Variation of accumulation rates over the last eight centuries on the East Antarctic Plateau derived from volcanic signals in ice cores

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Abstract. Volcanic signatures in ice-core records provide an excellent means to date the cores and obtain information about accumulation rates. From several ice cores it is thus possible to extract a spatio-temporal accumulation pattern. We show records of electrical conductivity and sulfur from 13 firn cores from the Norwegian-USA scientific traverse during the International Polar Year 2007–2009 (IPY) through East Antarctica. Major volcanic eruptions are identified and used to assess century-scale accumulation changes. The largest changes seem to occur in the most recent decades with accumulation over the period 1963–2007/08 being up to 25 % different from the long-term record. There is no clear overall trend, some sites show an increase in accumulation over the period 1963 to present while others show a decrease. Almost all of the sites above 3200 m above sea level (asl) suggest a decrease. These sites also show a significantly lower accumulation value than large-scale assessments both for the period 1963 to present and for the long-term mean at the respective drill sites. The spatial accumulation distribution is influenced mainly by elevation and distance to the ocean (continentality), as expected. Ground-penetrating radar data around the drill sites show a spatial variability within 10–20 % over several tens of kilometers, indicating that our drill sites are well representative for the area around them. Our results are important for large-scale assessments of Antarctic mass balance and model validation.

1. Introduction

The mass balance of the Antarctic ice sheet is a crucial parameter in climate research [*Alley et al.*, 2005; *Vaughan*, 2005] and is constantly under debate [*Vaughan et al.*, 1999; *Giovinetto and Zwally*, 2000; *Arthern et al.*, 2006; *van de Berg et al.*, 2006; *Horwath and Dietrich*, 2009] and a conclusive outcome is not yet reached, despite new and promising results and satellite techniques. For example, *Davis et al.* [2005] report growth of the Antarctic ice sheet over the time period 1992–2003. Recently, a study by *Velicogna* [2009] found a net mass loss over the time period 2002–2009 with an accelerating trend, based on data from the Gravity Recovery and Climate Experiment (GRACE) satellite mission. Yet interannual variations are large as are the uncertainties and there is no conclusive trend for individual drainage basins [*Horwath and Dietrich*, 2009]. *Rignot et al.* [2008] use radar interferometry and a climate model to assess recent Antarctic mass changes and obtain also a total mass loss with increases during the most recent decade. In addition to gravity missions, altimetry data give information about mass changes, derived from elevation changes. However, analyses of repeat altimetry measurements and accumulation pattern showed that observed elevation changes are largely determined by accumulation variability [*Davis et al.*, 2005], especially near the coast [*Helsen et al.*, 2008], while little is known about the impact on a continent-wide scale. Especially the East Antarctic interior is to a large degree uncovered by ground-based measurements and in situ data are scarce. *Turner et al.* [2009] review recent results of Antarctic mass balance and find that East Antarctica seems to be mostly quiescent with local exceptions. The results reported by *Turner et al.* [2009] range from zero to slightly positive values for the mass balance of East Antarctica, but again the error bars are large and errors can be as high as the variability itself. Moreover, *Turner et al.*

[2009] conclude that studies on Antarctic mass balance employing glaciological field data, e.g. *Vaughan et al.* [1999], give the most reliable results. *Genthon and Krinner* [2001] explain that especially the regions devoid of field observations introduce large errors in modeled assessments of a continent-wide accumulation pattern. Thus, it is important to obtain ground-truth for large-scale estimates of Antarctic mass changes.

The Norwegian-USA scientific IPY 2007–2009 traverse through East Antarctica aims to contribute a set of field data comprising among others firn-core records and ground-penetrating radar (GPR) data and thus help understanding the status of the East Antarctic ice sheet and its changes on scales of a few decades to more than one millennium. The traverse went from Norwegian Troll Station to South Pole in the austral summer 2007/08 and back on a different route via the Recovery Lakes area in 2008/09 (see Figure 1). We will refer to the route taken in 2007/08 as the first leg and the route from 2008/09 as the second leg in this paper. Together the two consecutive traverse legs covered large parts of the interior of Dronning Maud Land. Along the route shallow (20–30 m) and intermediate-depth (80–90 m) firn cores were drilled of which we present 13 records in total (9 shallow and 4 intermediate-depth). All the drill sites were linked by GPR data [*Müller et al.*, 2010].

Firn and ice cores are a valuable climate archive, allowing scientists to research climate variations as far back as 800000 years [*Lambert et al.*, 2008]. For the purpose of determining accumulation rates, mostly chemical species are used, often in conjunction with oxygen isotope data and electrical conductivity. Since all of these records tend to show an annual variation, they allow for identification of summer or winter peaks (depending on the species considered) and hence annual dating. However, in very low accumulation areas like the East Antarctic interior, an annual signal might not be preserved. Hence, identification of time markers is crucial in these areas for accumulation determination. Here, we focus on chemistry data (sulfur and sodium) and electrical conductivity to date the 13 firn cores by identifying known volcanic eruptions. This enables the calculation of accumulation rates and variability for the time periods between major eruptions.

2. Data and Methods

The firn cores NUS07-3, -4, -6, and -8 (Figure 1) from the first leg were analysed in the cold laboratory at Norwegian Polar Institute in Tromsø using the dielectric profiling (DEP) technique [*Moore et al.*, 1991; *Wilhelms et al.*, 1998]. From the measured capacitance and conductance we derived dielectric permittivity and electrical conductivity. The records have been presented and discussed in *Anschütz et al.* [2009] where we also give some more details about the measuring technique. The firn cores NUS07-1, -2, -5 and -7 were analysed for chemical composition (Figures 2 and 3) at the Desert Research Institute (DRI) in Reno, USA, using a sophisticated combination of continuous-flow analysis and mass spectrometry [*McConnell et al.*, 2002]. The record of NUS07-1 has also been shown by *Anschütz et al.* [2009] where sulfur, sodium and electrolytical conductivity (i.e., the conductivity of the meltwater) are discussed for this core. Note that this core is named ”site I” in *Anschütz et al.* [2009] due to a nomenclature of drill sites used during the expedition. The name has since been changed to ”NUS07-1” for the sake of consistency and we therefore also refer to this core as NUS07-1 here.

