



Stratigraphic expressions of the Holocene–Anthropocene transition revealed in sediments from remote lakes

Alexander P. Wolfe ^{a,*}, William O. Hobbs ^b, Hilary H. Birks ^c, Jason P. Briner ^d, Sofia U. Holmgren ^e, Ólafur Ingólfsson ^f, Sujay S. Kaushal ^g, Gifford H. Miller ^h, Mark Pagani ⁱ, Jasmine E. Saros ^j, Rolf D. Vinebrooke ^k

^a Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, AB, Canada T6G 2E3

^b St. Croix Watershed Research Station, Science Museum of Minnesota, Marine on St. Croix, MN, 55105, USA

^c Department of Biology and Bjerknes Centre for Climate Research, University of Bergen, NO-5020 Bergen, Norway

^d Department of Geology, University at Buffalo, Buffalo, NY, 14260, USA

^e Department of Earth and Ecosystem Sciences, Lund University, SE-22362 Lund, Sweden

^f Department of Earth Sciences, University of Iceland, Is-101 Reykjavík, Iceland, and University Centre in Svalbard (UNIS), 9171 Longyearbyen, Norway

^g Department of Geology and Earth System Science Interdisciplinary Center, University of Maryland, College Park, MD, 20740, USA

^h Institute of Arctic and Alpine Research and Department of Geological Sciences, University of Colorado, Boulder, CO, 80309-0450, USA

ⁱ Department of Geology and Geophysics, Yale University, New Haven, CT, 06520-8109, USA

^j Climate Change Institute and School of Biology and Ecology, University of Maine, Orono, ME, 04469, USA

^k Department of Biological Sciences, University of Alberta, Edmonton, AB, Canada T6G 2E9

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ABSTRACT

Stratigraphic boundaries are ideally defined by distinct lithological, geochemical, and palaeobiological signatures, to which a chronological framework can be applied. We present a range of observations that illustrate how the Holocene–Anthropocene transition meets these criteria in its expression in sediments from remote arctic and alpine lakes, removed from direct, catchment-scale, anthropogenic influences. In glaciated lake basins, the retreat of glaciers commonly leads to lithological successions from proglacial clastic sedimentation to non-glacial organic deposition. Sediments from the majority of lakes record marked depletions in the nitrogen stable isotopic composition of sediment organic matter, reflecting anthropogenic influences on the global nitrogen cycle. In all cases, siliceous microfossil assemblages (diatoms and chrysophytes) change markedly and directionally, with regional nuances. These stratigraphic fingerprints begin to appear in the sediment record after AD 1850, but accelerate in pulses between AD 1950 and 1970 and again after AD 1980. Our review indicates that recent environmental changes associated with humankind's dominance of key global biogeochemical cycles are sufficiently pervasive to be imprinted on the sediment record of remote lakes. Moreover, these changes are of sufficient magnitude to conclude that the Holocene has effectively ended, and that the concept of Anthropocene more aptly describes current planetary dynamics. The synthesis of these observations pertains directly to ongoing discussions concerning the eventual formalization of a new stratigraphic boundary.

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* Corresponding author.

E-mail address: awolfe@ualberta.ca (A.P. Wolfe).

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1. Introduction

Although the subdivision of geological time is fundamental in earth science, new stratigraphic boundaries are relatively rare. For example, the designation of the Ediacaran at the level of Period in the terminal Proterozoic was the first interval elevated to this status in 113 years, and the first application of Phanerozoic subdivisions to deeper time (Knoll et al., 2004). The Ediacaran serves in many ways as an exemplary template, in that it is bounded by distinct lithostratigraphic, chemostratigraphic, and biostratigraphic features. Ediacaran rocks succeed Marinoan glacial sediments and conformably overlie cap carbonates that register an extreme global negative $\delta^{13}\text{C}$ excursion. The fossil content of Ediacaran-aged rocks includes enigmatic but distinctive clades (e.g., vendobionts and rangeomorphs), as well as first occurrences of putative animal embryos, bilaterian body plans, calcified fossils, and locomotive traces. These all occur well before the Cambrian diversification of shelly organisms and the widespread distribution of bioturbation. Another prominent $\delta^{13}\text{C}$ excursion conveniently demarcates termination of the Ediacaran (Knoll et al., 2004).

