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# Orbitally Forced Ice Sheet Fluctuations in Snowball Earth

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**Snowball Earth theory provides a powerful framework for understanding Neoproterozoic panglaciations, although some of its predictions are apparently contradicted by geological evidence. For example, Snowball theory posits that the panglaciations were terminated after millions of years of frigidity by a positive feedback, in which initial warming from rising atmospheric $CO_2$ was amplified by reduction of ice cover and planetary albedo (1, 2). This threshold behaviour implies that most of the glacial record was deposited in a brief 'melt-back' period (3), an interpretation apparently inconsistent with geological evidence for glacial-interglacial cycles in low palaeolatitudes (4-6). Here w~~W~~e use geological and geochemical evidence combined with numerical modeling experiments to reconcile these apparently conflicting views. New evidence from Svalbard (Norwegian High Arctic) indicates oscillating glacier extent and hydrological conditions within continental deposits of a Cryogenian glaciation, during a period when $pCO_2$ was uniformly high. Modeling experiments show that such oscillations can be explained by orbital forcing in the late stages of a 'Snowball' glaciation, while $pCO_2$ was rising towards the threshold required for complete melt-back. This reconciles Snowball Earth theory with evidence for the complex successions observed at many other localities.**

The Wilsonbreen Formation in NE Svalbard contains a detailed record of environmental change during the Marinoan, the second of the major Cryogenian glaciations (650-635 Ma) (7, 8). At this time, Svalbard was located in the Tropics on the eastern side of Rodinia (9, 10). The Wilsonbreen Formation is up to up to 180 m thick and was deposited within a long-lived intracratonic sedimentary basin (11). It is subdivided into three members (W1, W2 and W3) based on the relative abundance of diamictite and carbonate beds (7, 8; Fig. 1; Supplementary Figures 1 & 2). The occurrence of lacustrine sediments containing both precipitated carbonate and ice-rafted detritus throughout the succession, and intermittent evaporative carbonates and fluvial deposits, indicates ~~deposition in a closed terrestrial basin.~~ ~~T~~that the basin remained isolated from the sea throughout deposition of the Wilsonbreen Formation, consistent with~~ due to~~ eustatic sea level fall of several hundred metres ~~during the Marinoan~~and limited local isostatic depression (Supplementary Information; 12). This makes it an ideal location to investigate the possibility of climate cycles within a Neoproterozoic panglaciation, as it provides direct evidence of subaerial environments and climatic conditions.

We made detailed sedimentary logs at ten known and new localities extending over 60 km of strike (Fig. 1; Supplementary Figure 1; see Methods). Seven sediment facies associations were identified, recording distinct depositional environments that varied in spatial extent through time (Supplementary Figure 3; Supplementary Information). These are: FA1: *Subglacial*, recording direct presence of glacier ice, FA2: *Fluvial channels*, FA3: *Dolomitic floodplain*, recording episodic flooding, evaporation and microbial communities; FA4: *Carbonate lake margin*, including evidence of wave action; FA5: *Carbonate lacustrine*, including annual rhythmites and intermittent ice-rafted debris; FA6: *Glacilacustrine*, consisting of ice-proximal grounding-line fans (FA6-G) and ice-distal rainout deposits (FA6-D); and FA7: *Periglacial,* recording cold, non-glacial conditions. Additional descriptions are provided in the Supplementary Information. The vertical and horizontal distribution of these facies associations (Fig. 1) allows the sequence of environmental changes to be reconstructed in detail.

(1) The base of the Formation is a well-marked periglacially weathered horizon with thin wind-blown sands (Supplementary Figure 4a-b). This surface records very limited sediment cycling in cold, arid conditions.

(2) At all localities, the weathering horizon is overlain by fluvial channel facies (FA2) and mudstones, marking the appearance of flowing water in the basin and implying positive air temperatures for at least part of the time (Supplementary Figure 5a).