From the second leg the firn cores NUS08-2, -3, -4, and -6 were analysed using DEP (Figure 4) and cores NUS08-4 and -5 for chemistry (Figure 5). From the large amount of species measured by the device at DRI we use sulfur and sodium here. The sodium records were used to calculate non-sea-salt (nss) sulfur (see e.g. *Traufetter et al.* [2004]) which differs less than 10 % from the total sulfur at these inland sites. In the following we will refer to the nss-sulfur data as the ”sulfur records” only. The DEP and sulfur records allow for detection of volcanic peaks as shown by several studies on Antarctic and Greenland ices cores [*Hofstede et al.*, 2004; *Traufetter et al.*, 2004; *Langway et al.*, 1995; *Cole-Dai et al.*, 2000]. We follow the criterion outlined by *Cole-Dai et al.* [1997] and other authors for identification of a volcanic peak: First, the large peaks likely stemming from volcanic input were removed from the records. Second, the mean (background value) and standard deviation were calculated. For a peak to qualify as a volcanic eruption it has to fulfill two criteria: (1) the value has to be at or above two times the standard deviation and (2) has to stay at that level for at least two consecutive samples, in order to exclude outliers in the measurement. As the electrical conductivity increases with depth, we followed the method outlined by *Karlöf et al.* [2000] and other authors and normalized the DEP data by first detrending the conductivity records and then dividing by the standard deviation. Again, a peak has to be at or above two times the standard deviation for at least two samples.

In order to derive accumulation rates from the dated horizons, information about density is needed. We measured the bulk density in the field and fitted a third order polynomial to these values [*Ren et al.*, 2010] to obtain a smooth density distribution. Often the Looyenga-based density is used for accumulation calculation where DEP data are measured [*Anschütz et al.*, 2009; *Hofstede et al.*, 2004]. However, we do not have DEP data available for the chemistry cores, therefore the bulk density was used here. A comparison between Looyenga-based density and bulk density for the DEP cores yields an average difference of 3–4 %, comparable to the values reported by *Hofstede et al.* [2004].

Error estimation follows the discussion by *Anschütz et al.* [2009] and *Müller et al.* [2010]: We assume an age uncertainty of three years between volcanic horizons (discussed below in more detail) [*Traufetter et al.*, 2004; *Hofstede et al.*, 2004], a depth error of two centimeters and a relative density error of 3.5 % of the respective density values [*Hofstede et al.*, 2004]. From error propagation we derive an overall mean error of the calculated accumulation rates of 4.8 % for the time periods considered here. Errors are given as absolute values for the respective results in Table 3. The relative errors for the period 1815–2007/08 are comparable with results by *Frezzotti et al.* [2005, 2007].

A reflection horizon at the corresponding depth of the Tambora layer (1815) was identified in the GPR data based on the dating of the firn cores. In order to evaluate the areal representativity of the firn core data, the layer was followed between two firn cores (Figure 6). Uncertainties in the GPR derived layer depth and conversion to accumulation rates originate from uncertainty in firn core dating, lateral density variability between the firn cores, digitization of the GPR data, and accuracy in layer picking. We estimate the combined effect of these error sources to be up to 8 % [*Müller et al.*, 2010].

3. Results

The records of electrical conductivity and sulfur were used to identify volcanic horizons by comparison with well-dated records [*Hofstede et al.*, 2004; *Traufetter et al.*, 2004]. Yet not all peaks could be assigned to known volcanic eruptions. Here, we focus on some prominent peaks, roughly one per century, in order to detect longer-term (century-scale) accumulation changes. The volcanoes and depths of the respective DEP or sulfur peaks in the different cores are given in Tables 1 and 2.

The DEP-signal responds to both enhanced acidity due to large volcanic eruptions and enhanced sea-salt input [*Moore et al.*, 1991]. In order to distinguish between conductivity peaks from volcanic events and peaks from enhanced sea-salt content, we also looked at the sodium data for the deep chemistry core NUS07-2 from the first leg and compared sodium peaks with peaks in electrolytical conductivity. A direct comparison between electrical conductivity and sodium is not possible since we do not have DEP data for this core, therefore we use the electrolytical conductivity here. Figure 2 shows that some peaks in the electrolytical conductivity record indeed seem to coincide with enhanced sodium. However, the peaks discussed here are not linked to enhanced sea salts, at least not for this core. Furthermore Figure 2 shows that peaks in sulfur and electrolytical conductivity coincide very well, strengthening also the dating of the DEP records by comparison with the sulfur records.

The most prominent peaks served as time markers, like the double peak Tambora (Indonesia) 1815/Unknown 1809 that has been observed widely in Antarctic ice cores [*Legrand and Delmas*, 1987; *Langway et al.*, 1995; *Karlöf et al.*, 2000; *Cole-Dai et al.*, 2000; *Hofstede et al.*, 2004, among others]. Thus, we used this double peak as an absolute time marker to date the other peaks in respect to the Tambora peak. Generally, a time lag of about one year between eruption and deposition is assumed by most studies, however, deposition dates are usually less certain than eruption dates, therefore all volcanic dates mentioned in this paper are eruption dates. *Traufetter et al.* [2004] report an uncertainty in deposition dates between ±1 year and ±5 years back to AD 1200. As has been already mentioned in the error discussion, we thus assume an average age uncertainty of ±3 years here, in accordance with *Anschütz et al.* [2009] and *Hofstede et al.* [2004].

One of the more recent peaks that is observed well in Antarctic ice cores corresponds to the eruption of Agung (Lesser Sunda Islands, Indonesia, 1963) [*Delmas et al.*, 1985]. Although the signal is not very large in most of our cores, we use this as the most recent time marker. The eruption of Pinatubo (1991), which would provide an even more recent time marker, is not unambigously detected in our firn-core records. Krakatau (Indonesia) erupted in 1883 and has been detected in several ice cores around the continent [*Traufetter et al.*, 2004; *Hofstede et al.*, 2004; *Karlöf et al.*, 2000]. The unknown peak from 1695 is reported by several authors, with slightly different dates, varying from 1693–1697 [*Ren et al.*, 2010; *Hofstede et al.*, 2004; *Cole-Dai et al.*, 2000; *Budner and Cole-Dai*, 2003]. Here, we use 1695 as the eruption date in accordance with *Hofstede et al.* [2004] and *Anschütz et al.* [2009]. The subantarctic volcano of Deception Island erupted in 1641 [*Aristarain and Delmas*, 1998], however, some authors ascribe a signal at that time to the eruption of Awu (Sangihe Islands, Indonesia) [*Stenni et al.*, 2002; *Karlöf et al.*, 2000] or Mount Parker (Philippines) [*Cole-Dai et al.*, 2000; *Traufetter et al.*, 2004]. Most likely, the signal is an overlap of several eruptions. Since Deception Island is the closest one to the Antarctic continent, we attribute the 1641 peak to this volcano. Another unknown eruption occurred in 1622 [*Hofstede et al.*, 2004], and in 1600 Huaynaputina (Peru) erupted, being also visible in several ice cores [*Cole-Dai et al.*, 2000; *Karlöf et al.*, 2000; *Budner and Cole-Dai*, 2003]. Here, we use the Huaynaputina peak where it is detectable and Deception Island or Unknown 1622 for cores that do not quite reach back to 1600. Before 1600 dating is less certain due to the sparsity of historic documentation of volcanic eruptions [*Traufetter et al.*, 2004]. However, some prominent peaks have been dated in deeper ice cores and allow us to assume reliable dating for several of our observed peaks as well. The eruption of Kuwae (Vanuatu, southwest Pacific) in 1453 is easily identified in ice cores from both hemispheres [*Langway et al.*, 1995; *Oerter et al.*, 2000; *Karlöf et al.*, 2000; *Ren et al.*, 2010] and in some studies it provided the largest peak in the entire record [*Gao et al.*, 2006; *Palmer et al.*, 2001]. The eruption of El Chichon (Mexico) in 1342 is seen less often than the one of Kuwae, but some authors report prominent peaks for this eruption as well [*Budner and Cole-Dai*, 2003; *Karlöf* *et al.*, 2000; *Hofstede et al.*, 2004; *Cole-Dai et al.*, 2000]. Here, it is not as large as the Kuwae signal, but visible in all of the deeper cores.