In the Quaternary, important boundaries have recently been ratified formally and designated their respective Global Stratotype Section and Point (GSSP). The bases of the Quaternary System/Period and of the Pleistocene Series/Epoch have been assigned a common stratotype at Monte San Nicola in Sicily (Italy), dated to 2.58 Ma, and corresponding to accelerated cooling of the climate system (Gibbard et al., 2009). The boundary between the Pleistocene and Holocene Series/Epochs has also been assigned a new stratotype, in Greenland ice (the NGRIP core, 75.1°N, 42.32°W), which dates to 11,700 yr BP and coincides with rapid warming of the northern hemisphere (Walker et al., 2009). The ice core record, which is primarily chemical in nature with some lithostratigraphic support (Holocene ice is less dusty), is bolstered by five auxiliary stratotypes, which include sediments from four lakes and one marine basin. Importantly, these exceptionally well-resolved records illustrate the breadth of sedimentological and palaeobiological change associated with the Pleistocene–Holocene transition.

These formal stratigraphic enquiries, coupled with the current acceleration of environmental changes associated with humankind (IPCC, 2007; Hansen et al., 2008; Kaufman et al., 2009), have stimulated the following question: does a boundary between the Holocene and the Anthropocene merit similar attention? We are not the first to ask this very question (Zalasiewicz et al., 2008, 2010). Instead, the objective of this review is to identify whether consistent stratigraphic markers of anthropogenic impacts can be identified in sediment records from remote lakes, thereby augmenting the available data array with which this emerging issue can be addressed. We use the term Anthropocene in the original sense of Crutzen and Stoermer (2000), Crutzen (2002), and Steffen et al. (2007), as referring to the interval of demonstrable human alteration of global biogeochemical cycles, beginning subtly in the late 18th century following James Watt's invention of the coal-fired steam engine, and accelerating markedly in the mid-20th century. A full account of the Anthropocene concept, including the temporal evolution of human processes that drive it and respond to it, is provided elsewhere (Steffen et al., 2011). While there is no question that humans have strongly

influenced the environment for millennia prior (Ruddiman, 2003), impacting both terrestrial and aquatic ecosystems (Birks, 1986; Renberg et al., 1993), our view is that these effects were for the most part local to regional in scale, although some produce geochemical signatures that are preserved in Greenland ice (Hong et al., 1994; Ferretti et al., 2005). However, the Anthropocene *sensu stricto* is more insidious than pre-industrial human activity because its consequences are unquestionably global and because the rates of several key components, including climate change associated with greenhouse gas emissions and anthropogenic emissions of reactive nitrogen (Nr), both have the potential to accelerate rapidly in the future (Galloway and Cowling, 2002; Hansen et al., 2008). Despite these realities, current suggestions for a Holocene–Anthropocene boundary focus on stratigraphic evidence that records direct and localized human modification of landscapes (Zalasiewicz et al., 2011) or of soils (Certini and Scalenghe, 2011). Here, we take an alternate approach by examining highly-resolved lake sediment records from regions removed from local human influences, thus targetting stratigraphic signatures expressed in relation to diffuse, but nonetheless discernible, anthropogenic influences. Lake sediments offer interpretable and reproducible archives of recent environmental change because they integrate efficiently the physical, chemical, and biological dimensions of the basin. This type of information is particularly useful for remote localities with fragmentary or lacking observational records (Smol, 2008).

