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Massive destabilization of an Arctic ice cap

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A B S T R A C T

Ice caps that are mostly frozen at the bedrock-ice interface are thought to be stable and respond slowly to changes in climate. We use remote sensing to measure velocity and thickness changes that occur when the margin of the largely cold-based Vavilov Ice Cap in the Russian High Arctic advances over weak marine sediments. We show that cold-based to polythermal glacier systems with no previous history of surging may evolve with unexpected and unprecedented speed when their basal boundary conditions change, resulting in very large dynamic ice mass losses (an increase in annual mass loss by a factor of ∼100) over a few years. We question the future long-term stability of cold and polythermal polar ice caps, many of which terminate in marine waters as the climate becomes warmer and wetter in the polar regions.

Significance statement: A tipping point is reached as a largely cold-based ice-cap outlet glacier with no previous history of surges advanced over low-friction marine sediments, causing it to accelerate and thin.

1. Introduction

High latitude cold and polythermal glaciers and ice caps cover ∼450,000 km^2 of the Earth's surface (Gardner et al., 2013) and hold ∼0.30 m of potential sea-level rise (Radić et al., 2014). These glaciers occur in polar desert environments that are characterized by sub-freezing average temperatures and low rates of annual precipitation, typically below 250 mm/yr. Cold-based glaciers are frozen to their bed and flow by internal deformation at rates of only a few meters per year. Ice in temperate glaciers is at the melting point throughout the body of the glacier with the glacier moving primarily by sliding. Polythermal glaciers are intermediate, with some parts of the glacier cold-based and other parts wet and temperate. Some 2,300 polythermal and temperate glaciers around the planet surge, undergoing short-duration accelerations in ice flow (Copland et al., 2003; Grant et al., 2009a; Jiskoot et al., 2000; Sevestre and Benn, 2015) that manifest as a rapid glacier advance, thinning in the accumulation area and thickening at the ice front. These surge mechanisms are not understood fully, but are thought to be driven by cyclic oscillations in basal hydraulic conditions (Clarke, 1991; Harrison and Post, 2003; Kamb et al., 1985; Raymond, 1987), controlled by a combination of sub-glacial geometry, the local climate and/or the basal thermal regime (e.g. Bindschadler et al., 1976; Dunse et al., 2015; Murray et al., 2003). Surges result in a rapid transfer of ice mass from the accumulation zone of a glacier to the ablation zone and often leave a long-term glaciological and/or geomorphological signature, such as looped moraines, that can be used to identify surging glaciers in their quiescent phase (Grant et al., 2009b; Ottesen et al., 2017; Ottesen and Dowdeswell, 2006). The quiescent phase is usually an order of magnitude longer in duration than the active phase. Here we present the first observations of an outlet glacier at a largely cold-based ice cap, the Vavilov Ice Cap on October Revolution Island, Severnaya Zemlya archipelago in the Russian High Arctic [Fig. 1], that is undergoing extraordinary ac-

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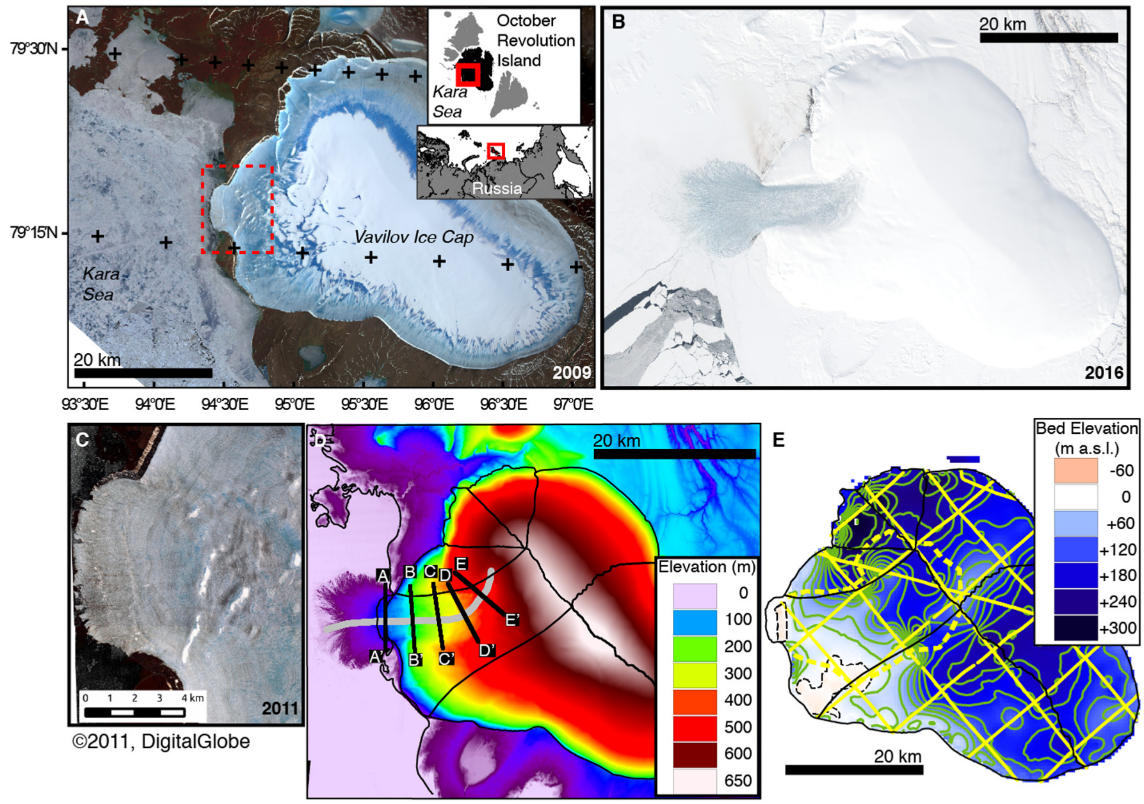