The ”1200-sequence” of several peaks in the late 13th century is another obvious time marker. This sequence has been detected in deeper cores from the Antarctic plateau [*Hofstede et al.*, 2004; *Ren et al.*, 2010; *Cole-Dai et al.*, 2000; *Karlöf et al.*, 2000] as well as some Greenland cores [*Langway et al.*, 1995]. We picked the oldest and - in most cores - the largest one of these four peaks for our discussion. It is believed to have occurred in 1259 where some authors attribute it to El Chichon in Mexico and some prefer to call it an unknown volcano. Since there has not been a conclusive attribution to El Chichon, we stay with the term ”Unknown” here.

4. Discussion

4.1. Temporal variability

In light of sea-level change it is important to assess the mass budget of the Antarctic ice sheet and determine accumulation rates and possible spatial and temporal changes. *Anschütz et al.* [2009] discuss temporal accumulation variability for some of the sites from the first leg (NUS07-3, -4, -6 and -8). They find a decreased accumulation averaged over the time period 1815–2007 in relation to the value for 1641–1815. They also give a comprehensive discussion of temporal variability in other cores from East Antarctica. Here, we present new results from the chemistry cores of the first leg (NUS07-2, -5 and -7, Figures 2 and 3) and the DEP (Figure 4) and sulfur records (Figure 5) of most of the cores from the second leg (NUS08-2, -3, -4, -5 and -6). We identified the eruption of Agung (1963) in all of the cores but NUS07-6 which enables us to address the question of recent accumulation changes. Arguably the Agung eruption is not always very clear in the DEP profiles as they are generally more noisy than the sulfur data. However, intercomparison of the records allows for a reliable identification also in most of the

DEP cores. Where identification is somewhat questionable due to noisy data or small peaks, a question marked is depicted in the respective figures. In the chemistry cores from the first leg the Agung peak is much smaller than the very prominent earlier peaks like Tambora and Kuwae. Thus, due to the scaling of the full record the Agung peak does not depict very well and therefore we show in addition a figure of the top meters of these records where Agung is visible (Figure 3). The accumulation rates averaged over the time periods between the respective volcanic horizons are depicted in Figures 7 and 8.

All the data from the first leg exhibit a slight decrease in accumulation since 1963, with the exception of the northernmost site NUS07-1 (Figure 7). NUS07-3 shows a very slight increase over the period 1963–2007 in comparison with 1883–1963, however, this increase is within the range of uncertainty. For the majority of the sites (NUS07-2, -4, -5, -7 and -8) the accumulation between 1963–2007 is the lowest in comparison to the other time periods considered in the respective record. NUS07-6 (depicted in *Anschütz et al.* [2009]) does not show the eruption of Agung due to lower core quality in the top meters, therefore only the period 1883–2007 is considered, which again reveals the lowest accumulation in the entire record from this site (Figure 7). These results show that virtually all of the highest elevation sites (above 3200 m) reveal a decreasing trend of accumulation over the last decades. This is in accordance with the findings of *Isaksson et al.* [1999] who report accumulation values from firn cores along a traverse line from the grounded coastal area up to the Amundsenisen plateau in Dronning Maud Land. They find that accumulation has been decreasing over the time period 1965–1996 for sites above 3250 m and mostly increasing below. Hence, they conclude that an accumulation increase as reported for instance by *Mosley-Thompson et al.* [1999]; *Hofstede et al.* [2004]; *Oerter et al.* [2000] is not necessarily valid for the whole plateau area of Dronning Maud Land.

In the 17th century accumulation at the three sites NUS07-3, -4 and -6 seems to be considerably higher than during the 20th century, whereas sites NUS07-2, -5 and -7 exhibit no such changes (Figure 7). This shows that temporal accumulation changes are site-dependent and can vary significantly between sites spaced several hundreds of kilometers apart. The largest changes in the long-term records from sites NUS07-2, -5 and -7 occur largely in the most recent decades, as the accumulation rates over the period 1963–2007 are mostly lower than during the other time periods considered here. This contrasts with results from some other studies on the East Antarctic plateau that found a recent increase in accumulation, for instance *Mosley-Thompson et al.* [1999]; *Frezzotti et al.* [2005]; *Stenni et al.* [2002]; *Hofstede et al.* [2004]. However, distances between individual study sites are large and observational time periods between the studies differ, rendering it difficult to compare changes in more detail.

The sites from the second leg are all located more westerly and at lower elevations compared to the ones from the first leg and the temporal accumulation pattern is quite different. Sites NUS08-2 and -4 show a decrease and sites NUS08-3, -5 and -6 an increase over 1963–2008. At sites NUS08-3 and -6 the recent accumulation (1963–2008) is in fact the highest in the entire record for the time periods considered here (Figure 8). Sites NUS08-4 and -5 are only spaced 55 km apart, yet the temporal accumulation pattern is rather different for the recent decades. NUS08-5 shows a slow, but continuous decrease of accumulation since 1600 with the exception of the most recent period (1963–2008). NUS08-4 shows a similar decrease since 1622, but here the decrease continues also over 1963–2008.