(3) Glacilacustrine deposits (FA6-D) record flooding of the basin and delivery of sediment by ice-rafting (Supplementary Figure 4c-d). Far-travelled clasts are common, indicating transport by a large, continental ice sheet.

(4) Warm-based, active ice advanced into the basin, indicated by traction tills and glacitectonic shearing (FA1; Supplementary Figure 4e-g).

(5) Ice retreat is recorded by a second periglacial weathering surface (FA7) developed on unconsolidated sediment at the top of member W1. This is overlain by fluvial channel, floodplain, lake-margin and carbonate lacustrine sediments of W2 (FA2-5; Supplementary Figure 5), recording a shifting mosaic of playa lakes and ephemeral streams. Lakes and river channels supported microbial communities. Millimetre-scale carbonate-siliciclastic rhythmites indicate seasonal cycles of photosynthesis. Overall, the environment appears to have been closely similar to that of the present-day McMurdo Dry Valleys in Antarctica, though with less extreme seasonality due to its low latitude (13).

(6) Water levels and glacier extent underwent a series of oscillations, recorded by switches between glacilacustrine diamictite (FA6-D) and fluvial, lacustrine and lake-margin sediments (FA2-5) in member W2. Sedimentation rates inferred from annual

rhythmites in member W2 suggest that each retreat phase may have lasted ~$10^4$ years.

(7) A second major ice advance marks the base of W3, with widespread deposition of subglacial tills and glacitectonism of underlying sediments. Basal tills are absent from the northernmost locality, but close proximity of glacier ice is recorded by grounding-line fans (FA6-G; Supplementary Figure 4h-i).

(8) Ice retreated while the basin remained flooded and glacigenic sediment continued to be delivered to the lake by ice rafting. Thin laminated carbonates (FA5) in W3 indicate periods of reduced glacigenic sedimentation, indicative of minor climatic fluctuations over timescales of ~$10^3$ years (Supplementary Figure 5g).

(9) A sharp contact with overlying laminated 'cap' carbonate (Supplementary Figure 2) records the transition to post-glacial conditions. At some localities, basal conglomerates provide evidence of subaerial exposure followed by marine transgression. The cap carbonate closely resembles basal Ediacaran carbonates elsewhere, and marks global deglaciation, eustatic sea-level rise and connection of the basin to the sea (1, 12, 14).

Environmental and atmospheric conditions during deposition of W2 and W3 can be further elucidated by isotopic data from carbonate-associated sulphate in lacustrine limestones (Fig. 2 and Supplementary Figure 6). These display negative to extremely negative $\Delta^{17}O$ values with consistent linear co-variation with $\delta^{34}S$, indicating mixing of pre-glacial sulphate and isotopically light sulphate formed in a $CO_2$-enriched atmosphere (15, 16). The observed values could reflect non-unique combinations of $pCO_2$, $pO_2$, $O_2$ residence time and other factors, but a box model (17) indicates $pCO_2$ was most likely ~10 to 100 mbar (1 mbar = 1000 ppmv).

These values are too high to allow formation of low-latitude ice sheets in the Neoproterozoic, but they are consistent with a late-stage Snowball Earth. For an ice-free Neoproterozoic Earth, model studies indicate mean terrestrial temperatures in the range 30-50°C for $pCO_2$ = 10 to 100 mbar (18). Formation of low-latitude ice sheets requires much lower $pCO_2$, on the order of 0.1 - 1 mbar (2, 19, 20), but once formed, high albedo ice cover can maintain low planetary temperatures despite rising $pCO_2$. This hysteresis in the relationship between $pCO_2$ and planetary temperature is a key element of Snowball Earth theory. It implies that W2 and W3 were deposited relatively late in the Marinoan, after volcanic outgassing had raised $pCO_2$ from 0.1 or 1 mbar to 10 or 100 mbar. Modelled silicate weathering and volcanic outgassing rates indicate that this would require $10^6$ to $10^7$ years (21).

