mechanism explaining the maturation of rupture (see the original paper and its Figure 6 for details).

Current climate warming is known to affect the temperature regime of hanging glaciers and consequently their stability [Faillettaz *et al.*, 2011; Preiswerk *et al.*, 2016]. Whether resulting from evolving englacial damage, initiation of sliding and/or enhanced surface crevasse formation, seismic monitoring during such temperature regime transitions may provide valuable information for forecasting efforts. This is particularly useful if seismometer pairs are used to allow for codetection of icequake signals [Faillettaz *et al.*, 2016]. Furthermore, icequake activity induced by teleseismic earthquakes [Peng *et al.*, 2014] suggests the possibility of dynamic triggering of englacial fracturing. Consequently, future studies should focus on the question to which degree the passage of earthquake shaking can influence the stability of hanging glaciers and steep glacier tongues.

3. Seismicity Related to Glacier Hydraulics

Hydraulic conditions at the ice bed control basal motion and thus have a profound impact on ice sheet and glacier flow. High water pressures tend to reduce basal resistance and thus increase sliding velocities, as described in section 2.2. Similarly, high subglacial water pressures can diffuse into pore space and thus weaken or stiffen subglacial sediments [Walter *et al.*, 2014; Bougamont *et al.*, 2014]. However, subglacial water pressures in turn depend on the configuration of the subglacial drainage system, which may change on seasonal scales with efficient channels operating under low pressures, in contrast to inefficient, pressurized linked cavity systems (for reviews, see Clarke [2005] and Irvine-Fynn *et al.* [2011]). Arguably, it is this sensitivity of subglacial water pressures on the type of subglacial drainage that makes basal motion one of the largest uncertainties in ice sheet models [Ritz *et al.*, 2015].

Measurements of subglacial water flow and hydraulic conditions within and underneath the glacier are inherently difficult and expensive. Hot water drilling and subsequent borehole instrumentation with pressure sensors, cameras, inclinometers, and temperature gauges provide important insights but only in the form of point measurements [Iken *et al.*, 1993; Harper *et al.*, 2005]. On the other hand, radio echo sounding and active source seismology can target a wide region and resolve drainage channel development within the ice or hydraulic changes within the bed [Nolan and Echelmeyer, 1999; Fountain *et al.*, 2005]. However, these measurements are difficult to repeat regularly over an extended time period and therefore provide distinct snapshots in time only.

In contrast, passive seismic techniques can target a region of the glacier or ice sheet, which is constrained only by the spatial extent of the monitoring network, noise level of background seismicity, and/or the strength of targeted seismic signals. Furthermore, on-ice seismometer installation techniques have undergone rapid improvements in recent years and multiseasonal deployments are now possible even in high-melt regions [Dalban Canassy *et al.*, 2016]. In the following, we discuss the seismic signature of the presence, pressure,

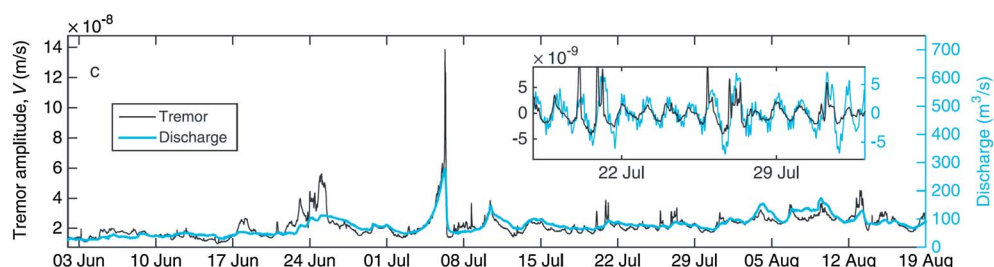


Figure 12. Glaciohydraulic tremor amplitude V and water discharge into Mendenhall Lake, Alaska. Inset shows V and discharge band-pass filtered for periods between 2 h and 2 days. Reprinted from the *Geophysical Research Letters* with permission of John Wiley and Sons [Bartholomaus et al., 2015b].

and movement of englacial and subglacial water and how this can provide important insights into hydraulic processes.

3.1. Hydrofracturing

The depth to which surface crevasses penetrate depends on tensile stresses, ice fracture toughness, glacier thickness, and water level within the crevasses [Weertman, 1973; Van der Veen, 1998]. Typical penetration depths are tens of meters in the absence of water, but if sufficient meltwater is available, hydrofracturing may extend crevasses to the glacier bed [Van der Veen, 1998, 2007]. The presence of crevasse icequakes at “intermediate depths” (that is below the surface crevasse zone) is therefore evidence for englacial water.

Various studies in temperate Alpine glaciers have presented evidence for icequakes at intermediate depth (~ 100 m) [Deichmann et al., 1979, 2000; Walter et al., 2009; Helmstetter et al., 2015b]. Full-waveform inversions argue for coseismic positive volumetric changes on the order of ~ 100 cm³, which is compatible with a tensile mechanism expected for hydrofracturing. Accordingly, the activity of intermediate icequakes in glacier ablation zones depends on the availability of meltwater. A study of 20–130 m deep icequakes on Glacier d’Argenti re (France) showed sporadic bursts in the activity of intermediate icequake clusters [Helmstetter et al., 2015b]. The waveforms of these icequakes are also consistent with tensile faulting, and variations in depths of icequakes belonging to a single cluster coincided with the warmest periods. This suggests that the gradual increase in icequake occurrence depth is related to increased water levels in cracks. The intermediate icequakes on Glacier d’Argenti re show a power law for the relation between peak amplitude and interevent time. However, the average rate of activity after a particular event did not depend on its amplitude, indicating that bursts of activity should be related to external forcing, such as oscillations in water pressure.

Current climate warming raises the question how hydrofracturing will affect cold polar regions, which so far are prone to little or no surface melt. Seismic measurements on the dry polar Taylor Glacier, Antarctica, revealed that even small meltwater input promotes hydrologically triggered fractures [Carmichael et al., 2012]. On a larger scale, collapses of entire Antarctic ice shelves can result from meltwater-driven crevasse propagation [Scambos et al., 2000]. On-ice seismic monitoring may therefore elucidate the vulnerability of the cryosphere to increases in meltwater production by recording detailed dynamics of hydrofracturing.

3.2. Water Flow: Seismic Tremor

Seismic networks serve as a monitoring tool for fluvial processes [Burtin et al., 2008, 2011], because bedload transport [Tsai et al., 2012] and water turbulence [Gimbert et al., 2014, 2016] within torrents and rivers emit seismic energy at frequencies between 1 and 20 Hz. These seismogenic processes are also active in subglacial [Winberry et al., 2009b] and englacial water flow, and the “glaciohydraulic tremor amplitude” (amplitude of continuous seismic noise in the 1.5–10 Hz range) can thus serve as a proxy for subglacial water discharge [Bartholomaus et al., 2015b] (Figure 12). This is particularly useful for tidewater-terminating glaciers, where subglacial discharge drives fjord water circulation, which enhances submarine melt and thus a large or even dominant part of the frontal ablation [Motyka et al., 2003, 2011, 2013; Rignot et al., 2010]. In such environments, seismology constitutes an unrivaled tool for subglacial discharge monitoring.

Although the correlation between glaciohydraulic tremor amplitude and subglacial discharge was only recently documented [Bartholomaus et al., 2015b], past studies already provided some evidence for such a relation. M taxian [2003] and R  sli et al. [2014] showed that seismic noise at a glacier is highest during the warmest part of the day, due to maximum melt runoff. This melt noise decreases the sensitivity of trigger

algorithms for seismic event detection leading to a spurious lower number of icequake detections during periods of high drainage [Walter *et al.*, 2008; Dalban Canassy *et al.*, 2012, 2013; Rössli *et al.*, 2014]. In order to tackle this biasing problem, event declaration schemes can be designed to compensate for changes in detectability. This has been implemented by assigning a score to each detection based on how far the event is observable throughout the monitoring network [Carmichael *et al.*, 2012, 2015].

3.2.1. Water Resonances

The seismic signature of englacial and subglacial drainage may contain resonances within the void spaces hosting the water flow. These resonances have attracted considerable attention, because they constrain the geometry of subsurface drainage networks, about which little information is known from alternative measurements.

Considering guided waves in a thin fracture, Lipovsky and Dunham [2015b] present a method to infer hydraulic fracture dimensions using seismic resonance peaks. For seismic observations, the most relevant asymptotic relation describes the “boundary layer limit,” in which most fluid displacement during vibration excitation occurs within a viscous boundary layer near the fracture walls. This asymptotic relation is given in Table A2; for more complete solutions, including fully developed flow, the reader is referred to the graphical method presented in Figure 7 of Lipovsky and Dunham [2015b]. The authors furthermore suggest the following useful diagnostic: resonances from fluid-filled fractures have noninteger spacing between resonant frequencies ($f_n/f_1 = n^{1.5}$), contrary to nondispersive waves with overtones (f_n) appearing at integer multiples (n) of the fundamental mode (f_1).

As an alternative to the crack wave model of Lipovsky and Dunham [2015b], Roeoesli *et al.* [2016b] use a one-side-closed organ pipe model for an englacial vertical drainage shaft to explain resonance peaks in hour-long tremor signals [Rössli *et al.*, 2014]. These “moulin” tremors (2–8 Hz) constitute the dominant ambient noise on the Greenland ice sheet and their resonance frequencies are determined by the water column within the drainage shaft [Roeoesli *et al.*, 2016b; Walter *et al.*, 2015a].

The idea of water resonances in glaciers was proposed as an analogue for hydraulic displacement of lava tubes in volcanic areas several decades ago by St. Lawrence and Qamar [1979]. In the authors’ view, changing water pressures within englacial conduits displace the conduit walls causing nearly monochromatic wave trains recorded near Alaskan glaciers. The triggering water pressure change may be caused by nearby fracturing, and thus also explains near-monochromatic coda, which is sometimes observed after deep icequakes [West *et al.*, 2010; Rössli *et al.*, 2014; Helmstetter *et al.*, 2015b]. This phenomenon is again analogous to “hybrid events” in volcanic environments [White *et al.*, 1998; Neuberg *et al.*, 2000; Harrington and Brodsky, 2007].

In general, changes in englacial water channels are particularly interesting, because they may provoke a change in overall glacier flow. So far, only few observational studies of water resonances have focused on this point: Heeszel *et al.* [2014b] found changes in hour-long tremor resonance during the englacial and subglacial drainage of an ice-marginal lake. Similarly, they found that resonance frequencies of basal icequake signals changed over the course of a day. Both observations were explained with changes in resonating fracture volume. Similarly, Helmstetter *et al.* [2015b] argued that a decrease in icequake resonance frequency may be an indicator of crack tip propagation.

New installation techniques allowing for multiseasonal, high-density seismic networks, will likely improve our understanding of water-generated seismic signals. Changes in seismic signature of water flow reflecting changes in the configuration of the subglacial drainage system would be particularly useful in the study of hydraulic controls on glacier flow.

3.3. Glacier Lake Drainages

The possibility to detect englacial water flow and hydrofracturing with seismic methods opens up new perspectives for studies of glacier-dammed lake outbursts. These floods are also referred to by the Icelandic term jökulhlaup and affect practically all glaciated regions worldwide. The draining water bodies can locate at the ice surface, the bed, within the ice column, and in between glacial ice and bedrock or sediments. Water discharges vary depending on lake and glacier geometries but commonly exceed tens of $\text{m}^3 \text{s}^{-1}$ and may have been as high as $20 \times 10^6 \text{ m}^3 \text{s}^{-1}$ for Pleistocene floods (for a review, see Roberts [2005]).

Glacier outburst floods are difficult to predict and thus constitute a considerable natural hazard [Kääb *et al.*, 2005]. A main reason is that ice dams can breach suddenly due to hydrofracturing and hydrojacking, and water flow quickly enlarges ice-walled channels due to heat dissipation at the channel walls [Roberts *et al.*, 2000;

Nye, 1976; Spring and Hutter, 1982]. The role of hydrofracturing is difficult to assess with conventional glaciological methods. Hydrofracturing is often responsible for drainages of Greenland's supraglacial lakes [Tsai and Rice, 2010; Doyle et al., 2013; Carmichael et al., 2015]. During such events, lakes with surface areas of several square kilometers, completely drain within a few hours via discharges of $8700 \text{ m}^3 \text{ s}^{-1}$, which route the water directly to the ice sheet bed [Das et al., 2008].

Precursory seismicity can, in principle, be detected with seismometers [Das et al., 2008; Jones et al., 2013]. However, focusing seismic source studies on hydrofracturing is complicated by the background seismicity associated with surface crevasses, which can also increase or shift as draining lakes perturb glacier flow [Roux et al., 2008; Carmichael et al., 2015].

Englacial and subglacial water drainage during glacier outburst floods raise the level of seismic background noise [Russell et al., 2010; Winberry et al., 2009b; Morgan et al., 2013]. These tremor signals may constitute a valuable signal for remote drainage monitoring and early warning systems, since flow rates of $1000 \text{ m}^3 \text{ s}^{-1}$ can emit seismic energy comparable to an $M_{2.5}$ earthquake [St. Lawrence and Qamar, 1979]. In addition, gliding spectral peaks in signals of lake drainage tremors may constrain geometries and discharge in water channels within and under the glacier [Heeszel et al., 2014a]. So far, no alternative to this seismic approach exists, which could provide information about water channel evolution needed to test existing mathematical descriptions of water flow through its solid phase [Nye, 1976; Spring and Hutter, 1982].