The changes between the periods 1883–1963 and 1963–2007/08 vary between +26 % at site NUS08-3 to -22 % at site NUS07-2. When compared with the long-term record for the respective core, the changes range from +17 % to -25 % (Table 3 and Figures 7 and 8). Even though the Agung peak is not as certain in some of our DEP cores as for example the Tambora peak, the overall picture as discussed above remains valid, where accumulation seems to have mostly decreased for the sites of the first leg and mostly increased for the second leg.

Ren et al. [2004] report accumulation values from snow pits along a traverse line from Zhongshan Station to Dome A. They find that higher-elevation sites (above 3400 m) show a decrease in accumulation for the recent decades whereas sites below that elevation show an increase. This fits well with our findings from both traverse legs.

In summary, there is no consistent trend over the area of the two traverse legs and different sites show a different temporal pattern. Yet for some of the sites the most recent changes seem to be the largest. This might implicate that recent changes are in fact occurring over different parts of the East Antarctic plateau, even though the direction of changes (decreasing or increasing) does not exhibit the same trend for all sites.

As for the earlier time periods, there is no evidence of the Little Ice Age in our deeper cores: the accumulation averaged over the period 1453–1815, i.e., between the eruptions of Kuwae and Tambora, results as 32.6 kg m^{-2} a^{-1} at site NUS07-2, 25.7 kg m^{-2} a^{-1} at site NUS07-5, 29.2 kg m^{-2} a^{-1} at site NUS07-7 and for the second leg 35.5 kg m^{-2} a^{-1} at NUS08-5. All these values differ only insignificantly from the long-term accumulation rates and the values over the period 1815 to present at the respective sites (Table 3). *Li et al.* [2009] report sharpely reduced accumulation rates for the period 1450–1850 from a drill site to the east of our investigation area in Princess Elizabeth Land (core DT263 at 76°32.5'S, 77°01.5'E and 2800 m asl). A comparison with their results stresses that a different temporal accumulation pattern over different parts of the East Antarctic plateau persisted also for earlier time periods and evidence of the Little Ice Age is not necessarily found in all cores around the continent.

4.2. Spatial variability

The South-Pole Queen Maud Land Traverses (SPQMLT) went through large parts of Dronning Maud Land in the 1960s [*Picciotto et al.*, 1971] and some of their sampling sites are relatively close to our drill sites (see Figure 1). They determined accumulation rates from snow-pit stratigraphy and at selected sites additionally from measurements of radioactivity, discovering fallout from nuclear tests in the 1950s and 1960s. *Anschütz et al.* [2009] compare accumulation values from the first leg with SPQMLT data and find that accumulation in this area is lower than reported by SPQMLT. For sites close to the area visited during the second leg of the traverse, *Picciotto et al.* [1971] report an accumulation value of 38 kg m^{-2} a^{-1} for their site SPQMLT-2-12 which is 31 km from our site NUS08-5 and 33 km from NUS08-4. The value of 37.6 kg m^{-2} a^{-1} at site NUS08-5 thus is in good agreement, whereas NUS08-4 shows a slightly lower value of 36.1 kg m^{-2} a^{-1}. For their site SPQMLT-2-16, 22 km from our site NUS08-6, *Picciotto et al.* [1971] obtain 35 kg m^{-2} a^{-1}. Here, our results are higher with 49.2 kg m^{-2} a^{-1}, yet this is one of the sites where a recent accumulation increase occurs. The 200-year mean of 39.2 kg m^{-2} a^{-1} is in better agreement with the results of *Picciotto et al.* [1971]. However, one should bear in mind that comparison is limited due to large spatial distances and different time periods. Moreover, *Magand et al.* [2007] demonstrate that older data sets, like some of the SPQMLT data, are often biased and tend to overestimate accumulation on the polar plateau.

In general, the spatial representativity of point measurements such as firn-core records can be assessed by GPR data. For the first leg, *Anschütz et al.* [2009] show 5.3 GHz-GPR data around the sites NUS07-4 and -6 and find a general variability of about 10–20 % over several tens of kilometers for the Tambora layer. *Müller et al.* [2010] follow GPR layers over an 860 km long profile of the first leg and find a mean accumulation of 23.7 kg m^{-2} a^{-1} over the period 1815–2007 with a standard deviation of 4.7 kg m^{-2} a^{-1} or 20 % over the entire GPR profile.

Figure 6 shows a radargram between NUS08-5 and -6 with the Tambora layer highlighted. The system used is an ultrawideband FMCW-radar with a center frequency of 1.75 GHz and a bandwidth of 2.5 GHz. System parameters and processing steps are discussed in detail by *Müller et al.* [submitted]. The layering over some parts of this stretch is very smooth. Yet especially in the northern part (towards NUS08-6) the amplitude of layer variation is larger (Figure 6). The average accumulation over the time period 1815–2008 over this 170 km long stretch is 36.8 kg m^{-2} a^{-1} with a standard deviation of 3.6 kg m^{-2} a^{-1} or 10 %. This is on the lower edge of the values reported by *Anschütz et al.* [2009] and *Müller et al.* [2010] for parts of the first leg. Our results of spatial variability of GPR layers are in good agreement with the findings from *Richardson and Holmlund* [1999]. Even though the core sites are thus representative for the area around them, comparison between individual sites is still limited by large spatial distances and spatial variability between them. However, a general pattern is obvious, as accumulation decreases with increasing elevation and distance to the coast (continentality). This has been reported in various studies [*Vaughan et al.*, 1999; *van de Berg et al.*, 2006; *Müller et al.*, 2010; *Isaksson et al.*, 1999] and is confirmed by our results as well.

Table 3 shows accumulation values for the most recent decades, averaged over the period 1963 to present, based on the detection of the eruption of Agung. For comparison, we also give the 200-year mean values, based on the eruption of Tambora in 1815 and the respective long-term mean for the individual cores. As explained above, the Tambora eruption was used as an absolute time marker, and the 200-year mean should give a sufficiently long time interval to obtain a stable accumulation result where possible decadal variations are smoothed out. Accumulation is mostly higher for sites on the second leg than on the first. This is clearly related to elevation differences (Table 3). The accumulation over parts of the Recovery Lakes area (NUS08-4 and -5) is in the range of the higher values of the first leg. In general, accumulation is very low on the high East Antarctic plateau, for parts of the first leg even lower than expected [*Anschütz et al.*, 2009] which fits the results from some other studies as well, e.g. *Genthon et al.* [2009].