The consistent co-variation of $\Delta^{17}O$ and $\delta^{34}S$ in lacustrine limestones in both W2 and W3 suggests no detectable rise in atmospheric $pCO_2$, as this would alter the slope of the mixing line (Fig. 2). This implies that the glacier oscillations recorded in~~deposition of both~~ W2 and W3 occurred ~~in~~ during a relatively short time interval ($<10^5$ years, 21)~~. We therefore infer that the glacier oscillations recorded in the Wilsonbreen Formation occurred during a relatively brief period~~ toward the end of the Marinoan. ~~, whereas the much longer period during which~~This implies a long hiatus in the geological record, while $pCO_2$ built up from the low values necessary to initiate low-latitude glaciation (0.1 ~~or~~ to 1 mbar) to those indicated by the geochemical evidence. ~~may have been characterized by cold, arid conditions represented only by t~~The basal weathering horizon may record cold arid conditions during part of this interval.

The evidence for ice-sheet advance/retreat cycles at low latitudes in a $CO_2$-enriched atmosphere motivated a series of numerical simulations to test the hypothesis that these cycles were linked to Milankovitch orbital variations. We employed asynchronous coupling of a 3D ice sheet model and an Atmospheric General Circulation Model using the continental configuration of (22). We first ran simulations with a modern orbital configuration to examine ice-sheet behaviour through a large range of $p\mathrm{CO}_2$ values from 0.1 to 100 mbar, as used in previous studies (23; Supplementary Figures 7-10). Consistently with previous results (2, 20), at low $p\mathrm{CO}_2$ (0.1 mbar), global ice volume reaches 170 x $10^6$ $km^3$ but substantial tropical land areas remain ice free due to sublimation exceeding snowfall (Supplementary Figure S10a). Ice volume remains relatively constant for $p\mathrm{CO}_2$ = 0.1 to 20 mbar (Supplementary Figure S10b), due to an increase in accumulation that compensates for higher ablation rates (Supplementary Figure 13). In contrast, above 20 mbar, ice extent in the eastern Tropics significantly decreases (Supplementary Figure 10c). At $p\mathrm{CO}_2$ = 100 mbar, most of the continental ice cover disappears except for remnants over mountain ranges (Supplementary Figure 10d).

To test the sensitivity of the tropical ice sheets to Milankovitch forcing, experiments with changing orbital parameters were initialized using the steady-state ice sheets for $p\mathrm{CO}_2$ = 20 mbar. Although obliquity has been invoked as a possible cause of Neoproterozoic glaciations (24), this mechanism remains problematical and cannot account for significant climatic oscillations at low latitudes (25, 26). We therefore focused on precession as a possible driver, and used two opposite orbital configurations favoring cold and warm summers, respectively, over the northern tropics (CSO: cold summer orbit and WSO: warm summer orbit) (Supplementary

Figure 14). Switching between these configurations causes tropical ice-sheets to advance/retreat over several hundred kilometers in 10 kyr (Supplementary Movie 1), with strong asymmetry between hemispheres (Fig. 3). Shifting from WSO to CSO causes ice retreat in the southern hemisphere, while ice sheet expansion occurs in the northern hemisphere (Supplementary Figure 14c-d). Significant ice volume changes occur between 30° N and S, but are less apparent in higher latitudes. This reflects higher ablation rates in the warmer low latitudes (Supplementary Figure 14e-h), and higher ice-sheet sensitivity to shifting patterns of melt. Larger greenhouse forcing at the end of the Snowball event implies increasing ice-sheet sensitivity to subtle insolation changes. Given a strong diurnal cycle (23), our simulations also predict a significant number of days above 0°C in the tropics (Supplementary Figure 15), consistent with geological evidence for ice rafting, liquid water in lakes and rivers, and photosynthetic microbial communities.