2. Study sites and summary of methods

We present palaeolimnological evidence pertaining to the Holocene–Anthropocene transition from lakes in the Canadian and American sectors of the Rocky Mountains, Baffin Island in the eastern Canadian Arctic archipelago, west Greenland, and Spitsbergen in the Svalbard archipelago of the Norwegian High Arctic (Fig. 1). Arctic and alpine lakes share several features including nival hydrological regimes, prolonged ice cover, short growing seasons, and typically low primary production. As such, both are recognized as sentinel ecosystems with regards to anthropogenic environmental change (Smol and Douglas, 2007a; Parker et al., 2008; Williamson et al., 2009). Weather station data, augmented by palaeoclimate proxies, reveal the amplitude and pattern of 20th century warming in each region, which can be visualized alongside trends of anthropogenic Nr emission and deposition as well as the inexorable rise of atmospheric CO_2 concentrations (Fig. 2).

We synthesize data from sites that are both published and unpublished (Table 1), and thus keep methodological details to a minimum here. All sediment cores are gravity-driven, preserving an intact mud-water interface, and extruded in the field at 0.25–0.50 cm continuous increments (Glew et al., 2001). Geochronology is based on sediment excess ^{210}Pb activities measured by α -spectroscopy, to which the CRS model has been applied (Appleby, 2001). Nitrogen isotopes were measured by isotope-ratio mass spectrometry (IRMS), and are expressed as $\delta^{15}\text{N}$ relative to air ($\delta^{15}\text{N}=0\%$). Sedimentary pigments were quantified using reverse-phase high-pressure liquid chromatography (Vinebrooke and Leavitt, 1999). Diatoms were prepared and enumerated using standard protocols (Battarbee et al., 2001), whereas data manipulations, including diatom flux and



Fig. 1. Locations of the sites considered in this study. Numbers correspond to individual lakes listed in Table 1. Shaded area indicates the Rocky Mountains.

taxonomic richness calculations, follow Wolfe (2003). For the purpose of synthesizing diatom data, we used indirect ordination by detrended correspondence analysis (DCA) to summarize dominant gradients of species compositional change (Hill and Gauch, 1980). All diatom species with relative abundances >1% in any one sample were used in these analyses, with detrending by segments, down-weighting of rare taxa, and non-linear rescaling.

In this review, we present thematic and largely graphical summaries of observations pertaining to the Holocene–Anthropocene transition. In keeping with the Ediacaran analogy made initially, lithological, geochemical, and biological observations are each outlined successively, and their causative factors discussed.

3. Stratigraphic evidence

3.1. Lithostratigraphy

Although glaciers are ubiquitous features in many arctic and alpine landscapes, for the most part they have increasingly negative

mass balances at the global scale, a feature that appears largely independent of glacier size (Dyrgerov and Meier, 2000; Oerlemans, 2005; Anderson et al., 2008; Rignot et al., 2008). As glaciers retreat, their hydrological and sedimentological influences on downstream lakes become moderated. In glacial lakes across the regions considered, we have documented abrupt lithological transitions from proglacial to non-glacial sedimentation (Fig. 3). The former is typically grey silt with <10% oxidizable organic matter (measured by loss on ignition at 550 °C), while the latter comprises olive, brown, or black sapropels with >20% organic content. Laminations may occur in either, or both, facies. Two mechanisms can explain these lithostratigraphic changes. The first is the development of sediment storage zones in recently deglaciated terrains, which trap and eventually starve down-stream lakes of clastic input (Desloges, 1994). The second and more striking manifestation is the spontaneous avulsion of drainage to the lake as the glacier retreats behind a topographic threshold, after which drainage is redirected elsewhere. Both styles of sedimentological response are illustrated. In the Geraldine valley of Jasper National Park, Alberta, Canada (Fig. 4), glacier retreat has