Several large-scale assessments have been carried out in order to derive a spatial pattern of accumulation for the entire Antarctic ice sheet, e.g. by *Vaughan et al.* [1999]; *Giovinetto and Zwally* [2000]; *Arthern et al.* [2006]; *Monaghan et al.* [2006]; *van de Berg et al.* [2006]. Even though a detailed comparison is limited due to the resolution of these studies (typically around 50–100 km or more), it is interesting to compare values for the area around our drill sites based on the large-scale assessments. *Anschütz et al.* [2009] discuss accumulation at sites NUS07-3, -4, -6 and -8 for the period 1815–2007 in comparison to the results by *Monaghan et al.* [2006] and *Arthern et al.* [2006]. They find lower in-situ values than these two studies. *Müller et al.* [2010] derive accumulation averaged over the time period 1815–2007 along an 860 km GPR profile for the first leg and likewise find lower values compared to the studies by *Monaghan et al.* [2006], *Arthern et al.* [2006] and *van de Berg et al.* [2006]. They conclude that this might support the suggestion that accumulation has been increasing for much of the East Antarctic plateau over the last 50 years, as the studies by *Arthern et al.* [2006] and *Monaghan et al.* [2006] represent largely this time period. This finding is not supported by our firn-core data from the first leg, highlighting again the complexity of the temporal accumulation behavior and the difficulties to draw conclusions for a large area from single drill sites.

Furthermore it is important to be aware that the values reported by *Anschütz et al.* [2009] and *Müller et al.* [2010] are point measurements and twodimensional profiles, respectively, and are averaged over a 200-year period, whereas the other studies give areal averages and look at more recent time periods of a few decades.

In Table 3 we compare our accumulation values over the period 1963 to present with the results by *Arthern et al.* [2006]. It is evident that the drill sites of the first leg show a much lower accumulation (up to 50% lower) compared to the study by *Arthern et al.* [2006], whereas the results from the second leg mostly fit well, with deviations between 2–12 %. The differences might be due to scarcity of in-situ observations available for the compilation by *Arthern et al.* [2006] as well as the reasons mentioned above, namely different time periods and resolution of this large-scale assessment. *Monaghan et al.* [2006] and *van de Berg et al.* [2006] both report values of 20–50 kg m^{-2} a^{-1} for our area of investigation with the exception of the area around South Pole where accumulation reaches 50–100 kg m^{-2} a^{-1} in both compilations. Thus, our in-situ values are largely on the lower edge or even below their assessments, especially for the sites of the first leg.

Our results show that some parts of the plateau with elevations above 3200 m exhibit less accumulation than obtained by large-scale assessments which has important implications for the determination of the overall mass balance of the Antarctic ice sheet.

5. Conclusions

In total 13 shallow and intermediate-depth firn cores from the East Antarctic plateau have been analysed for electrical conductivity and sulfur to establish a volcanic chronology and assess accumulation rates. The spatial accumulation distribution is influenced by elevation and continentality, fitting the expected pattern well. Spatial variability derived from GPR data is in the range of 10–20 % over several tens of kilometers which is in accordance with other studies from the interior of East Antarctica [*Richardson and Holmlund*, 1999; *Frezzotti et al.*, 2005]. The accumulation results for the high elevation sites above 3200 m are lower than values by the large-scale assessment of *Arthern et al.* [2006], yet the sites at lower elevations are in reasonably good agreement.

The temporal pattern does not show an overall clear trend, however, most of the sites of the first leg, i.e., the more easterly and higher elevation sites, reveal a decrease in accumulation over the period 1963–2007. For the second leg (the more westerly sites at comparatively lower elevations), there are some sites that show an increase over this time period in accordance with other results from East Antarctica [*Mosley-Thompson et al.*, 1999; *Hofstede et al.*, 2004; *Frezzotti et al.*, 2005]. The largest changes seem to have occurred in the most recent decades, with the longer-time pattern being mostly rather stable. Recent changes deviate from the long-term mean of the respective core by up to 25 %. No clear indication of the Little Ice Age could be found in our data.

Our study shows that temporal variability differs strongly between different sites, rendering difficulties to obtain a conclusive outcome for Antarctic mass changes based on individual ice-core studies. Hence, our results can serve, together with similar studies, as a valuable input for large-scale models and obtaining ground truth for satellite-based estimates of the mass balance of East Antarctica.

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Table 1. Snow depths of volcanic peaks in the cores from the first leg. All depth units are in meters and the date refers to the year of eruption as this is more certain than the year of deposition (see text).

volcano	**year**	**NUS07-1**	**NUS07-2**	**NUS07-3**	**NUS07-4**	**NUS07-5**	**NUS07-6**	**NUS07-7**	**NUS07-8**
Agung	1963	6.44	3.49	3.00	2.37	2.72	-	3.39	3.22
Krakatau	1883	14.44	10.48	7.62	6.93	7.66	5.63	9.1	9.22
Tambora	1815	20.70	15.24	10.98	10.33	11.62	8.98	13.42	13.57
Unknown	1695	-	22.96	16.98	16.03	18.12	13.76	20.37	-
Deception Island	1641	-	26.02	20.34	16.92	20.10	17.03	23.21	-
Unknown	1622	-	27.27	22.49	20.39	-	20.32	-	-
Huaynaputina	1600	-	28.96	25.33	-	22.77	-	25.29	-
Kuwae	1453	-	36.19	-	-	29.36	-	32.55	-
El Chichon	1342	-	42.29	-	-	34.72	-	36.39	-
Unknown	1259	-	46.75	-	-	38.44	-	42.01	-

Table 2. Snow depths of volcanic peaks in the cores from the second leg. All depth units are in meters and the date refers to the year of eruption.

volcano	**year**	**NUS08-2**	**NUS08-3**	**NUS08-4**	**NUS08-5**	**NUS08-6**
Agung	1963	7.19	5.51	4.92	4.76	7.33
Krakatau	1883	18.10	12.17	11.70	11.39	14.31
Tambora	1815	26.91	17.84	16.83	16.32	18.02
Unknown	1695	-	25.85	25.19	24.25	-
Deception Island	1641	-	29.27	28.43	27.61	-
Unknown	1622	-	-	29.71	28.86	-
Huaynaputina	1600	-	-	-	29.94	-
Kuwae	1453	-	-	-	38.05	-
El Chichon	1342	-	-	-	43.98	-
Unknown	1259	-	-	-	48.40	-

Table 3. Accumulation over the most recent decades, 200-year mean and long-term mean in the NUS-cores, compared with the results by *Arthern et al.* [2006]. The 200-year values for sites NUS07-1, -3, -4 and -6 have been taken from *Anschütz et al.* [2009].