Our results show that geological evidence for glacial-interglacial cycles (5-7) is consistent with an enriched Snowball Earth theory. Termination of the Marinoan panglaciation was not a simple switch from icehouse to greenhouse states but was characterized by a climate transition during which glacial cycles could be forced by Milankovitch orbital variations. The geochemical evidence presented here implies that at least the upper 60-70% of the Wilsonbreen Formation was deposited in ~$10^5$ years, on the assumption that a trend in $p\mathrm{CO_2}$ would be evident over longer timescales (21). Rates of $CO_2$ build-up, however, may have significantly slowed in the later stages of Snowball Earth due to silicate weathering of exposed land surfaces, so it is possible that the oscillatory phase was more prolonged.

~~The~~ Initiation of low latitude glaciation in the Neoproterozoic requires low $pCO_2$ (0.1 - 1 mbar, 2, 19, 20), implying that the oscillatory phase was preceded by a prolonged period (~$10^6$ to $10^7$ years) during which $pCO_2$ gradually increased by volcanic outgassing (21) ~~from the low levels (0.1 - 1 mbar) required for glacial initiation~~. ~~.~~ ~~In the Wilsonbreen Formation, this period appears to be represented solely by the basal weathering horizon, consistent with a 'deep Snowball' state with low temperatures and a limited hydrological cycle.~~ This timescale is consistent with recent dating evidence for the duration of Cryogenian glaciations (27).

Additional work is needed to refine the upper and lower limits of $pCO_2$ conducive to climate and ice-sheet oscillations in Snowball Earth. Factors not included in the present model, such as supraglacial dust or areas of ice-free tropical ocean (28-30), can be expected to make the Earth system more sensitive to orbital forcing. While many details remain to be investigated, our overall conclusions remain robust.

The Neoproterozoic Snowball Earth was nuanced, varied and rich. We anticipate that detailed studies of the rock record in other parts of the world, in conjunction with numerical modeling studies, will continue to yield insight into the temporal and regional diversity of this pivotal period in Earth history.

## Methods

**Sedimentology.** Lithofacies were classified based on grain size, internal sedimentary structures and deformation structures, and bounding surfaces. Detailed stratigraphic logs were made in the field, supplemented by drawings and photographs of key

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features. Samples were taken for polishing and thin sectioning, to allow detailed examination of microstructures in the laboratory. In addition, data were collected on clast lithology, shape, surface features and fabric. Diamictites of the Wilsonbreen Formation are commonly very friable, allowing included clasts to be removed intact from the surrounding matrix, allowing measurement of both clast morphology and orientation, using methods developed for unlithified sediments. Clast morphology (shape, roundness and surface texture) was measured for samples of 50 clasts to determine transport pathways. Clast fabric analysis was performed by measuring a-axis orientations of samples of 50 clasts with a compass-clinometer, and data were summarized using the eigenvalue or orientation tensor method. Orientated samples for measurement of Anisotropy of Magnetic Susceptibility (AMS) were collected using a combination of field-drilling and block sampling. AMS was measured using an AGICO KLY-3 Kappabridge operating at 875 Hz with a 300 A/m applied field at the University of Birmingham and an AGICO MFK-1A Kappabridge operating at 976 Hz with a 200 A/m applied field at New Mexico Highlands University.