4. Glacier-Covered Volcanoes and LP Events

The main interest in cryoseismic sources on glaciated volcanoes was initiated by the discovery that glaciers can interfere with volcanic signals [Weaver and Malone, 1976]. The reason for this confusion is because typical precursory volcanogenic signals have frequencies (0.5–5 Hz) and magnitudes ($M \leq 2.5$) that fall into the “busiest” domain of glacier-generated seismic signals (as will be shown in detail in section 6.2). With poorly constrained hypocenters, interpretation of such events is challenging. Some enigmatic low-period glacial events (hereafter “LP”) could be easily confused with volcanic activity [St. Lawrence and Qamar, 1979]. This is particularly dangerous because LP seismic events are often interpreted as eruptive precursors; hence, misidentification could lead to false alarms and unnecessary evacuations [Jónsdóttir et al., 2009]. Thus, it is of crucial importance to accurately distinguish between these event types, which have very different origins but comparable seismic signatures.

LP earthquakes of glacial origin can be caused by ice falls [i.e., a single force mechanism; Jónsdóttir et al., 2009], resonant water-filled ice cavities, or sudden changes in water flow rate [St. Lawrence and Qamar, 1979; Métaixian, 2003].

Stress drop thresholds could be used to distinguish icequakes from volcanic events. Due to the relatively weak strength of ice, it is reasonable to expect a smaller stress drop, $\Delta\sigma$ (see section 6.2 and Caplan-Auerbach et al. [2004]). Moreover, in theory, diagnostic imaging of fluid-filled fracture/conduit dimensions, obtained from quality factor and characteristic frequency based on the method introduced by Lipovsky and Dunham [2015b], may provide additional information due to significant differences in the material properties of solids and fluids (i.e., rock/ice, magma/water).

To avoid false alarms, the suggested intensification of cryoseismicity due to recent intense melting of ice bodies [Jónsdóttir et al., 2009] may require more attention from eruption forecasters. One practical solution for better recognition of glacial signals at volcanoes is to locate sensors as close to glaciers as possible [Métaixian, 2003].

In another context, Chaput et al. [2015] examined thousands of icequakes at Erebus volcano, Antarctica, to analyze scattering in its upper structure. They recover Green's functions, and suggest that coda correlation opens up new possibilities for seismological imaging of glaciated volcanoes.

5. Sea/Lake Ice and Icebergs

5.1. Sea Ice

By coupling into the solid earth, ocean waves produce surface wave signals, which can be detected far in the interior of continents [Hasselmann, 1963]. There are two types of such signals in the 1–20 s period band, and they are called microseisms. It is known that sea ice has an impedance effect on both types of microseisms

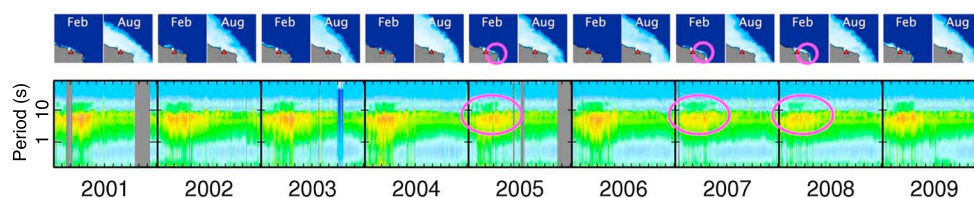


Figure 13. Seismic noise spectrum variability at Syowa station, Antarctica, from 2001 to 2009. Also shown are ice concentration images for February and August of each year from the NSIDC Sea Ice Index. The approximate location of SYO is indicated by a red triangle. Summer microseisms are of lower amplitude during 2005, 2007, and 2008, corresponding to residual ice near the Enderby Land coast (highlighted in magenta). Adopted from the *Geophysical Research Letters* with permission of John Wiley and Sons [Grob *et al.*, 2011].

[Hatherton, 1960] and that local sea ice variability may be indirectly derived from power spectral densities of broadband data [Grob *et al.*, 2011]. For instance, Grob *et al.* [2011] analyzed a decade of continuous seismic noise spectra from coastal stations in Antarctica, and demonstrated that systematic seasonality in microseismic power is associated with sea ice conditions (Figure 13). Moreover, higher amplitudes of microseisms during the local summer season, with less sea ice coverage, correspond to lower teleseismic detectability [Kanao *et al.*, 2012a; Storchak *et al.*, 2015]. Therefore, it can be expected that the increased duration of open water in the Arctic will influence the performance of coastal stations. Recent increase of microseisms' activity in Antarctic was interpreted as a consequence of climate change [Aster *et al.*, 2009].

In coastal East Antarctica, near Showa station (Queen Maud Land), it was observed that sea ice discharge from a bay can produce harmonic tremor lasting up to 6 h (30 July 1997) [Kanao *et al.*, 2012a]. Tens of such tremors (1–10 Hz) with durations between 15 min and 9 h can be detected monthly [Kanao, 2015]. It was also observed that tides can induce cracks in sea ice, as indicated by increased numbers of icequakes during falling tides [Kaminuma, 1994].

Sea ice seismology is of interest not only for glaciogenic studies (e.g., Marsan *et al.* [2012] derived sea ice thickness from dispersed ocean swell), but because sea ice is an alternative for sensor location when solid-rock installation sites are not available. For example, seismometers were placed on sea ice floes in the Arctic by Läderach and Schlindwein [2011], allowing location of regional earthquakes.

5.2. Lake Ice

The equivalent magnitudes of seismic events at lakes are only $M_w = 0.2 \sim 0.8$ [Ruzhich *et al.*, 2009; Carmichael *et al.*, 2012]. Nevertheless, it was recently shown that thermal cracking of lake ice can induce strong vibrations in coastal high-rise buildings [Makkonen *et al.*, 2010]. This makes it a rare type of cryospheric seismicity detectable by people and can even cause panic. Some land-terminating glaciers may generate low-magnitude events due to thermal cracking of ice on marginal lakes nearby [Carmichael *et al.*, 2012]. Interestingly, multiyear passive seismic observations of the mechanical behavior of Lake Baikal ice provided a basis for treating it as a model for large-scale lithospheric processes [Ruzhich *et al.*, 2009].

5.3. Large Tabular Icebergs

As drifting icebergs impact and scratch shallow-ocean bottoms, they may cause high-magnitude events with long-lasting signals detectable at teleseismic distances. This phenomenon is beautifully referred to as “singing icebergs” [Müller *et al.*, 2005]. Recently, Pirli *et al.* [2015] demonstrated the potential of regional seismic networks for iceberg monitoring in the Antarctic.

Icebergs act as low-pass filters of oceanic waves for seismometers placed on top of them [Kristensen *et al.*, 1982]. Therefore, rare (and thus unique) seismic experiments, with instruments installed on drifting icebergs, have allowed researchers to identify teleconnections with distant storms and tsunamis, as well as dispersive microtsunamis produced by detached icebergs [Okal and MacAyeal, 2006; MacAyeal *et al.*, 2006, 2009].

Moreover, colliding icebergs are known to generate up to thousands of short stick-slip events daily, which can merge into long tremor episodes with distinct aseismic “eyes” or gaps [MacAyeal *et al.*, 2008] (Figure 7b). Seasonal cycles of hydroacoustic noise produced by icebergs were identified recently as one of the most dominant ocean sound sources in the Southern Hemisphere (see a reference in MacAyeal *et al.* [2015]).

Ambient-noise-correlation analysis of a few broadband seismometers located on the same iceberg has identified an asymmetric noise correlation function, meaning the noise field is strongly directional and mainly

originates from the iceberg collision zone [MacAyeal *et al.*, 2015]. The nonisotropic nature of the noise prohibits ambient noise tomography [Zhan *et al.*, 2014]; however, this still provides an opportunity to distinguish phases in wave propagation, as demonstrated by MacAyeal *et al.* [2015]. The approximately diurnal rhythm of the noise was suggested to be produced by diurnal tides, which are strong drivers of iceberg motion [MacAyeal *et al.*, 2015]. For multiple phase speeds identified by the latter authors (P , S , hydroacoustic, and flexural gravity waves), an interesting observation has been made regarding the relationship between long-wavelength cutoff and water thickness below an iceberg. Because water is a layer of minimum velocity, it acts as an efficient waveguide and thus behaves like a high-pass filter. The filter's cutoff frequency is given by $f_c \approx V_{p,\text{water}}/d$, where f_c is the lowest passing frequency and d is the water column thickness [MacAyeal *et al.*, 2015]. This suggests the possibility of a combined seismic/hydroacoustic mode, with more seismic energy at low frequencies and more hydroacoustic energy at high frequencies.

6. Discussion

6.1. Seismic Portrait of the Cryosphere

The wide variety of cryospheric seismic sources covers a broad range of frequencies (from 10^{-3} to 10^2 Hz; i.e., ~ 17 octaves), magnitudes ($-3.2 \leq M \leq 5.1$, $\max(M) = 7.0$), and physical processes (Figure 14). Two broad signal categories can be identified: (1) teleseismic (epicentral distance $\Delta > 30^\circ$) and (2) regional/local. The former can be detected through global seismic networks with source station distances of thousands of kilometers (e.g., the Global Seismographic Network, GSN) and are caused by glacial earthquakes associated with large-scale glacier calving, large-scale stick-slip motion of the Whillans Ice Stream, and iceberg collisions with the ocean floor. The latter category may be detected only with regional/local arrays with source station distances of tens to hundreds of kilometers, due to relatively weak, highly attenuated signals originating from smaller events associated with ice and snow fractures, ice falls, icebergs, basal motion, glacier surge, calving, and ice-water interactions.

From the perspective of risk mitigation, the seismic emissions of the largest cryoseismic sources associated with GEQs in Greenland and the large-scale Whillans Ice Stream stick-slip events ($M_{5.1}$ and 7, respectively) do not generate any strong ground motion, or shaking, dangerous to human life or infrastructure. This is so because these events release energy at a much slower rate than tectonic earthquakes of comparable magnitude (see Ekström *et al.* [2003] and Larmat *et al.* [2008] for examples). Accordingly, cryospheric seismicity is interesting mainly as an insight into ice dynamics, as a proxy for climate change, or as a factor interfering with the interpretation of volcano-monitoring data [Méteaxian, 2003] and local earthquake seismology [Kanao *et al.*, 2012a].

Nevertheless, ice shocks generated by sea ice pressure are familiar to polar explorers [Nansen, 1897]. Furthermore, large icequakes have been familiar to Greenlanders living in Quanaaq area (T. Oshima, personal communication, 2015) or were felt during field campaigns near the calving front of Bowdoin Glacier (M. Funk, personal communication, 2015).

6.2. Characteristic Frequency and Magnitude

The data summarized in Figure 14 illustrate several key points.

1. First, Figure 14 shows the general trend of decreasing characteristic frequency with magnitude. From seismic observations and source spectra theory of tectonic faults (i.e., shear slip) it is known that the largest earthquakes produce energy at the lowest frequencies, due to the larger rupture volume [Stein and Wysession, 2003]. This relationship can be seen on theoretical source spectra of earthquakes. More specifically, as the seismic moment increases, the so-called corner frequency, f_c , is decreasing. The corner frequency is defined as the intersection between the flat spectrum segment at low frequencies and the declined spectrum segment at high frequencies. This corresponds to the recommended "best practice" of estimating earthquake magnitudes around the corner frequency, where the maximum energy is centered [Bormann, 2012]. The corner frequency can be calculated from a seismic moment and a so-called stress drop (i.e., the difference between acting stress at the source area before and after the rupture, or, in other words, the stress released by the earthquake).