led to progressive reductions of mineral sediment input to downstream lakes. Although remnant glacier ice below a former icefall still feeds the lakes, even the highest of these, Upper Geraldine, has now become free of suspended glacial sediment, with attendant limnological and sedimentological responses since the early 1990s (Figs. 3 and 4). At Kongressvatnet, the largest lake above the Holocene marine limit in western Spitsbergen, the retreat of a small local glacier behind a topographic threshold defined by a bench in Palaeozoic bedrock has redirected discharge from the lake northward to Grønfjorden (Fig. 5). This avulsion resulted in a dramatic cessation of clastic sedimentation. At Qipisarqo and Goose lakes in west Greenland (Fig. 3D and E), the same scenario was manifested when the inland ice retreated from its Little Ice Age maximum to positions behind bedrock knolls that define the limit of both lake catchments. Once again, sediment lithology reveals abrupt transitions from clastic to organic sedimentation, in both cases having occurred in the mid-20th century (Kaplan et al., 2002; Briner et al., 2010).

These events necessarily engender important limnological changes in the affected lakes. The reduction of allochthonous clastic inputs dramatically alters the light and ultraviolet radiation regimes of these lakes, and lake-water residence times are prolonged. Warmer summers prolong the growing season available to photoautotrophs and facilitate stratification of the water column. Furthermore, melting glaciers and perennial snowpacks have the potential to release a legacy of accumulated pollutants (Blais et al., 2001; Saros et al., 2010). It remains difficult to ascertain with precision the chronology of these events because sedimentological changes influence the efficiency of ^{210}Pb recruitment to lake sediments. Nonetheless, these changes occur within sediments that contain excess (unsupported) ^{210}Pb , and therefore post-date ~AD 1850. Sedimentological responses to glacier retreat are predicted to be time-transgressive, because they are largely controlled by local geology and geomorphology.

However, the proportion of lakes we have sampled with glaciers in their catchments is relatively small, because we typically aim to minimize the potential influences of local edaphic factors when investigating the regional consequences of climate change and atmospheric deposition. Thus, the majority of sampled lakes (Table 1) are not glacially influenced and sediments deposited in recent millennia are typically composed of uniform organic-rich sapropels. Such sediments are well-suited for stable isotopic measurements and the analysis of siliceous microfossils. These lakes form the basis of the following results and discussion.

3.2. Chemostratigraphy

3.2.1. Nitrogen stable isotopes

A recurrent observation throughout our study areas is the 20th-century decline of $\delta^{15}\text{N}$ values obtained from total sediment organic matter. This is typically in the range of 1–3‰, beginning near AD 1900, and becoming more pronounced in the latter decades of the 20th century (Wolfe et al., 2001, 2003, 2006; Holmgren et al., 2010; Holtgrieve et al., 2011). When raw sediment $\delta^{15}\text{N}$ values are normalized to mean pre-1900 values and compiled regionally as excursions from this baseline (i.e., as $\Delta\delta^{15}\text{N}$), all regions register pronounced isotopic depletions (Fig. 6). Regional trends from lake sediments are strikingly similar to the record of nitrate- $\delta^{15}\text{N}$ from ice at Summit, Greenland (72.5°N, 38.4°W), obtained using the