**Geochemistry.** Laboratory procedures for extracting, purifying, and measuring the triple oxygen ($\delta^{18}O$ and $\Delta^{17}O$) and sulfur ($\delta^{34}S$) isotope composition of CAS in bulk carbonates are detailed in ref 16. Briefly, fresh carbonate-bearing rock chips were crushed into fine grains and powders using mortar and pestle. Rinsing the fines with 18 MΩ water revealed little water-leachable sulphate in all the Wilsonbreen carbonates. Subsequently, ca. 10 to 30 g carbonates were slowly digested in 1-3 M HCl solutions. The solution was then centrifuged, filtered through a 0.2 μm filter, and acidified before saturated $BaCl_2$ droplets were added. $BaSO_4$ precipitates were collected after >12 hours and purified using the DDARP method (see Supporting Information). The purified $BaSO_4$ was then analyzed for three different isotope parameters: 1) $\Delta^{17}O$, by converting to $O_2$ using a $CO_2$-laser fluorination method; 2) $\delta^{18}O$, by converting to CO through a Thermal Conversion Elemental Analyzer (TCEA) at 1450 °C; and 3) $\delta^{34}S$, by converting to $SO_2$ by combustion in tin capsules in the presence of $V_2O_5$ through an Elementar Pyrocube elemental analyzer at 1050 °C. The $\Delta^{17}O$ was run in dual-inlet mode while the $\delta^{18}O$ and $\delta^{34}S$ in continuous-flow mode. Both the $\Delta^{17}O$ and $\delta^{18}O$ were run on a MAT 253 at Louisiana State University whilst the $\delta^{34}S$ was determined on an Isoprime 100 continuous flow mass spectrometer at the University of Lancaster, UK. The $\Delta^{17}O$ was calculated as $\Delta^{17}O \equiv \delta'^{17}O - 0.52\times\delta'^{18}O$ in which $\delta' \equiv 1000\ln(R_{sample}/R_{standard})$ and R is the molar ratio of $^{18}O/^{16}O$ or $^{17}O/^{16}O$. All $\delta$ values are in VSMOW and VCDT for sulphate oxygen and sulfur respectively. The analytical standard deviation (1$\sigma$) for replicate analysis associated with the $\Delta^{17}O$, $\delta^{18}O$, and $\delta^{34}S$ are ±0.05‰, ±0.5‰, and ±0.2‰, respectively. Since the CAS is heterogeneous in hand-specimen, the standard deviation is for laboratory procedures. $\delta^{34}S$ values were corrected against VCDT using within run analyses of international standard NBS-127 (assuming $\delta^{34}S$ values of +21.1 ‰). Within-run standard replication (1 SD) was <0.3 ‰. All geochemical data are included in Supplementary Table 1.

Details of the numerical modelling are provided in the Supplementary Information in the online version of the paper. Code for the GCM LMDz is freely available at: http://lmdz.lmd.jussieu.fr but the ISM GRISLI (GRenoble Ice Shelf and Land Ice model) is in limited access.

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## Author contributions

Field data were collected and analyzed by IJF, DIB, EJF, MJH, EAMcM, MSP, PMW and CTES. Geochemical analyses were conducted by HB and PMW. Model experiments were designed and conducted by GLeH, YD, CD and GR. The manuscript and figures were drafted by DIB, IJF and GLeH, with contributions from the other authors.

## Competing financial interests

The authors declare no competing financial interests.