To test how the observed trend of decreasing frequency with magnitude of cryoseismic events relates to this concept, we assume that it is valid for glacial sources. As it was shown in the previous sections, various seismic sources in the cryosphere, such as tensile surface cracks, single force-calving events, and

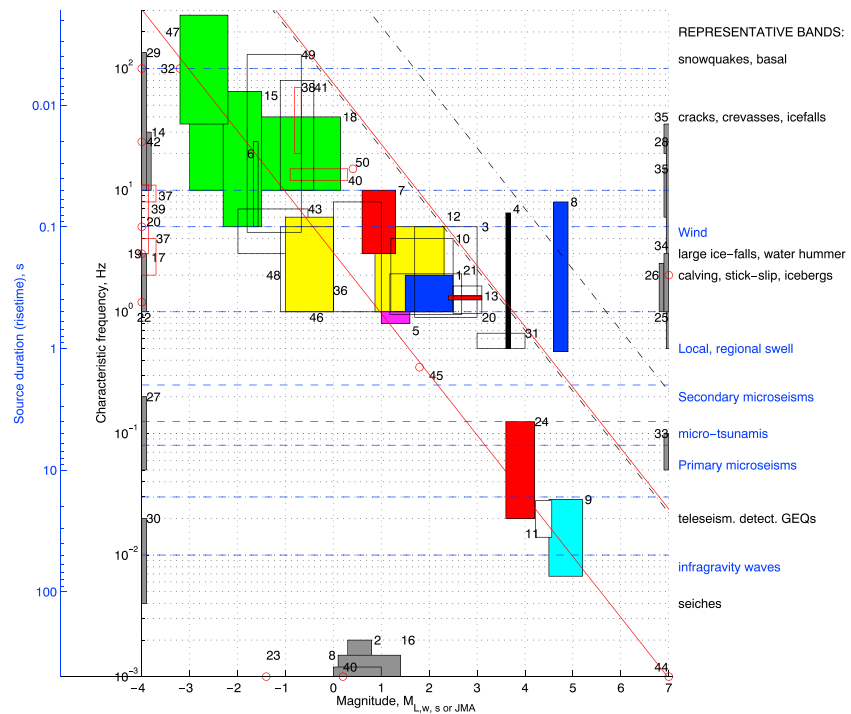


Figure 14. Magnitudes ($M_{L,W,S,JMA}$) and characteristic frequencies of various cryospheric sources. Gray rectangles or red circles, “touching” x or y axes correspond to data where at least one piece of information was missing (frequency or magnitude). In general, values are provided only to illustrate the main ranges or “working” frequencies that yield valuable seismological information for each individual study. Lower and upper red straight lines indicate f_c for ice with $\Delta\sigma = 10^{-4}$ and 1.5 MPa, respectively. Black dashed lines indicate f_c for rock with $\Delta\sigma = 10^{-1}$ and 100 MPa, respectively. Numbers correspond to bibliographic references as follows: 1 – Wolf and Davies [1986], 2 – Ruzhich et al. [2009], 3 – VanWormer and Berg [1973], St. Lawrence and Qamar [1979], 4 – Müller et al. [2005], 5 – Danesi et al. [2007], 6 – Roux et al. [2010], 7 – Sinadinovski et al. [1999], 8 – Rial et al. [2009], 9 – Nettles and Ekström [2010], 10 – Jónsdóttir et al. [2009], 11 – Chen et al. [2011], 12 – O’Neel et al. [2010], 13 – Adams [1969], 14 – Faillettaz et al. [2011], 15 – Walter et al. [2009], 16 – Kanao and Kaminuma [2006], 17 – O’Neel and Pfeffer [2007], 18 – Roux et al. [2008], 19 – Winberry et al. [2009b], 20 – Qamar and St. Lawrence [1983], 21 – Weaver and Malone [1979], 22 – Métaxian [2003], 23 – Anandakrishnan and Alley [1994], 24 – Wiens et al. [2008], 25 – Kanao et al. [2012b], 26 – Zoet et al. [2013a], 27 – MacAyeal et al. [2009], 28 – Martin et al. [2010], 29 – Stuart et al. [2005], 30 – Bromirski et al. [2010], 31 – Hatherton and Evison [1962], 32 – Harrison et al. [1986] and Neave and Savage [1970], 33 – Walter et al. [2012], 34 – Bartholomaeus et al. [2012], 35 – West et al. [2010], 36 – Caplan-Auerbach and Huggel [2007], 37 – Górski [2004], 38 – Walter et al. [2010b], 39 – Rössli et al. [2014], 40 – Carmichael et al. [2012], 41 – Mikesell et al. [2012], 42 – Jones et al. [2013], 43 – Heeszel et al. [2014a], 44 – Winberry et al. [2013], 45 – Zoet et al., 2012 [2012, 2013b], 46 – Thelen et al. [2013], 47 – Helmstetter et al. [2015a], 48 – Allstadt and Malone [2014], 49 – Smith et al. [2015], 50 – Carmichael et al. [2015].

fjord seiches, are distinct from source mechanisms of tectonic earthquakes. Nevertheless, in this section, we will show that stress drops typical for ice and glaciers fit the presented data reasonably well.

2. The above reasoning makes it tempting to predict the corner frequency, f_c . However, assumptions must be made about the source in order to calculate f_c from a seismic moment and stress drop, $\Delta\sigma$. For the sake of simplicity, we consider a circular shear fault as a source mechanism. Assuming this geometry and source model allows us to recall that for large and microearthquakes, seismic moment, M_0 , is proportional to f_c^{-3} . For instance, Allmann and Shearer [2009] proposed an empirical relationship in the context of tectonic shear ruptures,

$$M_0 = \frac{\Delta\sigma}{\left(\frac{f_c}{0.42V_s}\right)^3}, \quad (1)$$

where V_s is shear wave velocity. It is reasonable to assume that the stress drop $\Delta\sigma$ is higher than the strength of snow (10^{-4} MPa) and lower than that of ice (1.5 MPa).

This relation, expressed in terms of M_w and shown in Figure 14, can be applied only if we adopt a liberal definition of slip, in order to incorporate the diversity of icequake events. By “liberal” we mean that a classic shear model is used to represent a vast variety of sources, as, for example, it is commonly done for hydraulic

fracture or for ice rift-widening events [e.g., *Eaton et al.*, 2014; *Bassis et al.*, 2007]. Obviously, with a mixture of magnitudes and source mechanisms, which are sometimes different from the shear fault assumption [e.g., *Walter et al.*, 2011], it is difficult to draw final conclusions. Nevertheless, most of the cryospheric data plotted in Figure 14, as well as previous stress drop calculations for seismic signals of glacial origin [*Teisseyre et al.*, 2004; *Caplan-Auerbach et al.*, 2004; *Bassis et al.*, 2007; *Winberry et al.*, 2009b; *Walter et al.*, 2011; *Helmstetter et al.*, 2015a], are in line with Point 1 above (where we noted that the largest earthquakes produce energy at the lowest frequencies). Overall, the indicated dependence is consistent with common values of glaciological stress (measured driving stresses for terrestrial ice bodies range between 14 and 410 kPa) [*Cuffey and Paterson*, 2010]. Note, for example, that for tectonic earthquakes, with significantly higher $\Delta\sigma$ (e.g., 0.1–100 MPa) and higher V_s (e.g., 3.9 km s^{−1} in the lower crust for PREM [*Dziewonski and Anderson*, 1981]), a uniform f_c will be shifted to the right and thus will miss most of the cryospheric data (as indicated in Figure 14).

We also remark that some data, seemingly at lower frequencies than our calculations suggest (Figure 14), were indicated as having lower frequency energy than could be reliably detected by the authors of the corresponding studies [*Heeszel et al.*, 2014a; *Allstadt and Malone*, 2014; *Thelen et al.*, 2013]. In particular, studies of repeating stick-slip icequakes at Mount Rainier volcano by *Thelen et al.* [2013] and *Allstadt and Malone* [2014], suggested that the low-frequency content (1–5 Hz) of events with the magnitude range approximately between −1 and 0 was a result of wave propagation through heterogeneous, low-velocity layers in volcanic terrain. This was supported by active and passive source records obtained nearby on ice and farther away on rock. Another example is a study of microseismicity within a propagating ice shelf rift by *Heeszel et al.* [2014a]. Their analysis did not consider high-frequency data (>10 Hz) for the sake of stability of the full-moment-tensor inversions. Thus, we believe that large ranges of reported frequencies, which are shown in Figure 14, do not contradict our reasoning, since outliers can be explained by other factors.

3. The most noticeable gap in characteristic frequency (Figure 14) that is not covered in previous studies lies between 0.1 and 0.5 Hz (period 2–10 s), a range which partly overlaps with the secondary microseism and therefore provides limited opportunity for signal detection [*Walter et al.*, 2010b; *Bartholomäus et al.*, 2012]. It is possible that this background noise inhibits the detection of events with sizes of $M_w \approx 2.5$ –4.0. Hence, some authors have previously noted a lack of glacial earthquakes with such sizes [*Tsai and Ekström*, 2007; *Nettles and Ekström*, 2010].
4. For incomplete data (where one piece of information, such as magnitude or frequency, is missing), Figure 14 serves as a first-order approximation to interpolate a missing piece of information from a particular publication, in order to facilitate comparisons of different sources. In addition, this figure may serve as a guide for the design of new seismic arrays by facilitating the choice of appropriate sampling frequencies, array apertures, and types of instruments.
5. Figure 14 also includes valuable information about expected source durations (or rise times), as indicated by the secondary y axis. This axis shows source duration as a function of the most energetic frequency, f ; i.e., $\Delta t \approx 1/(2f)$ [*Stein and Wyssession*, 2003]. Glacier-related seismic sources last from ~1 ms to 50 s. Some studies have established these source durations by focusing on the physics of possible source mechanisms [*Helmstetter et al.*, 2015a; *Bartholomäus et al.*, 2012]. This is also somewhat consistent with theoretical studies [*St. Lawrence and Qamar*, 1979], which show that sudden closure of a subglacial tunnel within 0.4 s (e.g., by valving causing a rapid water pressure increase) can generate earthquakes with Richter magnitudes of −0.4 to 2.5. However, since this relationship comes from tectonic earthquakes, some major exceptions should be noted, such as the duration of stick-slip events [*Wiens et al.*, 2008] or fjord seiches, which can last 1000 s instead of 10 s (due to water resonances in fjords and tilt signal rather than elastic wave emission) [*Walter et al.*, 2013b]. For a more detailed discussion, see sections 2.2.1 and 2.5.

In addition to rise time, another Y axis showing the source length L could be added if rupture velocity V_{Ri} was relatively uniform: in general, $L \approx \Delta t V_{Ri}$ [*Helmstetter et al.*, 2015a]. Unfortunately, the wide range of rupture velocity (0.1–1.5 km h^{−1}; Table A1) does not allow this approximation, even to first order. Roughly, it can be estimated that cryospheric source lengths range between 10 cm and 100 km.

6. The longest reported seismic wave periods were related to seiches and infragravity waves [*Walter et al.*, 2013b; *Bromirski et al.*, 2015]. In fact, regarding ultralong period signals (detectable only with broadband seismometers), varying viscosity is another candidate mechanism. For this it is important to bear in mind that ice is a non-Newtonian fluid (meaning that the viscosity of ice depends on strain rate). Assuming viscoelastic rheology of ice, we can refer to the concept of Maxwell relaxation time (for equation see Table A1), which describes the deformation time scales when viscous behavior becomes important. If these time

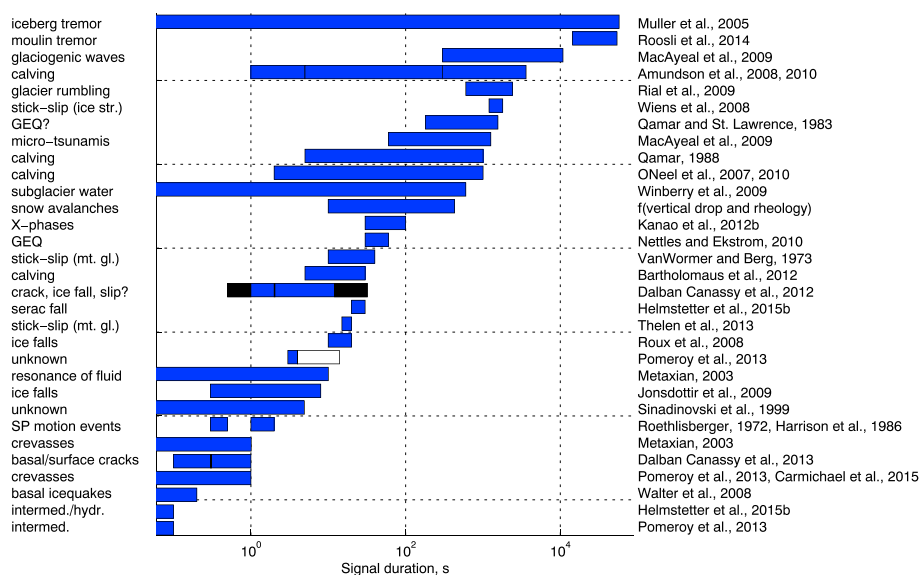


Figure 15. Typical or maximum (shown by bars “touching” the y axis) duration of reported cryogenic seismic signals. Intermediate black segments indicate mean/median duration of separate instances of different types of emissions. Black filling for the study by Dalban Canassy et al. [2012] corresponds to cracks on the left and icefalls on the right. Note that duration may be longer due to dispersion [MacAyeal et al., 2009; Bartholomaeus et al., 2012].

scales are short enough, long-period seismic waves can be related to viscosity variations [Nimmo and Manga, 2009].

For a very low viscosity (about 10^{12} Pa s), the Maxwell relaxation time can be on the order of 1000 s, which is similar to low-frequency seismic waves, like those excited by fjord seiches. (We note that 0.001 Hz falls below what most broadband seismometers reliably detect). Otherwise, for common viscosity values of terrestrial ice bodies (10^{14} Pa s), the Maxwell relaxation time is much higher (0.5–10 days) and the corresponding signals are thus undetectable with seismic instruments.

- Finally, it is interesting to consider how the presented seismic portrait of the terrestrial cryosphere (Figure 14) might look for an icy moon, such as Europa or Enceladus (moons of Jupiter and Saturn, respectively). The low strength suggested for the upper several kilometers of cold brittle ice shell [Nimmo and Manga, 2009] corresponds to lower maximum possible stress drop in equation (1) and therefore means that expected values of corner frequency, f_c , lie only slightly lower than the maximum values indicated on Figure 14. Therefore, it is unlikely that any significant differences would be observed. However, if cold-temperature ice has comparable strength to experimental values obtained at extremely cold temperatures (e.g., ~ 70 MPa at -170°C) (M. Arakawa, personal communication, 2015), then the corner frequency line will move upward. This indicates that the event-rich high-frequency band will be largely missed by broadband seismometers.