core name	**lat.**	**long.**	**elevation m a.s.l.**	**acc. 1963–2007/08 kg m^{-2} a^{-1}**	**acc. 1815–2007/08 kg m^{-2} a^{-1}**	**long-term acc. kg m^{-2} a^{-1}**	**acc. from *Arthern et al.* [2006] kg m^{-2} a^{-1}**
NUS07-1	73°43' S	07°59' E	3174	55.9±3.9	52.0±2.0	-	58
NUS07-2	76°04' S	22°28' E	3582	28.0±2.0	33.0±0.7	33.3±1.2[1]	42
NUS07-3	77°00' S	26°03' E	3589	23.7±1.7	22.0±0.5	27.8±1.0[2]	40
NUS07-4	78°13' S	32°51' E	3595	17.5±1.2	19.0±0.5	20.9±0.8[3]	36
NUS07-5	78°39' S	35°38' E	3619	20.1±1.4	24.0±0.5	26.0±0.9[1]	37
NUS07-6	80°47' S	44°51' E	3672	-	16.0±0.4	21.1±0.7[2]	32
NUS07-7	82°04' S	54°53' E	3725	26.1±1.9	29.4±0.6	29.5±1.0[1]	30
NUS07-8	84°11' S	53°32' E	3452	30.0±2.1	32.0±1.2	-	40
NUS08-2	87°51' S	01°48' W	2583	63.4±4.2	67.4±2.6	-	65
NUS08-3	84°08' S	21°54' E	2625	45.3±3.1	40.1±1.0	38.8±1.4[4]	43
NUS08-4	82°49' S	18°54' E	2552	36.1±2.1	36.7±0.9	37.2±1.3[3]	34
NUS08-5	82°38' S	17°52' E	2544	37.6±2.3	35.0±0.8	35.5±0.8[1]	34
NUS08-6	81°42' S	08°34' E	2447	49.2±3.4	39.2±1.5	-	41

[1] 1259–2007/08
[2] 1600–2007/08
[3] 1622–2007/08
[4] 1641–2007/08

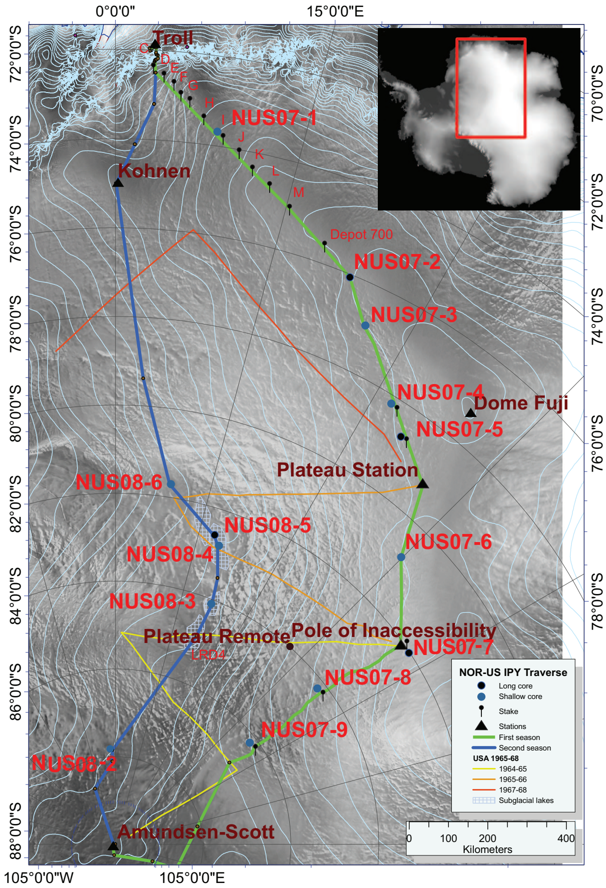

Figure 1. Map of the traverse route 2007/2008 (green line) and 2008/2009 (blue line) with drill sites from both legs indicated (NUS07-X and NUS08-X). The South Pole Queen Maud Land Traverse routes [*Picciotto et al.*, 1971] are indicated by the yellow-orange lines and relevant stations in the area of investigation are shown as well. Other dots indicate science stops along the traverse routes not relevant for this paper but shown for the sake of completeness. Elevation contour lines are in 100 m intervals. The map was compiled by K. Langley and S. Tronstad (Norwegian Polar Institute).

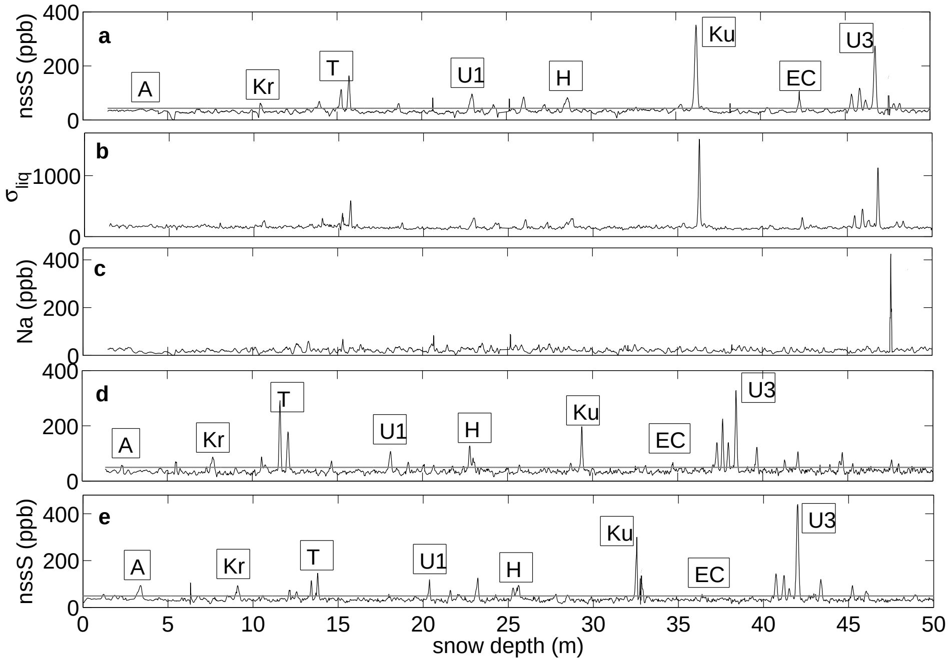

Figure 2. Records of chemistry data for the cores NUS07-2 (a: nss-sulfur, b: electrolytical conductivity, c: sodium), NUS07-5 (d: nss-sulfur) and NUS07-7 (e: nss-sulfur). The two-fold standard deviation is indicated by the grey line in the sulfur records. A: Agung 1963, Kr: Krakatau 1883, T: Tambora 1815, U1: Unknown 1695, H: Huaynaputina 1600, Ku: Kuwae 1453, EC: El Chichon 1342, U3: Unknown 1259. Note that only the top 50 m are shown here as they fully cover the time period we are concerned with here.