Fig. 1. Vavilov Ice Cap, Severnaya Zemlya Archipelago, Russian High Arctic. **A)** True color ASTER image from late summer, 2009: note front position and extent of summer melt. Red dotted box details region shown in C. Insets show location of Severnaya Zemlya and October Revolution Island (black). **B)** Landsat-8 true color image on May 8th 2016, note front position and crevasse field extending into the interior of the ice cap. **C)** Worldview-02 true color image showing front moraine, terminus position and continuous sediment bands in 2011. Image © 2011, DigitalGlobe, Inc. **D)** Elevation of Vavilov Ice Cap from spring 2016. Thin black lines – drainage basins from Randolf Glacier Inventory version-5 (Pfeffer et al., 2014). Thick grey line – centerline profile in Figs. 2, 3 and 4. Thick black lines – transverse profiles in Fig. 4. DEM created from DigitalGlobe Inc. stereo imagery. **E)** Bed elevation derived from airborne radio echo sounding. Green contours are 25 m intervals; dotted black contours are sea level, solid black basins outlines are from Randolf Glacier Inventory version-5. Yellow lines are flight lines used to collect radar echo sounding data; yellow dotted line is extent of crevassing in 2011 sub-meter optical imagery. Note scales in panels **A**, **B**, **D** and **E** are identical. (For interpretation of the colors in the figure(s), the reader is referred to the web version of this article.)

celeration and thinning but displays no previous surface evidence of surging (Dowdeswell and Williams, 1997), confirmed with high resolution imagery in Fig. 1c).

The ~300 to 600 m thick, 1,820 km^2 Vavilov Ice Cap [Fig. 1] is frozen to its bed over the majority of its area, apart from a region along its western margin where basal sliding is potentially important for faster flow (Bassford et al., 2006a). Similar polythermal conditions are seen elsewhere in the Arctic (Gilbert et al., 2016). Ice-core drilling near the summit suggests the oldest ice at the base of the ice cap is Pleistocene age and shows the ice cap sits on top of frozen clay and sandstone deposits (Stiévenard et al., 1996). Palynological study of the ice core suggests precipitation at the ice cap became increasingly focused on the southwest side of the ice cap about 500 yrs ago (Andreev et al., 1997; Bassford et al., 2006b). Bassford et al. (2006b) suggest this change in precipitation pattern caused the western margin to advance slowly through time. During periods of intense melt the ice cap is stripped almost entirely of its snow cover and narrow supraglacial streams flow to the ice cap periphery. When intense melt occurs, continuous sediment bands in the ice can be traced in high resolution imagery at low elevation around the full periphery of the ice cap. At the western margin, these sediment bands are disturbed compared to elsewhere on the ice cap [Fig. 1C]. The disturbed but continuous bands coincide with the region of slightly faster ice flow and slow terminus advance.

Satellite radar interferometry from 1996 observed this western region of the ice cap moving at a peak rate of around 20 m/yr (~5.4 cm/day), a rate at which modeling required some of the motion to be accommodated by basal sliding. The rest of the ice cap moved at speeds of less than 5 m/yr, consistent with internal deformation (Bassford et al., 2006b).

2. Results

In 2010, the ice in the western region started to accelerate. The front of this "fast" flow region was vertical and had a sediment apron, a configuration that is indicative of cold-based ice at the glacier front (Chin, 1991). In 2011 crevasses were observed from the western margin inland to within 5 km of the ~700 m a.s.l. summit plateau of the ice cap using sub-meter resolution satellite imagery [Fig. 1]. The extent of the crevasses matches that of the inland extent of the ongoing ice acceleration and roughly coincides with an inherited, shallow sub-glacial valley [Fig. 1]. The increasingly rapid motion observed since 2010 requires a component of basal sliding. This motion may be accommodated by the plastic deformation of subglacial sediments or at the glacier-bed interface (Bassford et al., 2006b) or possibly due to enhanced deformation in Pleistocene ice (Gilbert et al., 2016) at the bottom of the ice cap, or at the boundary between Pleistocene and Holocene ice (Stiévenard et al., 1996).

From 1952 to 1985 the western part of the ice cap advanced around 400 m (~12 m/yr) (Bassford et al., 2006b). The rate of advance accelerated significantly to ~75 m/yr between 1998 and 2011. By 2000, the grounded ice front had moved offshore into the shallow Kara Sea pushing sediment before it. Rates of advance were steady between 2000 and 2009. By the spring of 2013 the