6.3. Duration of Events

Typical time scales of most seismic signals, from less than a second to tens of minutes or hours, correspond to processes too short for detection by most remote sensing techniques (Figure 15). This means that seismology is a unique approach to investigate poorly understood fast dynamic processes in the cryosphere [MacAyeal et al., 2009; Dalban Canassy et al., 2012], often taking place in remote, dark, inaccessible regions and thus giving insights into glacier dynamics, including calving (section 2.3), basal motion (section 2.2), strain variations (section 2.1), and subglacial water discharge (section 3.2). Following published reports of some of these rapid processes, some cryospheric researchers adopted high-frequency GPS (up to 0.067 Hz for long-term campaigns and 1 Hz for short-term deployments) to improve monitoring capabilities [Pratt et al., 2014]. In seismology, the 1 Hz GPS has been used since at least the year 2000 [Larson et al., 2003].

6.4. Frequency-Magnitude Relations

The Gutenberg-Richter (GR) frequency-magnitude relationship ($\log_{10} N = a - Mb$) describes an inverse logarithmic relationship between magnitude M and number of earthquakes N with magnitudes $\geq M$ in a given time interval. For tectonic events, the slope of this decay (b -value) is generally about 1. It has been suggested

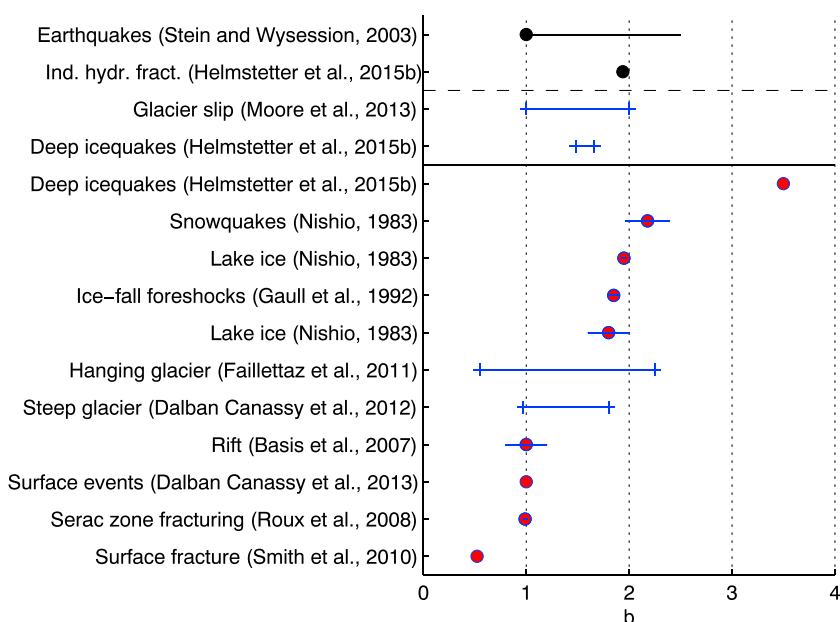


Figure 16. Comparison of b -values from different studies: data points shown below the solid line correspond to peak amplitude or relative magnitude; values indicated above the solid line are relative to interevent time (for comparison purposes, tectonic values are also indicated). In general, a higher b -value corresponds to lower stress, resulting from earthquake/volcanic swarms or induced earthquake hydraulic fracturing [Helmstetter et al., 2015a; Stein and Wyss, 2003].

that this pattern emerges from the concept of scale invariance [Stein and Wyss, 2003; Uhl et al., 2015]. Because the b -value depends on faulting style (normal faulting corresponds to a low stress, and thrust to a high stress), it can be considered a “stress meter” in the Earth’s crust [Schorlemmer et al., 2005; Schorlemmer and Wiemer, 2005]. Thrusts are producing the lowest b -values and normal faults the highest. Moreover, a change or a difference in the b -value may provide information about the rupture mechanisms [Pisarenko and Sornette, 2003; Amitrano, 2012]. Figure 16 illustrates b -values reported by different studies for a variety of cryospheric processes (instead of M , these papers refer to event peak amplitudes; this approach is known to produce the same b -value as in classic GR-relationship [Helmstetter et al., 2015b]). The dissimilar appearances of such values for different regional studies mean that there is no single characteristic magnitude-frequency relationship for most types of icequakes [Basis et al., 2007; Roux et al., 2008].

A high b -value (up to about 3.5) indicates many small quakes relative to large ones; such high values are usually associated with earthquake swarms, where no main shock occurs [Helmstetter et al., 2015a]. These swarms are often observed in volcanic environments and are related to fluid migration or caldera development [Stein and Wyss, 2003]. Nishio [1983] found that snowquakes due to thermal cracking (near Mizuho station, East Antarctica) seemed to follow the Gutenberg-Richter law with a very high b -value of 2.2 (Figure 16). However, for icequake swarms concurrent with ice shelf rift propagation, Basis et al. [2007] reported b -values only around 1. They noted, though, that due to the design of their analysis (i.e., with neglected geometric spreading and attenuation), the inferred relationship may not be accurate. Therefore, the applicability of magnitude-frequency power law scaling remains unclear for icequake swarms.

Studies focused on precursory icequakes have shown that temporal variations in b -values are predictors of ice falls [Gaull et al., 1992; Faillettaz et al., 2011; Dalban Canassy et al., 2012]. A change in b -value was associated with a change in rupture process by Faillettaz et al. [2011]. In particular, a significant change of the exponent was considered a candidate proxy for approaching catastrophic instability.

Some tectonic studies suggest that a low b -value (<1) may be associated with a locked patch (or asperity) [Stein and Wyss, 2003; Schorlemmer et al., 2005]. The corresponding conceptual model offered by Schorlemmer et al. [2005]; Schorlemmer and Wiemer [2005]; Uhl et al. [2015] interprets low b -value as an indicator of high differential stress in the fault, and high b -value as an indication of creeping sections, which continuously release stress through many small events. This hints that low b -values could potentially be observed

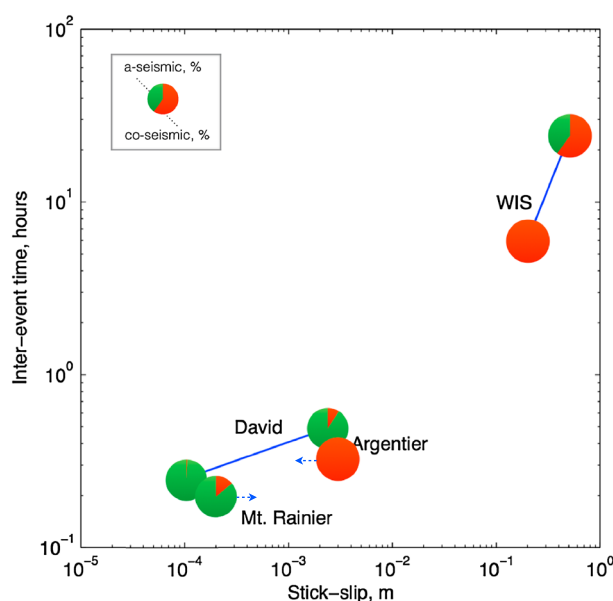


Figure 17. Examples of previously reported slip events and typical recurrence intervals [Zoet *et al.*, 2012, 2013b; Wiens *et al.*, 2008; Allstadt and Malone, 2014; Danesi *et al.*, 2007; Helmstetter *et al.*, 2015a]. Dashed arrows indicate that the amount of slip is poorly known and can be higher or lower. Pies indicate a coseismic/aseismic portion of annual ice flow (here the minimal value of slip at Mount Rainier was taken from Zoet *et al.* [2013b] and the long-term ice speed estimate of 0.8 m d^{-1} from Allstadt and Malone [2014]). Notably, a rapid slip of about 1 mm, which was not accompanied by any seismic emissions, was reported by Moore *et al.* [2013]; Bassis *et al.* [2007] estimated 1 cm coseismic slip events at the tip of the ice shelf rift.

for stick-slip events at ice streams. However, Danesi *et al.* [2007], investigating an asperity rupture under David Glacier, Antarctic, found a Gaussian-like frequency distribution rather than a power law relationship, with a characteristic magnitude $M_L = 1.3$. For the same glacier, Zoet *et al.* [2012] found a typical $M_w = 1.8$. At the same time, surprisingly, glacier slip observed at temperate Engabreen glacier, Norway, was not accompanied by any seismicity. However, according to acoustic monitoring it was associated with a reduction in b -value for acoustic events before slip, suggesting similarities to the rupture of a locked tectonic interface [Moore *et al.*, 2013]. Finally, we recall that the distribution of equivalent GEQ magnitudes in Greenland was found to deviate strongly from the Gutenberg-Richter relation and was suggested to have a characteristic size specific to each glacier ($0.6 \times 10^{14} \text{ kg m}$) [Tsai and Ekström, 2007; Nettles and Ekström, 2010].

It is argued that b -values for peak amplitude distributions are dependent on the failure mechanism, such as tensile failure induced by hydraulic fracturing and geothermal activity or shear failure for induced earthquakes [Helmstetter *et al.*, 2015b], as well as on style of faulting corresponding

to different amounts of stress [Schorlemmer *et al.*, 2005]. However, microseismic swarms associated with hydraulic fracture treatment (i.e., “fracking”) do not always follow the Gutenberg-Richter law [Eaton *et al.*, 2014]. More specifically, Eaton *et al.* [2014] showed that a widely observed falloff in large-magnitude events, universally quantified using b -value, may in some cases be an artifact of the strongly laminated character of the stimulated oil and gas reservoirs.

The exact relationship between b -value and failure mechanism remains to be understood for different types of icequakes and faulting styles (deep, intermediate, shallow, and other cases).

6.5. How Much Does Stick-Slip Contribute to Glacier Flow?

The total annual amount of coseismic slip was long believed to be small compared to aseismic movement of ice flows [Blankenship *et al.*, 1987; Anandakrishnan and Bentley, 1993]. At WIS, the opposite was shown: bi-diurnal coseismic speedups are comparable to, or larger than, the total daily aseismic movement of the ice stream [e.g., Winberry *et al.*, 2013]. A similar conclusion was reached by Helmstetter *et al.* [2015a] who recently argued that up to 100% of the total daily basal motion of an alpine glacier can be accommodated by cumulative microseismic slip. To answer the question whether or not stick-slip motion is generally a significant sliding mechanism will most likely require independent measurements, because fault dimensions, slip, and stress drop derived exclusively from seismic data are subject to large uncertainties [Abercrombie, 2015].

We provide a summary of previously reported coseismic stick-slip events at ice sheets and temperate glaciers in Figure 17, which shows total slip versus recurrence time. In several cases, up to 100% of annual flow could be accommodated by coseismic stick-slip events.

6.6. Is Glacial Stick Slip a Slow Earthquake?

Whillans Ice Stream has been frequently suggested as one of the most “simple” and suitable geological objects on the planet for studying slow earthquakes (i.e., slip events releasing energy within hours to months) and

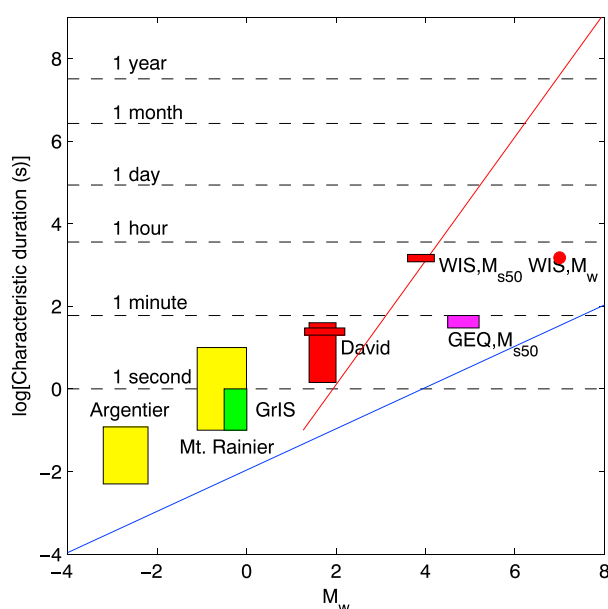


Figure 18. Comparison between magnitudes and characteristic durations of glacier stick-slip and typical scaling relations for slow earthquakes (red line) and regular earthquakes (blue line); after *Ide et al.* [2007]. Note that GEQs are not consistent with a shear slip source [Nettles and Ekström, 2010] and are shown only as an example of an unusual event. Data credits: Helmstetter et al. [2015a], Allstadt and Malone [2014], Zoet et al. [2012], Danesi et al. [2007], Wiens et al. [2008], and Nettles and Ekström [2010]; data for GrIS are courtesy of C. Röösli (personal communication, 2015).

sliding phenomena [Walter et al., 2011; Winberry et al., 2013; Pratt et al., 2014]. However, any similarity to other tectonic processes remains to be clarified.

Ide et al. [2007] found that the seismic moments of slow earthquakes are proportional to their characteristic durations ($M_0 \approx T \times 10^{12-13}$, where M_0 is in units of N m, and T in units of seconds) which is clearly different from the scaling of regular earthquakes ($M_0 \approx T^3 \times 10^{15-16}$). Both of these phenomena arise from shear slip on the plate interface, but slow earthquakes may continue for about 5 months. *Ide et al.* [2007] suggested that the source spectrum of slow earthquakes follows f^{-1} spectral decay at high frequencies; while the spectra of regular earthquakes follow f^{-2} decay. The physical mechanisms behind the two different scaling relationships remain to be understood and fall beyond the scope of the present paper. Nevertheless, it is noteworthy that there is a distinct gap between the two types of earthquakes, which can be filled only by events with a composite of two rupture modes [Ide et al., 2007].