**Figures:**

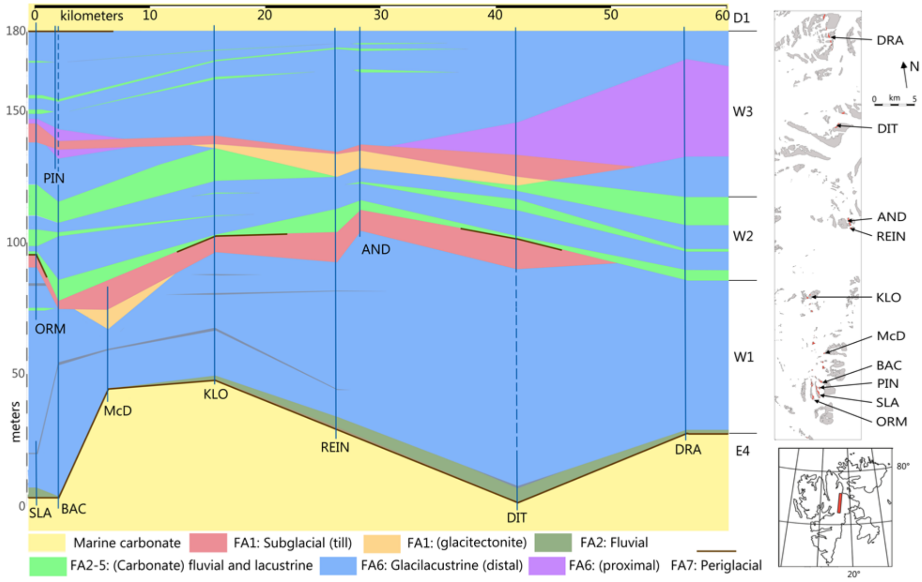


**Figure 1: Sedimentary architecture and palaeoenvironments of the Wilsonbreen Formation.** Regional correlation of facies associations and members W1, W2 and W3 across NE Svalbard. From north to south, study locations are: DRA: Dracoisen; DIT: Ditlovtoppen; AND: East Andromedafjellet; REIN: Reinsryggen (informal name); KLO: Klofjellet; McD: MacDonaldryggen; BAC: Backlundtoppen - Kvitfjellet ridge; PIN: Pinnsvinryggen (informal name); SLA: Slangen and ORM: Ormen.

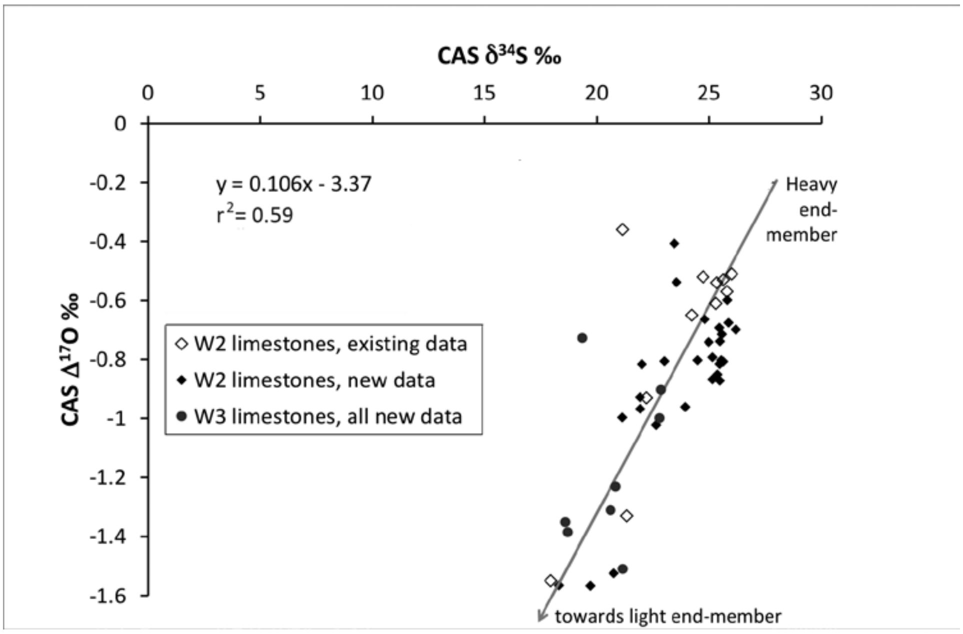


**Figure 2: Co-variation of $\Delta^{17}O$ and $\delta^{34}S$ from carbonate-associated sulphate in W2 and W3.** 'Existing data' (ref. 16) and new data define a mixing line between pre-glacial sulphate (top) and an isotopically light sulphate formed by oxidation of pyrite including incorporation of a light-$\Delta^{17}O$ signature from a $CO_2$-enriched atmosphere. Data from W2 and W3 lie on closely similar trend lines, indicating no detectable change in $pCO_2$ between deposition of the two members.