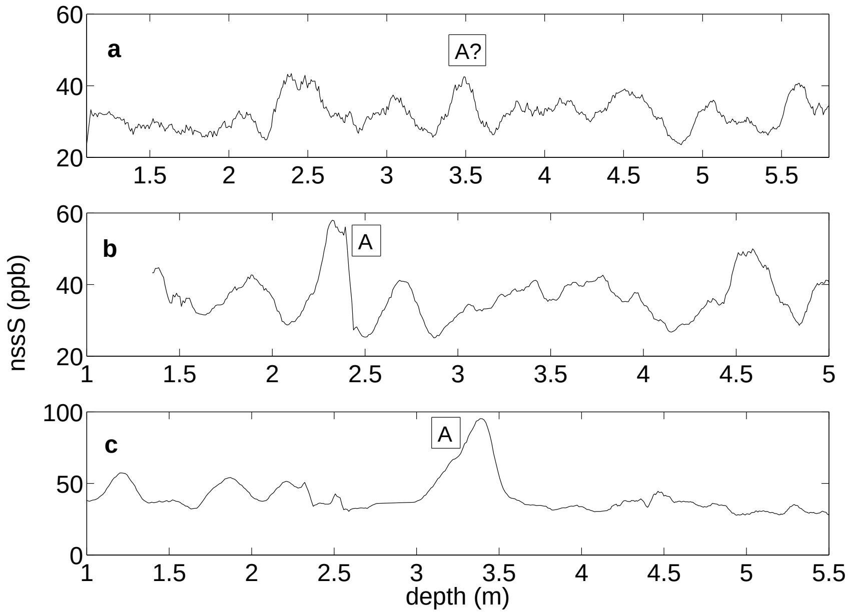

Figure 3. The Agung eruption in the deep cores from the first leg. a) NUS07-2, b) NUS07-5, c) NUS07-7. Since the peak in NUS07-2 is just at the two-fold standard deviation (see Figure 2) and also less clear than in the other cores, it is displayed with a question mark here.

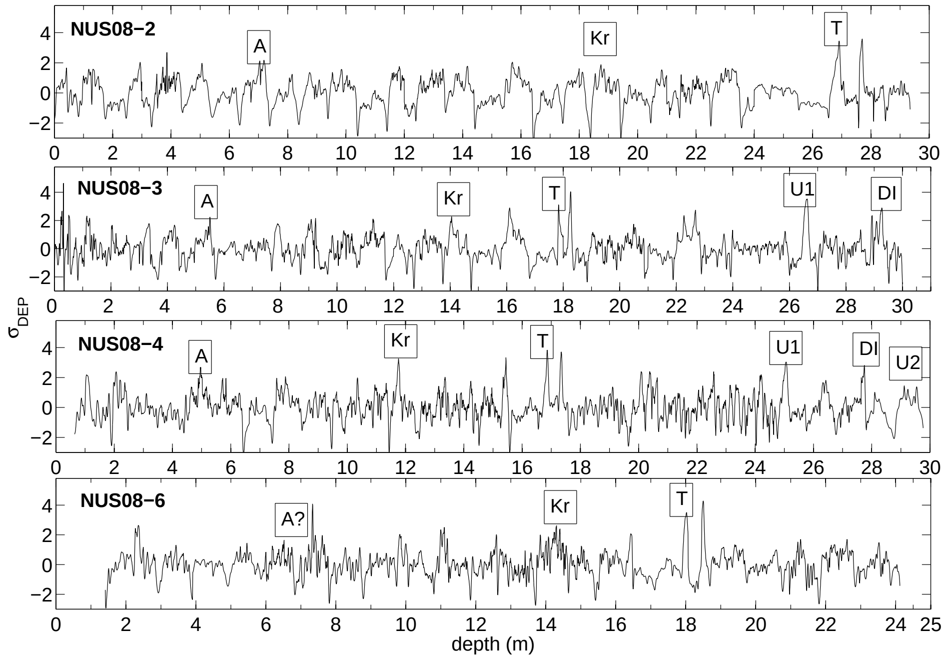

Figure 4. Normalized DEP-based conductivity for the cores NUS08-2, -3, -4 and -6 from the second leg. The volcanoes discussed in the text are indicated. DI: Deception Island 1641, U2: Unknown 1622; other abbreviations see Figure 2. The negative spikes in parts of the records are due to varying core quality and slightly differing diameter and are not eliminated here completely as full elimination would induce data gaps.

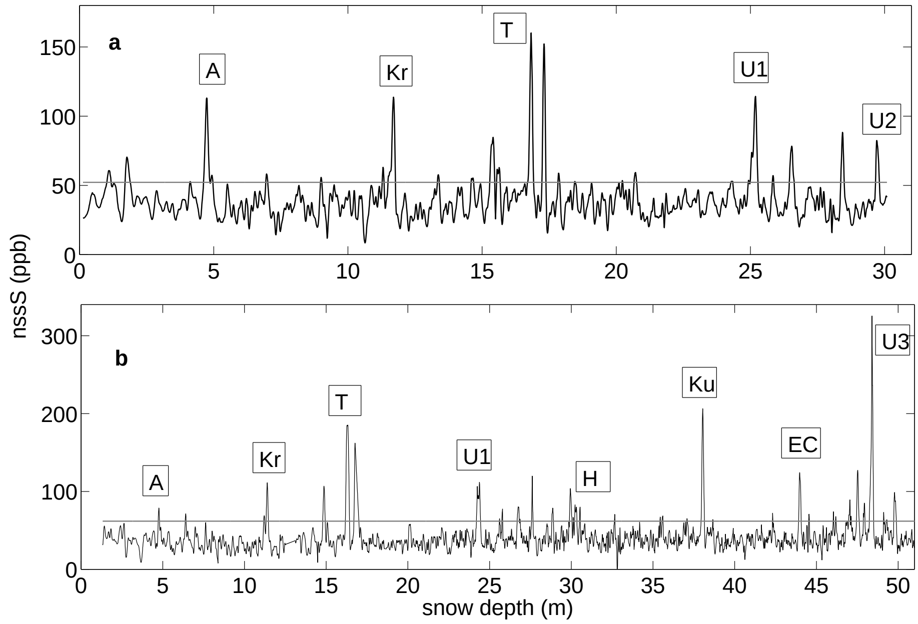

Figure 5. Sulfur data for the cores NUS08-4 (a) and NUS08-5 (b) from the second leg. The two-fold standard deviation is indicated by the grey line. Same abbreviations as in Figure 2. Note that only the top 50 m of NUS08-5 are displayed here, covering the period back to about 1250 AD that we are concerned with in this paper.