If we compare events of glacial origin that are explained through shear slip, it becomes apparent that none of the previously proposed relationships between seismic moment and characteristic time hold (Figure 18). Durations of some icequakes scale with moment magnitudes (Figure 18) with the same slope as for regular earthquakes, but with a weighing factor at least 4–5 orders of magnitude lower ($M_0 \approx T^3 \times 10^{11}$), which is clearly different from the well-documented scaling relation for tectonic earthquakes.

Moreover, WIS events of $M_w \approx 7$ fall exactly into the “gap” between the two modes, presenting an unusual earthquake phenomenon. Note that glacial earthquakes, which are consistent with a Centroid Single Force mechanism [Nettles and Ekström, 2010], instead of a shear dislocation, are also indicated in Figure 18 in order to highlight other behaviors atypical of regular earthquakes.

7. Future Directions and Research Frontiers

Many scientific questions remain to be addressed with glacier seismology; however, the following outlook focuses on a few selected research topics. Within these topics, recent progress suggests that future research efforts could lead to substantial advances in our understanding of processes in the cryosphere and seismology.

7.1. Fault Zone Analysis

Ice streams and glaciers provide a unique way to study fault zones by proxy because (i) it is impossible to get as close to an active tectonic fault as can be done, for example, at WIS with near-source observations and (ii) glaciers are very seismically active. Therefore, whatever the relationship between slow earthquake and glacier sliding (section 6.6), monitoring of the latter has a potential to improve our basic understanding of slip phenomena and fault friction in general.

In addition, it is possible to measure rupture speed at ice streams geodetically. It was reported that stick-slip rupture speed at WIP was significantly lower than rupture speeds of earthquakes; values varied between 0.1 and 0.3 km s⁻¹ [Wiens et al., 2008; Walter et al., 2011]. More recent analysis showed higher initial rupture speeds,

around $1-1.5 \text{ km s}^{-1}$, which were close to the theoretical maximum values for rupture velocity (90% of shear wave velocity V_s in ice) [Pratt *et al.*, 2014].

The use of local high-density arrays to improve understanding of ice shelf stick-slip source mechanisms may allow us to reexamine the behavior of ice streams on older teleseismic records, from periods without array monitoring [Pratt *et al.*, 2014].

7.2. Source Processes of Glacial Earthquakes

Thirteen years after the canonical paper of Ekström *et al.* [2003], source mechanisms of glacial earthquakes are still debated and refined: the initially proposed terminus sliding model has been replaced by the action of contact forces due to iceberg capsizing, for which direct observational evidence exists. High-precision GPS measurements have shown that elastic glacial retreat due to impacts from rotating icebergs is one possible process for glacial earthquakes [Murray *et al.*, 2015]. Further research will show how the latter finding can help to answer why GEQ epicenters are not always located at calving fronts [Veitch and Nettles, 2012]. A solid understanding of glacial earthquakes would therefore not only facilitate monitoring of calving activity but could also provide insights into large-scale elastic deformation of glacier ice.

7.3. Passive Approaches to Ice Structure Imaging

Density, water content, porosity, fabric, and other properties of Earth materials affect the velocities and damping of seismic waves. For this reason, seismic waves reaching a surface seismometer from below can be used to image the structural properties of the subsurface material. The rich seismic wave field on glaciers and ice sheets eliminates the need for an active seismic source such as typically used in exploration geophysics.

7.3.1. Seismic Interferometry

The past one to two decades have witnessed rapid advances in imaging and monitoring of the Earth's subsurface structure with passive signals such as noise or naturally occurring earthquakes. These techniques have yet to be fully harnessed in seismological studies of the cryosphere.

In solid Earth seismology, the developments were sparked by the discovery that the impulse response ("Green's Function") can be reconstructed from purely passive recordings (see Curtis *et al.* [2006] and Wapenaar *et al.* [2010], for reviews). In particular, cross correlation of the two records from seismometer pairs yields the impulse response as long as the sources of seismic background noise are stationary and/or their emitted wavefield is equipartitioned. Equipartition, in turn, requires that seismic waves equally occupy all propagation modes and all source information is consequently lost [e.g., Campillo and Paul, 2003]. Since no seismic source has to be active at the locations where the seismic records are cross correlated, the recovered impulse response is often called a signal of a virtual source. This technique, referred to as "seismic interferometry," can be applied to image and monitor seismic velocity structures on continental tomography scales [Yang *et al.*, 2007], in volcano [Brenguier *et al.*, 2008a] and fault zones [Brenguier *et al.*, 2008b], and in small-scale samples in the laboratory [Hadziioannou *et al.*, 2009]. With this method, recently, Mordret *et al.* [2016] proposed to infer Greenland's ice sheet mass balance using ambient seismic noise in near real time.

Inhomogeneities in the Earth's crust give rise to coda waves consisting of reflections and multiply scattered waves [Aki and Chouet, 1975]. The same effect can provide an equipartitioned wavefield to allow for Green's Function retrieval via cross correlation [Hennino *et al.*, 2001; Campillo and Paul, 2003]. In contrast, glacier ice is highly homogeneous with scattering crevasses mostly confined to the surface crevasse zone, typically 20 m thick [Cuffey and Paterson, 2010]. As a result, icequakes usually lack a sustained coda. This is illustrated in Figure 19, which compares an icequake record from the flat Gornergletscher tongue with a record from an unstable high-altitude hanging glacier. Whereas pervasive fracturing within the hanging glacier generates a pronounced coda, the surface crevasses on Gornergletscher's tongue have little effect on the propagation of seismic waves. Consequently, the direct *P* and Rayleigh waves dominate the Gornergletscher seismogram.

The lack of scattering in glacier ice inhibits use of coda waves or continuous seismic background noise to retrieve impulse responses between on-ice seismometers. However, in principle, the surface waves of an impulse response can also be obtained from seismic sources distributed on a loop, which surrounds a seismometer pair [Wapenaar, 2004]. In this case, the cross correlations between the pair record have to be integrated over the entire loop. Surface crevasse icequakes can thus be used to recover virtual surface waves traveling between a seismometer pair [Walter *et al.*, 2015a] (Figure 3). In the presence of enough and well-distributed surface icequakes, the recovered virtual signals in principle allow for monitoring the glacier's subsurface structure.

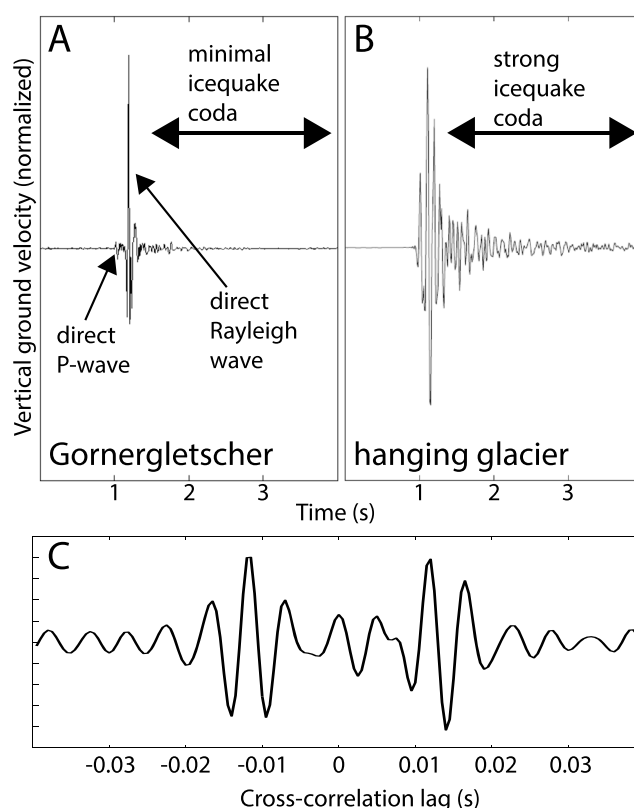


Figure 19. (a, b) Icequake signals and (c) Green's Function retrieval. Due to lack of scatterers, glacier ice typically gives rise to only limited coda waves (Figure 19a). An exception is the highly damaged ice of an unstable hanging glacier (Figure 19b). Seismic interferometry on direct surface wave arrivals can nevertheless retrieve the Green's Function response between two seismometers (Figure 19c). Reprinted from the *Geophysical Journal International* [Oxford Journals; *Walter et al.*, 2015a, Figure 14a].

split into two waves polarized along the symmetry axes of anisotropic media (see *Wuestefeld et al.* [2010], for a review). In contrast to explosions, naturally occurring seismic sources, such as shear dislocations, efficiently generate *S* waves. For microseismic stick-slip events in Antarctica, shear wave splitting seems to be a prominent feature [*Blankenship et al.*, 1987; *Harland et al.*, 2013; *Smith et al.*, 2015]. Using basal stick-slip icequakes beneath Rutford Ice Stream, Antarctica, *Harland et al.* [2013] analyzed shear wave splitting to infer an anisotropy. This anisotropy can be explained by a superposition of crystal alignment near the ice stream base expected for high shear deformation and preferred directions of englacial fractures. Together with records of seismic waves refracted into the underlying bed substrate [*Roeoesli et al.*, 2016a], shear wave splitting makes deep icequakes a promising tool to study the beds of ice streams and glaciers.

7.3.3. Receiver Functions

Seismic receiver functions are a powerful technique to elucidate conversion between *P* and *S* energy. These conversions occur as seismic waves cross planes between contrasting material in the Earth's crust, such as the ice-bed interface. A common approach is to isolate the path effect as teleseismic *P* waves travel nearly vertically upward toward a receiver at the surface. Equivalently, the receiver function strips teleseismic seismograms of their source time function, typically achieved by deconvolving one seismogram component from another [e.g., *Park and Levin*, 2000; *Zhu and Kanamori*, 2000; *Helfrich*, 2006].

The receiver function technique highlights *P-S* conversions at the bed of ice sheets but also depends on the *P*-to-*S* velocity ratio of ice. Consequently, when ice sheet thickness is known from radio echo sounding or deep drilling, the ice's seismic velocity structure can be measured. Receiver functions of seismic records on the polar ice sheets thus show a bottom-ice layer making up about 1/3 of the ice column, which exhibits seismic anisotropy and low *S* wave velocities due to the presence of liquid water [*Wittlinger and Farra*, 2012, 2015].

7.3.2. Basal Signals and Shear Wave Splitting

Structural anisotropy influences both the elastic [*Diez and Eisen*, 2015] and viscous [*Placidi et al.*, 2006] properties of glacier ice. Seismic signals can therefore be used to study differences in viscous deformation of different ice layers. Since seismic waves of basal icequakes sample the entire ice column and are abundant in stick-slip regions, they are particularly suitable for anisotropy studies.

Ice anisotropy has attracted much scientific attention as it results from and influences ice flow: On the one hand, anisotropy may result from preferred crystal alignment as a consequence of a specific deformation mode like simple shear or uniaxial compression [*Diez et al.*, 2014]. On the other hand, the effect of crystal orientation and size can change ice viscosity by a factor of 50–100 [*Dahl-Jensen et al.*, 2013]. Anisotropy measurements can therefore provide key insights into ice deformation history and provide parameter input into predictive ice sheet models.

The velocity of seismic *S* waves depends on their polarization with respect to the anisotropic symmetry axes. Arbitrarily polarized seismic *S* waves are therefore

Receiver functions can also detect sediments between the ice sole and the bedrock. As the acoustic impedance between ice and sediment is weaker than that of sediment and bedrock, *P*-to-*S* conversion primarily occurs at the sediment-bedrock interface [Anandakrishnan and Winberry, 2004]. Sediment thicknesses can thus be derived from the mismatch between measured receiver functions and those expected for an ice sheet, which lies directly on bedrock and whose thickness is known independently. The technique has shown the presence of tens to hundreds of meter thick sediments beneath the Antarctic and Greenland ice sheets [Anandakrishnan and Winberry, 2004; Wittlinger and Farra, 2012; Walter et al., 2014].

The presence or absence of subglacial sediments plays a key role in ice dynamics [Anandakrishnan et al., 1998]. Therefore, receiver functions provide an important piece of information for understanding large-scale ice sheet flow. As receiver functions can be calculated for single seismometer stations at the ice surface, future seismic deployments will be a valuable substitute or supplement to other geophysical methods targeting ice sheet beds.

7.3.4. Other Passive Techniques

Even though limited englacial scattering complicates the interferometry approach, various traditional seismological techniques are available to study the structure of glacier ice. Recent investigations have focused on Antarctic ice shelves, which can suddenly disintegrate in the event of pervasive fracturing [MacAyeal et al., 2003]. Fracture toughness of shelf ice is related to ice density and elastic modulus [Rist et al., 2002]. Therefore, seismic imaging and monitoring of ice structure may reveal how resistant an ice shelf is to external factors promoting disintegration, such as hydrofracturing in response to meltwater ponding [Scambos et al., 2003] and tsunamis [Brunt et al., 2011].

Continuous noise records at the ice shelf surface contain signals from reverberations within the sub-ice shelf cavity and the spectral character of this signal thus provides insights into the velocity structure and thickness of the ice shelf [Zhan et al., 2014]. Alternatively, Rayleigh-Lamb waves can be used. These phases are typical for thin plate geometries, travel within ice shelves and are efficiently generated when long-period (50–250 s) infragravity waves impact the ice front [Bromirski et al., 2010, 2015].