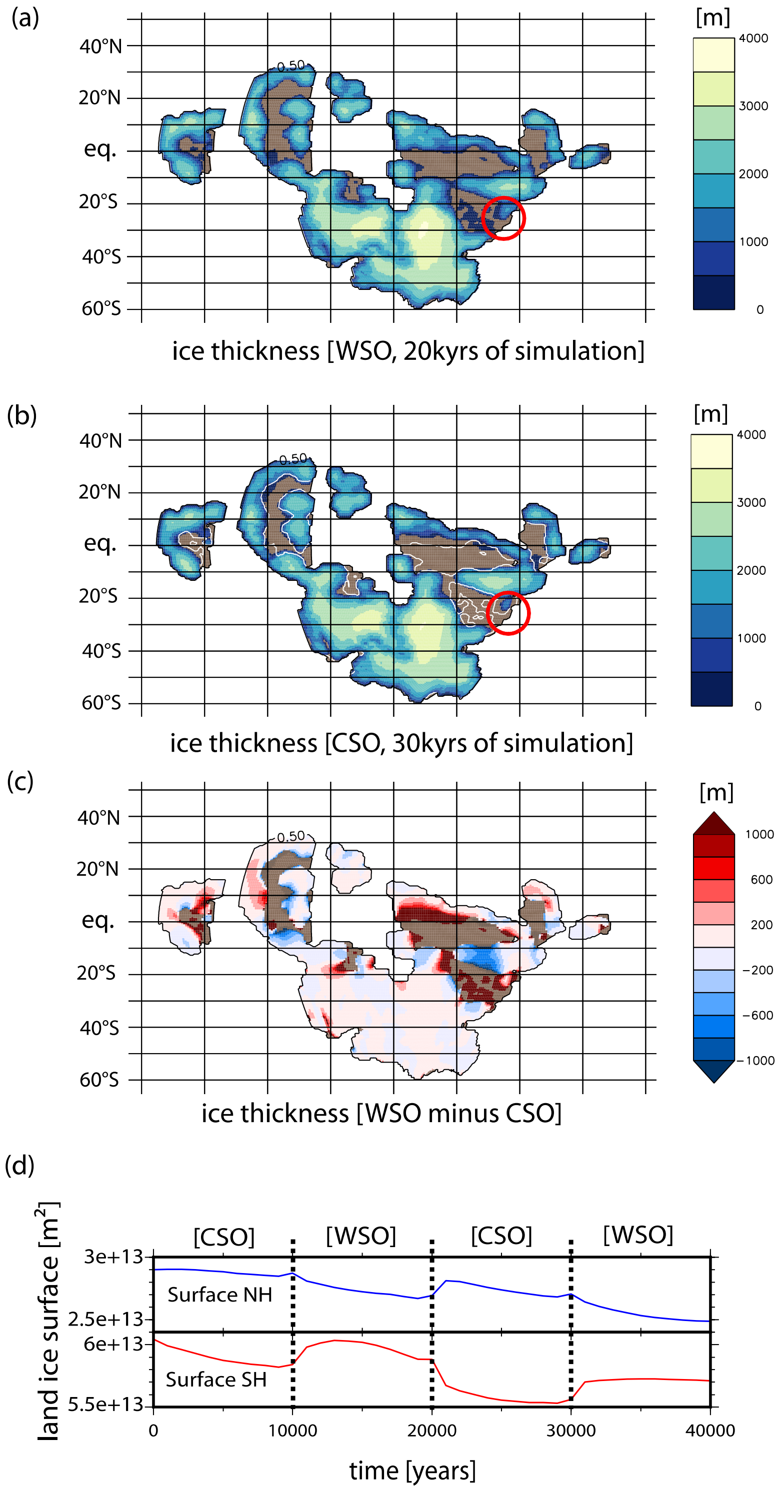


**Figure 3: Modelled ice sheet oscillations in response to orbital forcing.** (a), (b)

shaded contours show land ice thickness obtained with 20 mbar of carbon dioxide in response to changes of orbital forcing (WSO and CSO, warm/cold summer orbit for the northern hemisphere) over the course of two precession cycles (40 ky of simulation). In light brown continental areas without ice, the white line is used to represent the old ice-sheet extension (WSO case). The Svalbard area is indicated by a red circle. (c) ice thickness variation in 10 ky (WSO case after 20 ky minus CSO case after 30 ky of simulation) (d) surface of hemisphere covered by ice ($m^2$) through time ([WSO] and [CSO] indicate which orbital configuration is used).