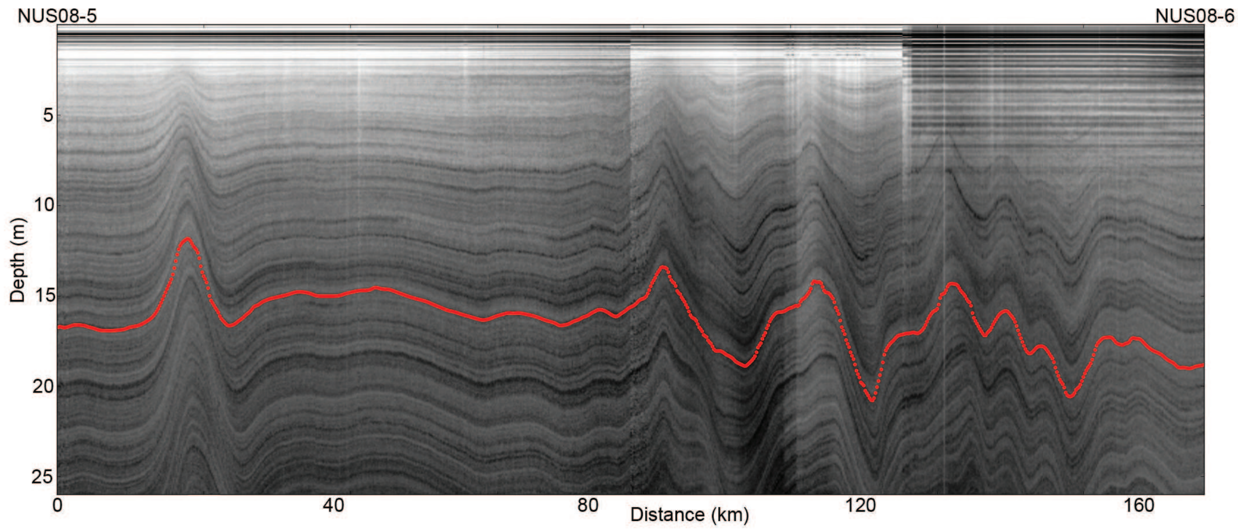

Figure 6. Radargram of the stretch between NUS08-5 and -6. The Tambora layer is highlighted by the red dashed line.

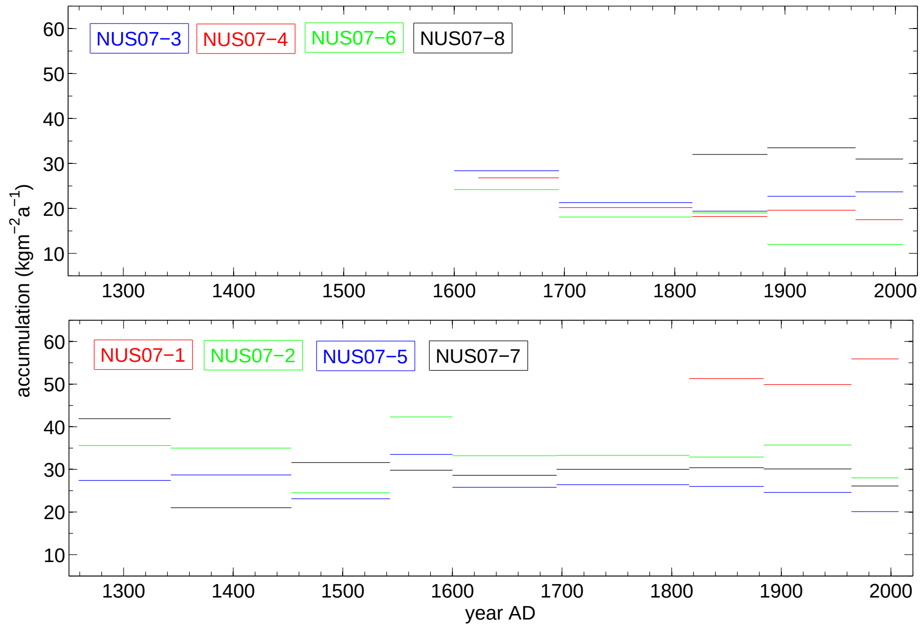

Figure 7. Temporal variability of accumulation rate in the cores from the first leg. Top: DEP cores; bottom: chemistry cores.

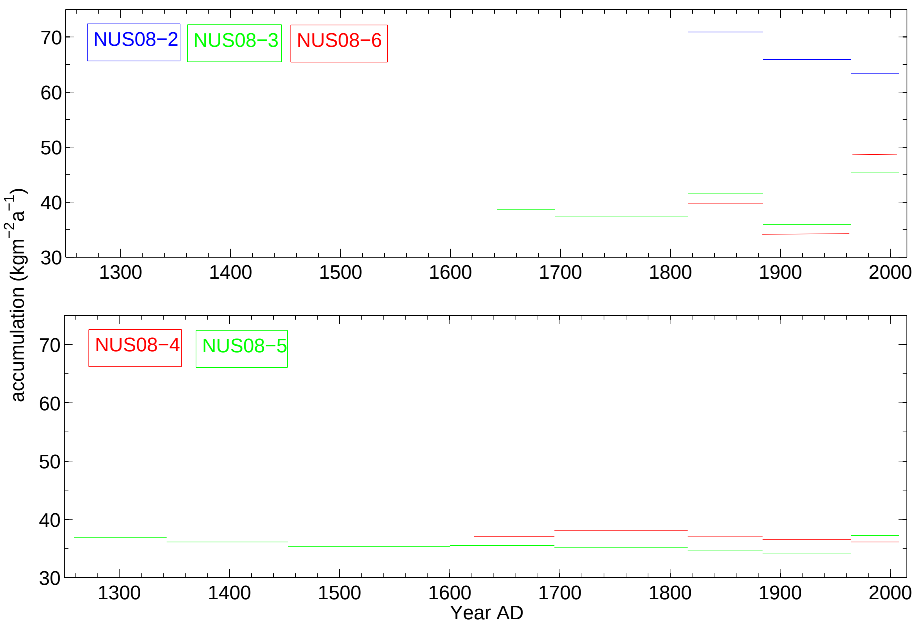

Figure 8. Temporal variability of accumulation rate in the cores from the second leg. Top: DEP cores; bottom: chemistry cores.