The phenomenon of trapped and guided waves has also been observed for earthquake signals traveling through the Greenland Ice Sheet [Toyokuni et al., 2015a]. A characteristic type of ice sheet-guided *S* wave, resulting from superimposition of surface/bed reflections, was proposed and named “Le.” Due to the influence of the ice sheet, records of deep teleseismic earthquakes may thus appear with amplified high-frequency content [Tsuboi et al., 2015].

Finally, when seismic signals are coherent throughout the recording network, array techniques can be used to calculate phase velocities and propagation direction of incoming waves [e.g., Rost and Thomas, 2002]. The coherence condition is often met for small networks (less than 1 km aperture) on glacial ice at frequencies below 10–18 Hz. If surface phases dominate the wavefield, determining phase velocities at different frequencies yields a dispersion relationship, which can be used to estimate ice thicknesses and determine structure as it was done locally for the Greenland Ice Sheet [Walter et al., 2015a] and for the Ross Ice Shelf [Diez et al., 2016]. Equivalently, MacAyeal et al. [2015] used waveform coherence to identify seismic, hydroacoustic, and flexural gravity waves recorded on the top of a tabular iceberg in Antarctic.

7.3.5. Ambient Seismic Noise and Water Tremor

Continental Antarctica and its ice streams have low ambient seismic noise levels with some borehole stations being candidates for the most quiet stations in the world—see Anthony et al. [2015]. In contrast, water flow through englacial and subglacial drainage channels constitutes a prominent ambient noise source in high-melt areas, such as the ablation zone of the Greenland Ice Sheet [Röösli et al., 2014] or on mountain glaciers [Pomeroy et al., 2013; Heeszel et al., 2014b]. The potential to use this water tremor signal to quantitatively monitor subsurface drainage channels in glaciers has only recently been recognized [Bartholomaus et al., 2015b; Gimbert et al., 2016; Roeoesli et al., 2016b]. Passive seismic monitoring could therefore soon fill a critical observational gap in glacier hydraulics.

7.4. Monitoring

Long-term on-ice seismometer deployments to monitor glacier dynamic processes are often complicated by difficult site conditions, in particular high surface melt rates. To overcome this, shallow borehole seismometers requiring minimal maintenance have proven useful, but technological improvements are still needed to maximize data return [Dalban Canassy et al., 2016]. Similarly, seismic monitoring of unstable ice masses for the

purpose of break-off precursor detection or process studies [Zoet *et al.*, 2013a; Kanao *et al.*, 2012b; Faillettaz *et al.*, 2016] will benefit from improvements in real-time data communication.

Importantly, increasing volumes of seismic data streams from glacier environments demand more efficient analyses. Identifying ongoing changes in the cryosphere with seismic catalogs will require improved automated waveform discriminators [O'Neel *et al.*, 2007; Köhler *et al.*, 2012; Hammer *et al.*, 2015]. Specifically, statistical analysis and quantification of calving volumes still suffer from poor accuracy and will have to move toward site independence to avoid laborious seismic detection tuning, which is unrealistic when studying extended geographic regions.

8. Concluding Remarks

The present review of cryoseismology covered various topics including iceberg calving, basal motion, glacier hydraulics, and englacial fracture development. This wide range of observations reflects the variety of geophysical processes, which emit seismic waves. Although there exist many other important multidisciplinary research fields in Earth science, it is striking that most of the presented discoveries were made only recently, in the last one to two decades. This likely reflects improvements in seismic instrumentation, which has become portable enough to allow deployment in remote terrain and harsh polar conditions.

To date there is no sign that the growth of cryoseismology research will decelerate. Glaciologists are only beginning to harness the three major advantages, which seismology has over traditional glaciological measurement techniques: First, seismic waves can be used to look into the glacial body, whereas by convention glaciologists have mostly targeted the ice surface. Similarly, passive seismic measurements provide insights into the processes of major calving fronts, which belong to the most inaccessible environments in the cryosphere. Second, a single seismometer network can target an extended region of a glacier and ice sheet including the ice-bed interface, which otherwise can only be accessed at distinct locations with expensive deep drilling. Third, recording at hundreds or even thousands of Hz, seismology provides an unrivaled temporal resolution.

The study of basal sliding best highlights the potential of seismology in cryosphere research: Seismometers played the pivotal role in the discovery of sudden stick-slip motion. This, in turn, can only be explained in terms of elastic strains and frictional sliding, which were neglected in traditional sliding theories of hard and soft glacier beds.

At the same time, cryosphere research is also of high value to seismologists. As discussed, seismic sources in the cryosphere occupy parameter ranges of stress drop, signal duration, and fault mechanisms, which are distinct from tectonic earthquakes. Cryoseismology therefore challenges traditional seismic source theories by shedding light from new directions. In this sense, the Whillans Ice Stream is arguably the most important study site, because it hosts various modes of sliding processes, which have their counterparts on much less accessible tectonic faults.

The mutual benefit of combining glaciological and seismological research also requires an in-depth dialog between the two scientific communities. It is the goal of this review to facilitate such a dialog. With future improvements in sensor technology and data handling, we expect seismology to soon provide new key observations of ice dynamic processes, for which few or no alternative monitoring techniques exist.

Appendix A: Technical Notes

These technical notes provide (i) an introduction to the basic, commonly used in seismology analytical methods, software, techniques of data handling and instrumentation; (ii) tables with an overview of key material properties and useful equations, instrumentation used in previous geophysical experiments, as well as a list of main semipermanent networks in glacial regions; finally, (iii) a figure showing examples of some seismic arrays on ice surface and around glaciers.

A1. Suggestions for Analysis Tools

A1.1. Key Material Properties and Useful Equations

For additional details on the fundamentals of glaciology, seismology, commonly used array techniques, and active seismic exploration in cold regions, see Cuffey and Paterson [2010], Stein and Wysession [2003], Bormann [2012], Rost and Thomas [2002], and Röthlisberger [1972].

Table A1. Various Properties Important to Seismic Methods in Cryospheric Environments

Property	Value	Bibliographic Reference
P wave velocity in ice, $V_{p,ice}$	3600–3900 m s ⁻¹	Röthlisberger [1972]
Shear wave velocity in ice, $V_{s,ice}$	1700–1950 [1610, 2250] m s ⁻¹	Röthlisberger [1972] (outliers reported by Helmstetter <i>et al.</i> [2015b] and Bassis <i>et al.</i> [2007] respectively)
Rayleigh wave velocity in ice, $V_{R,ice}$	1650–1668 m s ⁻¹	Roux <i>et al.</i> [2010], Mikesell <i>et al.</i> [2012] for $f = 45$ Hz
Q_p - Earth model	600	Walter <i>et al.</i> [2009]
Q_s - Earth model	300	Walter <i>et al.</i> [2009]
P wave velocity in firn, $V_{p,firm}$	500 m s ⁻¹	Zhan <i>et al.</i> [2014]
P wave velocity in cold sea water, $V_{p,water}$	1460–1470 m s ⁻¹	Bartholomäus <i>et al.</i> [2012], MacAyeal <i>et al.</i> [2015]
P wave velocity in air, $V_{p,air}$	333 m s ⁻¹	Walter <i>et al.</i> [2010b]
Rupture velocity in ice, V_{Ri}	0.1–1.5 km h ⁻¹ or $0.9 \times V_{s,ice}$	Wiens <i>et al.</i> [2008], Walter <i>et al.</i> [2009], Walter <i>et al.</i> [2011], Zoet <i>et al.</i> [2012], Pratt <i>et al.</i> [2014]
P wave velocity in granitic bedrock, $V_{p,rock}$	5000 m s ⁻¹	Walter <i>et al.</i> [2010b]
S wave velocity in granitic bedrock, $V_{s,rock}$	2550 m s ⁻¹	Walter <i>et al.</i> [2010b]
Poisson ratio, ν_{ice}	0.32	Heeszel <i>et al.</i> [2014a]
Seismic efficiency, e	1–10%	VanWormer and Berg [1973], Weaver and Malone [1979], St. Lawrence and Qamar [1979]
Young's modulus of ice, G_{ice}	9.4 GPa	Heeszel <i>et al.</i> [2014a]
Shear modulus of ice, μ_{ice}	$\mu_{ice} = V_{s,ice}^2 \times \rho_i$	e.g., 2.3–3.55 GPa [Helmstetter <i>et al.</i> , 2015a; Mikesell <i>et al.</i> , 2012], Carmichael <i>et al.</i> [2012], Heeszel <i>et al.</i> [2014a]
The first Lamé constant, λ	$\frac{\nu G_{ice}}{(1+\nu)(1-2\nu)}$	(e.g., 5.71–6.55 GPa [Mikesell <i>et al.</i> , 2012; Carmichael <i>et al.</i> , 2012; Heeszel <i>et al.</i> , 2014a])
Tensile strength of ice, σ_{ice}	1.5 MPa	
Effective viscosity of ice, η_{ice}	$10^{12} - 10^{15}$ Pa s	
Maxwell relaxation time, τ_M	$\frac{\eta_{ice}}{G_{ice}}, 2.8 \times 10^2$ to 3.4×10^5 s	Nimmo and Manga [2009], Stein and Wysession [2003]
Characteristic damping of a resonating subglacial conduit, Q_{iw}	3.0, 3.4	$f = 3.0, 1.2$ Hz [Métiexian, 2003; Winberry <i>et al.</i> , 2009b]; see Lipovsky and Dunham [2015b] for evaluation methods of Q_{iw}
Quality factor, Q_{ice} - strongly depends on wave type (surface, body; P or S , etc.), frequency and temperature	2.2–7.7, 35	$f = 3.5, 20$ Hz [Jones <i>et al.</i> , 2013; Rösli <i>et al.</i> , 2014], Peters <i>et al.</i> , 2012]

Typical properties of ice and snow used for velocity models and other modeling purposes are summarized in Table A1. Note that many parameters are generalizations and vary depending on temperature, density, salinity, etc. (see Röthlisberger [1972] for details). Thus, seismic velocities may vary, for example, with the seasons [Hunkins, 1960]. Some useful “quick-pocket” formulas are shown in Table A2.

The most crucial element for seismic data analysis is the velocity model. For ice bodies it can be composed of several layers representing firn, ice, and water. It can be 1-D [Zhan *et al.*, 2014], 2-D, or 3-D [Dalban Canassy *et al.*, 2013] depending on the objectives of a study. Indeed, this also implies that bedrock topography or bathymetry should be known for defining a geometric domain and interpreting source locations. Therefore, availability of radar- or sonar-derived profiles is important for reliable judgments about the observed phases and sources.

Constructing an accurate velocity model may be more difficult if underlying sediments, like glacial till, or bedrock layers are to be considered. These usually do not have well-known thickness and mechanical properties but can strongly affect seismic wave radiation, especially from basal events.

Many techniques allow one to estimate seismic velocity structure. Here we consider only ways previously used in cryospheric studies: (i) active seismology using explosives [Mikesell *et al.*, 2012]; (ii) ambient noise

Table A2. Simple Equations Commonly Used in Relevant Analysis

Evaluated Property	Equation	Bibliographic Reference
Fundamental resonance frequency, f_r	$f_r = \frac{V_s}{2H}$	<i>Métaxian [2003], O'Neel et al. [2007]</i>
Basal shear stress, S_b	$S_b = \rho g h \alpha$, where α is the surface slope	<i>Cuffey and Paterson [2010]</i>
Radiated seismic energy due to slip, E_s	$E_s = e S_b A_b d$, where S_b - shear modulus (the rigidity at the source depth), A_b - the fault area, d - dislocation on the fault	<i>Weaver and Malone [1979]</i>
Poisson ratio, ν	$\nu = \frac{(V_p^2/V_s^2)-2}{2(V_p^2/V_s^2)-1}$	<i>Deichmann et al. [2000]</i>
Rayleigh wave velocity, V_R	$0.92 V_{s,ice}$, for $\nu = 0.25$	e.g., <i>Roux et al. [2010]</i>
Distance from a station to a source	$r = \frac{V_p V_s (t_s - t_p)}{V_p - V_s}$	(assuming isotropic homogeneous half-space); see also <i>Helmstetter et al. [2015b]</i> for possible improvement of accuracy of this method
Moment magnitude, M_w	$M_w = 2/3(\log_{10} M_0 - 9.1)$, where M_0 is in Nm (10^7 dyn cm)	<i>Bormann [2012]</i>
Energy-magnitude relationship	$\log_{10} E_0 = 1.5(M_w) + 11.8$, where E_0 is in ergs (10^{-7} J)	<i>Stein and Wysession [2003]</i>
Duration magnitude, M_d	$M_d \approx -0.9 + 2 \log_{10}(d)$, where d is in seconds	see <i>Barruol et al. [2013]</i> for reference
Volumetric change due to tensile opening, ΔV or δV (m^3)	$\delta V = \frac{M_{iso}}{\lambda + 2\mu} = \frac{M_{iso}}{\rho \alpha^2}$ $\Delta V = \frac{M_{iso}}{\lambda + 2\mu/3}$ where M_{iso} is the isotropic moment of a moment tensor, λ is the first Lamé constant, μ is the shear modulus, ρ is the mass density, α is the P wave speed. Note, that $\Delta V/\delta V = 1.8$ if $\lambda = \mu$; however, both equations are correct, but δV is preferable [<i>Richards and Kim, 2005</i>].	see <i>Carmichael et al. [2012, 2015], Walter et al. [2009]</i> , or <i>Mikesell et al. [2012]</i> for reference
Attenuation of the wave amplitude due to geometric spreading and anelasticity (for a homogeneous medium) at distance R_i from the point source	$A(R_i) = A_0 \frac{e^{-\alpha R_i}}{R_i^n}$, where $\alpha = \frac{\pi f}{Q\beta}$, A_0 - amplitude at the source, β - wave velocity, and $n = \begin{cases} 1 & \text{for body waves,} \\ 0.5 & \text{for surface waves.} \end{cases}$	see <i>Mikesell et al. [2012], Dalban Canassy et al. [2012], Jones et al. [2013] or Rösli et al. [2014]</i> for references
Glacial hydraulic fracture half width, ω_0 , and length, L	$L = \frac{1}{2} \left[\pi \nu \left(\frac{G^*}{\rho_0} \right)^2 \frac{Q_{iw}^2}{f^5} \right]^{\frac{1}{6}}$, $2\omega_0 = Q_{iw} \sqrt{\frac{\nu}{\pi f}}$, where $G^* = \frac{\mu_{ice}}{(1-\nu_{ice})}$, ν - kinematic viscosity of water ($1.787 \times 10^{-6} m^2 s^{-1}$ at $0^\circ C$), ρ_0 - density of water ($1000 kg m^{-3}$), Q_{iw} is a quality factor, and f is a characteristic resonant frequency.	<i>Lipovsky and Dunham [2015b]</i>

Table A2. (continued)

Evaluated Property	Equation	Bibliographic Reference
Relationship between dispersion and source to receiver distance (deep water: $h > 0.5\lambda$)	$\frac{df}{dt} = \frac{g}{4\pi\Delta}$, where g is acceleration due to gravity, h is water depth, and λ is wavelength.	<i>MacAyeal et al.</i> [2009]
Shallow water ($\lambda > h$) undispersed relation for gravity wave phase (and group) velocity	$c = \sqrt{gh}$	<i>MacAyeal et al.</i> [2009]
Critical refraction angle (α_c)	$\alpha_c = \arcsin(V_1/V_2)$	<i>Helmstetter et al.</i> [2015a]

tomography [Zhan et al., 2014; Diez et al., 2016]; and (iii) phase velocity measurements using naturally occurring glacier seismicity [Walter et al., 2015a];

Software packages for analysis and visualization:

For data analysis, signal processing, picking, and other basic or more advanced operations, there are several seismic processing and plotting software packages available for most common data formats, including SAC, mini-SEED, and .mat (an HDF5 variant). Many of these are open source. A short list appears below (see the more comprehensive, albeit no longer maintained, Orfeus list at <http://www.orfeus-eu.org/software.html>).

Software packages:

1. SeisComP 3;
2. Seismic Analysis Code (SAC) [Goldstein et al., 2003];
3. ObsPy (open source) [Krischer et al., 2015];
4. PQL II (open source) [McNamara and Boaz, 2011];
5. Pyrocko (Snuffler, etc.) (open source);
6. Kiwi tools for moment tensor inversions (open source) [Cesca and Heimann, 2013];
7. Qseis for Green's functions (open source) [Wang, 1999];
8. Geopsy package for inversion of Rayleigh wave dispersion (www.Geopsy.org);
9. MSNoise package for retrieving dv/v from seismic noise (www.msnoise.org); and
10. GMT (open source) [Wessel et al., 2013].

High-level programming languages:

1. Matlab (e.g., the GISMO suite);
2. Octave (open source; note that Octave runs most Matlab code without modification); and
3. Scilab (open source).

A2. Detection and Location Algorithms

The choice and design of a signal detection algorithm, and its degree of complexity, depends on study objectives, the type of icequakes to be identified, and trace quality/characteristics. Below, we list the most common approaches useful for icequake signal detection and source location.

1. Classical short-term/long-term average (STA/LTA) threshold-based triggering algorithms for time domain [Bassis et al., 2007; Walter et al., 2008; Roux et al., 2010; Dalban Canassy et al., 2012; Carmichael et al., 2012; Barruol et al., 2013; Rösli et al., 2014; Köhler et al., 2015; Podolskiy et al., 2016]. STA/LTA in the frequency domain may be useful for detecting emerging tremors with highly localized frequency content [O'Neel et al., 2007]. To discriminate between cryogenic and tectonic events, Köhler et al. [2012] applied a self-organizing map algorithm on STA/LTA output. Jones and van der Baan [2015] developed an improved STA/LTA alternative using a two-state hidden Markov model, which might be more appropriate for the closely spaced events of icequake swarms.
2. Automatic icequake signal identification with intelligent event discrimination routines, e.g., treating the data with a hidden Markov model, HMM [e.g., Hammer et al., 2015]. Application of HMMs dates back to at least Ohrnberger [2001]. Similar earlier papers by Beyreuther and Wassermann [2008] and Beyreuther et al. [2012] took a nearly identical approach to Hammer et al. [2015]. For additional references, see Benitez et al. [2007] and Ibanez et al. [2009].
3. A search for similar waveforms (or detection of so-called "multiplets") can be performed by choosing template events and cross correlating them with continuous seismic data from one or all channels [Helmstetter et al., 2015a; Carmichael et al., 2012; Mikesell et al., 2012; Thelen et al., 2013; Allstadt and Malone, 2014].

A high degree of similarity (i.e., high correlation) can mean that events share a source mechanism and hypocentroid [Helmstetter et al., 2015a; Harris, 2006].

4. For events with clear *P* and *S* waves, source to receiver distances can be inferred from the time lag between these phases (Table A2). Furthermore, arrival directions (back azimuth and incidence angle) can be estimated from the polarization of the particle motion after making necessary corrections related to free-surface reflections [Helmstetter et al., 2015b]. This can be more accurately derived through a tensor decomposition approach, taking into account the diversity of body wave velocities [Raimondi et al., 2016].
5. Amplitude decay models (see Table A2) and grid search location algorithms are useful for tremor-like signals with emergent or indistinguishable first arrivals [Jones et al., 2013; Rösli et al., 2014].
For a more comprehensive survey of techniques, see examples of amplitude-based grid searches in the volcanic tremor and ETS literature [Konstantinou and Schlindwein, 2003]. Other useful techniques appear in Ghosh et al. [2009], Jones et al. [2012], and Husker et al. [2012].
6. Nonlinear probabilistic location approaches (using phase arrival times and associated picking uncertainties) provide realistic, nonelliptical location uncertainties (usually within 95% confidence limits) [Carmichael et al., 2012; Dalban Canassy et al., 2013; Rösli et al., 2014; Pratt et al., 2014; Helmstetter et al., 2015a; Smith et al., 2015; Roeoesli et al., 2016a].
7. Beamforming methods [Bormann, 2012; Barcheck et al., 2013; Pratt et al., 2014; Koubova, 2015]: A given seismic signal from a given arrival direction (i.e., back azimuth, incidence) appears at each array station at a different time. Beamforming methods grid search over a multiparameter solution space (often back azimuth and apparent velocity) based on corresponding time delays. The optimal solution corresponds to the maximum energy of the summed signal. In other words, the single traces, shifted for the best delay times, after summation give the largest amplitudes due to coherent interference of the signals and suppression of incoherent noise [Bormann, 2012].
8. A “jackknife” technique, consisting of removal of some stations from a location routine for cross validation of the results, can verify that the spatial patterns do not depend strongly on array geometry [Bassis et al., 2007; Walter et al., 2013c]. Similarly, subsets of available data can be removed for testing the stability of waveform inversion [Sergeant et al., 2016].

A3. Waveform Modeling/Matching

Polarity of *P* wave arrivals on the vertical component may give hints about rupture type [Stein and Wysession, 2003]. For instance, upward (compressional) polarity is typical for tensile basal icequakes [Walter et al., 2009; Helmstetter et al., 2015b] or for tensile openings in sea ice [Crary, 1955]. Downward (dilatational) polarity can indicate crack closure [Heeszel et al., 2014b]. Since a *P* wave has the highest-frequency energy ($f > 200$ Hz), undersampling can make it impossible to determine the arrival and polarity [Helmstetter et al., 2015a].

More detailed analysis requires a full-moment-tensor inversion [Walter et al., 2009], which requires good spatial coverage of the main seismic lobes and very high frequency sampling (up to 1000–2000 Hz). (Here we remind that as already explained in the main text, many cryoseismic sources are different from shear mechanism, which can be described with a moment tensor. Therefore, other approaches are used to invert the source force history from seismograms, as it was done, for example, by Sergeant et al. [2016] without any a priori constraint on the source time function.)

The main objective of waveform modeling is to find the most likely source mechanism that matches the features of recorded signals. How close the modeled waveform matches the record is typically judged based on the misfit between observed and synthetic waveforms [Walter et al., 2009].

Calculation of synthetic seismograms, representing propagation from an assumed source to a receiver, is an iterative four-step procedure [Stein and Wysession, 2003; Walter et al., 2009, 2012]:

1. Assumption of a source time function.
2. Assumption of a velocity model.
3. Calculation of Green’s functions (response of the defined media to a force impulse of fundamental fault) and synthetic seismograms.
4. Evaluation of misfit between data and synthetics.

A4. Array Techniques

Formally, if the aperture is larger than the correlation radius of the signals (the maximum distance between stations at which time series are correlated), the group of sensors is called a network; otherwise, it is an array [Bormann, 2012].

Sensor arrays have two main advantages. (1) Signal-to-noise ratios can be improved by time domain stacking. The expected gain is $G \propto \sqrt{N}$, where N is the number of sensors in the array. (2) Compared to a single three-component sensor, signal detection capability can be improved by application of “stack-and-sum” time domain beamforming techniques, which constructively sum coherent signals and cancel out incoherent random noise [Rost and Thomas, 2002].

Two main methods of array analysis concern signals and noise:

1. F-K analysis (back azimuth and slowness) [Eckstaller et al., 2007; Hammer et al., 2015; Koubova, 2015]. This array technique provides information about the arrival directions of seismic energy and their apparent velocities (expressed as slowness = v_a^{-1}). This is particularly useful for separating nearby events from distant events that impinge on an array with higher apparent velocities [Hammer et al., 2015]. It can also be used, for example, to distinguish the air phases that lag behind some calving-generated events [Koubova, 2015]. Combination of at least two arrays enables location of seismic sources from beamforming techniques alone.
2. Site characterization. Tectonic microzonation studies usually perform S wave inversions from surface wave dispersion curves, as derived from noise (say, a 40 min record), for characterizing upper layers of sediments. This also can be done in near real time for landslide monitoring (M. Pilz, personal communication, 2013). Such an approach is new to glaciers but promises to open windows into the spatial and temporal evolution of the material properties of ice. Another method is based on receiver functions [Stein and Wyssession, 2003], which use teleseismic earthquakes to evaluate the properties of subsurface layers, like basal ice, subglacial sediments, or till [Wittlinger and Farra, 2012, 2015; Walter et al., 2014].

We note that the geometry of a seismic array both affects sensitivity to energy of different wavelengths and introduces azimuthal dependence to the accuracy of event locations. In particular, arrays have a resolution limit which is approximately equal to the aperture of an array (the longer the wavelength, the larger the array has to be) [Poggi and Fäh, 2010; Bormann, 2012]. Moreover, the orientation of an array with respect to the source is important for location quality [Bassis et al., 2007; Jones et al., 2013]. Array-based locations become increasingly inaccurate in the near field, as the assumption of a planar wave front becomes invalid [Rost and Thomas, 2002].

The best practice to minimize the azimuthal dependence of location resolution is to choose a circular geometry with irregular station spacing [Rost and Thomas, 2002]. This was done at the Watzmann array in Antarctic, for example [Eckstaller et al., 2007; Hammer et al., 2015]. Another practical wisdom is to avoid large altitude differences between arrays used to compute the back azimuths of local events; otherwise, incorrectly modeled travel distances will distort the calculation [Koubova, 2015].

High-velocity glacial flow, which is often nonhomogeneous and deforms the initial geometry of arrays, does not present significant problems for arrays installed for only a few months. For instance, using methods of Picozzi et al. [2010], it can be shown that the hypothetical horizontal shift of an array on a glacier with a parabolic traverse velocity does not change the so-called Array Response Function, or ARF, which represents a kind of spatial filter for the wavefield [Picozzi et al., 2010; Rost and Thomas, 2002].

A5. Instrumentation

A5.1. Seismometers

The broad frequency range of cryospheric seismic signals and their large dynamic amplitude range impose requirements on seismic sensors; i.e., often, there is a need to combine broadband (BB) and short-period (SP) instruments. Accelerometers are rarely used for studying glaciers due to concentration of the main frequencies of interest below 100–200 Hz [Stuart et al., 2005] and insufficient sensitivity. However, they can be useful for recording high-frequency intermediate and deep icequakes just beneath sensors [Helmstetter et al., 2015a, 2015b], which can be missed otherwise due to aliasing.

Furthermore, for near-field arrays, a low acquisition frequency may be insufficient for moment tensor inversion [Smith, 2006], and the optimal balance between data storage capacity and maximum expected frequency of interest must be chosen (i.e., according to the Nyquist criterion $f_s > 2f_{\max}$). A common seismological practice is to use sampling rate 4 times higher than the maximum frequency of interest [Stein and Wyssession, 2003].

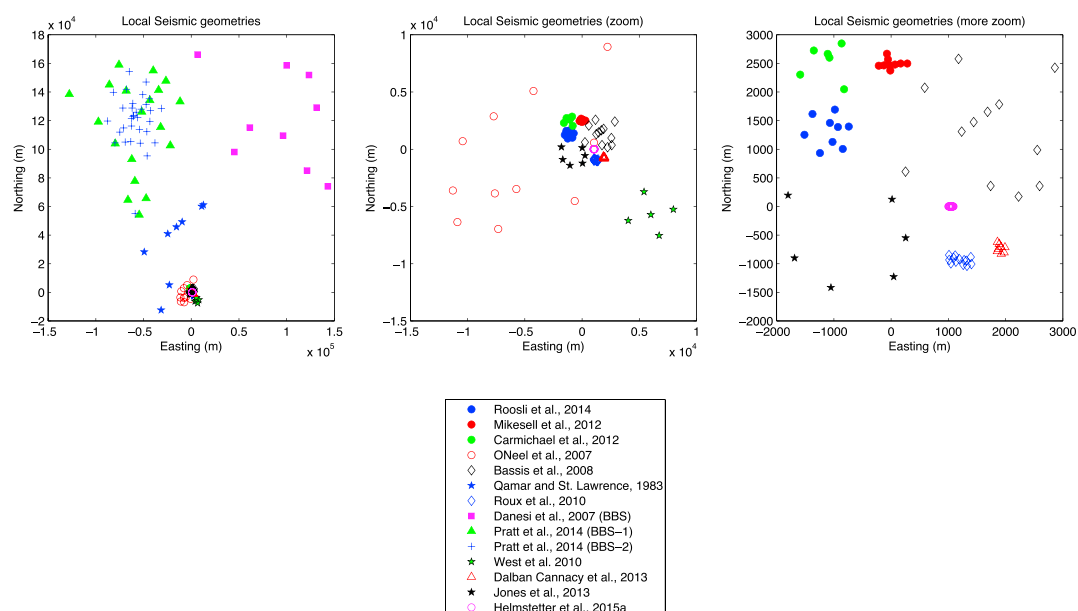


Figure A1. Examples of previously used seismic arrays on ice surface and around glaciers.

This explains why some icequake studies have relied on 500–1000 Hz acquisition frequencies [Röösli *et al.*, 2014; Helmstetter *et al.*, 2015b]. Studies on the Gornergletscher, an alpine glacier in the Swiss Alps, have used up to 4000 Hz [Walter *et al.*, 2008, 2009].

It is known that some high-amplitude near-field energy arrivals of glacial origin can clip BB or LP sensors [e.g., Walter *et al.*, 2013d; R. Genco, personal communication, 2015]. To extend observations to lower frequencies without introducing clipping, and to record static displacements, colocated GPS sensors may become a crucial supplement to broadband seismometers [Barcheck *et al.*, 2013; Pratt *et al.*, 2014]. Using horizontal component seismogram as a record of lateral motion and combining it with GPS measurements allows to reconstruct broadband displacement [Pratt *et al.*, 2014].

An aspect of on-ice instruments, which often surprises “tectonic” geophysicists, is that sensors for passive cryospheric seismology may be installed on nonstationary ice surfaces, which not only move up to several meters per day but also subside due to melting during the ablation season. Here the most crucial issue is tilting of sensors due to glacier movement or surface melting. This may be addressed by laborious weekly or even daily leveling of instruments [Métaxian, 2003; Walter *et al.*, 2008; Podolskiy *et al.*, 2016], by making snow/ice pits or by drilling boreholes [Walter *et al.*, 2013d; Morgan *et al.*, 2013; Röösli *et al.*, 2014]. For some studies, a sensor was dug into a glacial crevasse [Weaver and Malone, 1979]. Indeed, making pits for sensors reduces noise from wind and rain [Röthlisberger, 1972; Nishio, 1983; Anthony *et al.*, 2015; Smith *et al.*, 2015], while releveing causes inevitable short-term data gaps [Röösli *et al.*, 2014; Podolskiy *et al.*, 2016].

Good sensor-ice coupling is essential to ensure high-quality data [Mikesell *et al.*, 2012]. To improve it, sensors may be mounted onto concrete slabs or tile blocks before burial into ice [Jones *et al.*, 2013; Winberry *et al.*, 2013; Röösli *et al.*, 2014; Pomeroy *et al.*, 2013].

High surface ablation rates (up to 10 cm/d) are a very challenging environment for placing seismic stations on an ice surface; they may require reinstallation and releveing, and a natural consequence is a reduced window for data collection [Pomeroy *et al.*, 2013]. In order to simplify routine leveling, relatively light seismometers can be placed on specially constructed tripods [Röösli *et al.*, 2014]. For summer ablation zones, efficient melt-water drainage should also be provided by digging small channels to direct water away from seismometers. Additionally, a high-albedo cover over a pit is highly recommended; it will reduce the amount of absorbed heat (so that the pit, in which the sensor is placed, does not become too shallow) and will protect instruments from the wind and rain [Röösli *et al.*, 2014; Pomeroy *et al.*, 2013].

In cold conditions, without any melting, sensors can be placed in shallow snow pits. In this case, they can be covered with a protective hemispherical dome or bucket, then buried in snow. In heavy snow regions,

Table A3. Examples of Different Types of Recent Instrumental Setups for Observations of Glaciogenic Seismicity (the Main Provider Is IRIS)^a

Study	Site	Duration	Sensor (The Number of Sensors Is Shown in Square Brackets.)	Digitizer and Recorder	Time Lapse Photo Camera	Microphone for Audio Record	Water Level Sensor
Walter <i>et al.</i> [2010a]	Columbia Glacier (bedrock)	BB: Long-term (19 months), SP: Short term (within 4 months)	Güralp 40T BB [1]; Mark Products L22 SP [6]	Quanterra Q330, Quanterra PB44 Packet Baler or Reftek RT130	°	x	x
West <i>et al.</i> [2010]	Bering Glacier (ice surface)	Short term (3 months)	Güralp CMG-6TD [2]; Mark Products L22 SP [3]	Quanterra Q330, Quanterra PB44 Packet Baler	x	x	x
Amundson <i>et al.</i> [2008, 2010]	Jakobshavn Isbræ (bedrock)	Short term (1–4 months)	Mark Products L22 SP [1]	Quanterra Q330, Quanterra PB44 Packet Baler	°	°	°
Roux <i>et al.</i> [2010]	Gornergletscher (ice surface, borehole)	Short term (1 month)	Lennartz LE-3-D (surface) [13], Geospace GS-20DH (borehole) [1]	Geometrics GEODE (2), recording laptop computer	x	x	°
Walter <i>et al.</i> [2008, 2009]	Gornergletscher (ice surface, borehole)	Short term (1–2 months)	Lennartz LE-3-D (surface) [14–15], Geospace GS-20DH [1–3] & GS-11D [6] (shallow and deep borehole)	Geometrics GEODE (2), recording laptop computer	x	x	°
Faillottaz <i>et al.</i> [2008]	Weisshorn hanging glacier (ice)	1 month	Lennartz LE-3Dlite MkII [1]	Taurus Portable Seismograph	°	x	x
O'Neel and Pfeffer [2007]	Columbia Glacier (bedrock, ice surface/pit)	Short term (about 3 months)	Güralp 40T BB [1], L-28 geophone [6]	Reftek RT130 and experimental loggers	x	x	x
O'Neel <i>et al.</i> [2010]	Columbia Glacier (bedrock)	Long-term (19 months)	Güralp 40T BB [1], Mark Products L-22 SP [10]	Reftek RT130	°	x	x
Martin <i>et al.</i> [2010]	Iceberg (B15A) Antarctic (snow pit) + onshore bedrock	Long-term (Description of 3 days)	Güralp 40T BB [1 per iceberg] + records from Antarctic stations	Quanterra Q330, Quanterra PB44 Packet Baler	x	x	x
MacAyeal <i>et al.</i> [2009]	Icebergs (C16, B15A, B15K) Antarctic (snow pit), ice shelves, sea ice	several terms during 3 years	Güralp 40T BB [9]	Quanterra Q330, Quanterra Baler	x	x	x

^aThe power supply is usually a solar panel and Lithium batteries.

like Greenland, 5 m of snow may accumulate over the sensor during a few winter seasons, meaning that some equipment (particularly solar panels) might need to be dug out and reinstalled on the surface [Toyokuni *et al.*, 2015b].

Deep borehole sensors [Neave and Savage, 1970; Walter *et al.*, 2008; Dahl-Jensen *et al.*, 2010; Rösli *et al.*, 2014] allow better determination of hypocenter depth and significantly reduce noise [Anthony *et al.*, 2015]. In theory, if these are frozen into the ice, they may also provide better coupling.

Seismic experiments by Qamar [1988], O'Neel *et al.* [2007], Dalban Canassy *et al.* [2012], and Carmichael *et al.* [2012] confirmed that it is possible to monitor weak seismic glacial emissions with SP sensors installed on

Table A4. Main Permanent Seismic Networks Operating in Glaciated Regions and Other Useful Data Sources

Name	Location	Description	Reference	Web Page
GLISN	Greenland	33 stations	<i>Larsen et al.</i> [2006], <i>Dahl-Jensen et al.</i> [2010]	www.glisn.info
POLENET	Antarctic	growing	<i>Kanao et al.</i> [2012a]	polenet.org
USArray	Alaska	ongoing installment		www.usarray.org
Spitsbergen seismic network (NORSAR, GEOFON, GSN)	Spitsbergen	three stations and an array with nine seismometers (SPITS)	<i>Köhler et al.</i> [2015]	www.norsardata. no/NDC/bulletins/dpep
GEQ database	Greenland, Antarctic, Alaska	--	<i>Nettles and Ekström</i> [2010]	www.globalcmt.org

nearby rock surfaces or large rocks. However, for most severe environments, like continental Antarctica, rock outcrops have been found to be less favorable than pits because of stronger coupling with wind-generated noise [Anthony *et al.*, 2015]. Furthermore, in general, direct coupling of a sensor with an ice surface is recommended, because this reduces path effects; that is, it facilitates recording of clearer signals by eliminating highly “disturbed” arrivals that propagate through heterogeneous materials along the ice-rock interface. When installing multiple arrays for event location, it is important to avoid large elevation differences between arrays, which can reduce the quality of results due to increased raypath distances [Koubova, 2015].

One consequence of the requirement that sensors function in harsh, cold, and wet environments is significant improvements in the robustness of instruments [Rössli *et al.*, 2014; Anthony *et al.*, 2015]. For example, the IRIS Polar Program is now supplying seismometers and recorders for key international year-round polar networks (e.g., a cold-rated Guralp CMG-3T tested at -55° or Quanterra Q330 operating at $>-45^{\circ}$ with Lithium batteries).

Some examples of commonly used instrumental setups and network configurations (with aperture ranges from 10 m to hundreds of kilometers) are provided in Figure A1 and in Table A3. Array/network design and aperture strongly depend on the objectives of a study. As a generalized example, a typical installation for calving monitoring might comprise one or more Guralp BB seismometer(s), a few short-period instruments, and a Quanterra digitizer and recorder, with a total cost of at least 25,000 USD. More substantial costs, however, are related to helicopter/aircraft operation.

As a minimal number of stations, we recommend two, because electric spikes (for example) may resemble icequakes. Note that two stations may be sufficient for localizing calving events with emergent wave onsets [Mei *et al.*, 2016]. A common number of seismic stations is around nine. This number seems to emerge as an optimal choice for a typical limited-duration campaign with limited work hours. It is also a good compromise between maintenance cost and ideal network coverage.

A5.2. Supplementary Tools/Sensors

Although rare in tectonic studies, time lapse photography in glaciology is a relatively common accompaniment to seismic observations [Harrison *et al.*, 1986; Amundson *et al.*, 2008] (see also <http://extremeicesurvey.org>). Contrary to tectonic events, for which fault rupture is nearly impossible to document with photography, this methodology has provided some of the strongest clues about the true nature of glacier earthquake sources and provides dramatic records of glacier behavior, such as iceberg capsizing or Worthington jets launching ice fragments >100 m into the air after relatively small calving events [e.g., Amundson *et al.*, 2008; Bartholomaeus *et al.*, 2012].

To improve interpretation of complex signals, cryoseismic monitoring may also be complemented by infrasonic monitoring (e.g., to improve calving location) [Richardson *et al.*, 2010]. Connection of an infrasound sensor (e.g., from *iTem geophysics s.r.l.*) to a seismic station is relatively simple if the recorder has an auxiliary sensor plug (e.g., Guralp CMG-DAS-S6). Hydroacoustic [MacAyeal *et al.*, 2008] and water level sensors [Amundson *et al.*, 2010; Rössli *et al.*, 2014] are also useful.

A6. Networks

Recent international collaboration has significantly expanded seismic networks into polar regions, especially in the Arctic (see Table A4) [Storchak *et al.*, 2015]. The main objectives of such networks (e.g., POLENET, Greenland Ice Sheet Monitoring Network (GLISN), USArray, and SPITS) focus on tectonics and glacial changes,

as well as monitoring isostatic rebound of the lithosphere due to ice cap mass loss [Dahl-Jensen *et al.*, 2010]. Increasing teleseismic coverage coupled with on-ice near-field observations allows better location and understating of source regions [Pratt *et al.*, 2014; O'Neel *et al.*, 2010; Bartholomäus *et al.*, 2012]. Another advantage of expanding polar networks is the potential to detect meteorites strikes in remote polar regions [Kanao *et al.*, 2012a]. Many records from permanent networks are now freely available online in near real time (Table A4); in addition to those networks in the table, GLISN will soon become a part of the Federation of Digital Seismograph Networks (FDSN).

A detailed description of the evolution of seismic networks in polar regions can be found in a paper by Storchak *et al.* [2015].

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