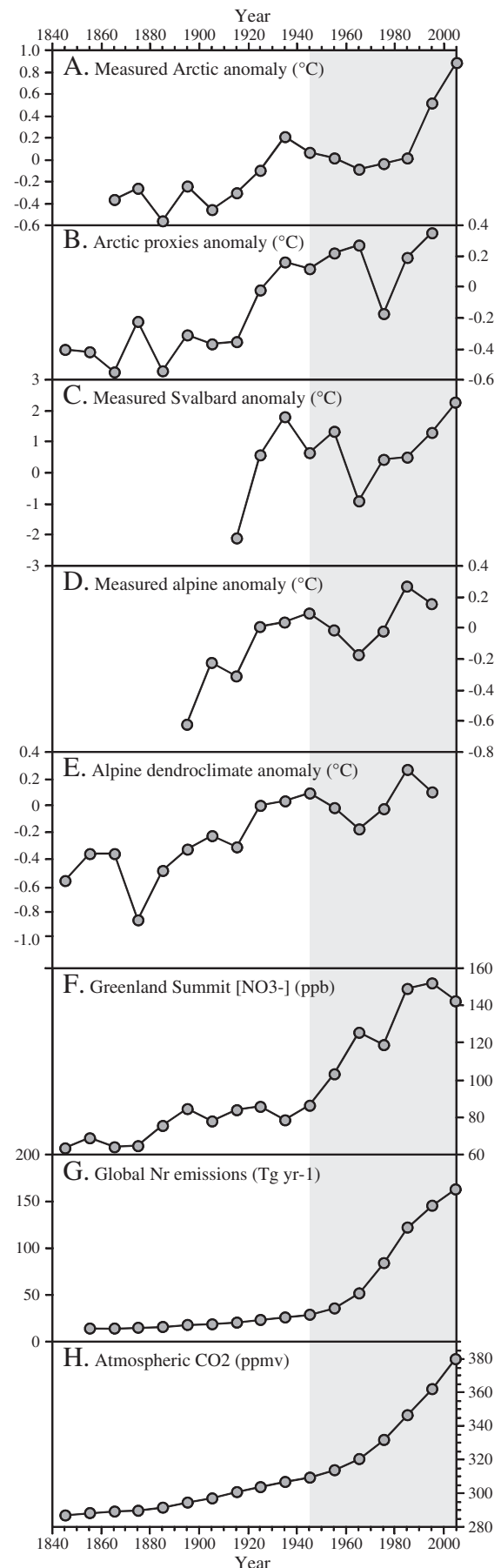


Fig. 2. Environmental context for the Holocene–Anthropocene transition. Measured and proxy-based climate records (A–E) are shown alongside Greenland summit ice-core NO_3^- concentrations (F), global Nr emissions (G), and CO_2 concentrations from spliced ice core and atmospheric measurements. In order to mimic the stratigraphic resolution attained in typical lake-sediment records, the data are reported as decadal averages. Primary sources include Kaufman et al. (2009), New et al. (2002), Isaksen et al. (2007), Luckman and Wilson (2005), Hastings et al. (2009), Galloway et al. (2004), Neftel et al. (1985) and Tans (2007). The shaded area indicates the “Great Acceleration” beginning AD 1945 (Steffen et al., 2007).

Table 1
Lakes considered in the present study.

Region	Lake name	Latitude	Longitude	Altitude (m asl)	Figure (s)	Map label	Reference
<i>Arctic</i>							
Spitsbergen, Svalbard, Norway	Borgdamene	78.06°N	13.93°E	39	6,15,16	1	This study
	Kongressvatnet	78.01°N	13.97°E	94	3,5,6,14,15,16	2	This study
	Tjornskardet	77.97°N	13.87°E	125	6,15,16	3	Holmgren et al., 2010
	Skardtjørna	77.95°N	13.81°E	67	6,8,12,13,15,16	4	Holmgren et al., 2010
Baffin Island, Nunavut, Canada	CF-11	70.47°N	68.67°W	96	6,15,16	5	Wolfe et al., 2006
	CF-10	70.43°N	69.12°W	435	6,15,16	6	Wolfe et al., 2006
	Lost Pack	69.85°N	68.95°W	241	6,15,16	7	Michelutti et al., 2005
West Greenland	Goose	69.06°N	49.90°W	277	3	8	Briner et al., 2010
	Qipisarqo	61.01°N	47.45°W	7	3	9	Kaplan et al., 2002
<i>Alpine</i>							
Jasper National Park, Alberta, Canada	Curator	52.80°N	117.87°W	2232	3,6,9,10,11,15,16	10	Hobbs et al., 2011
	Upper Geraldine	52.56°N	117.93°W	2280	3,4	11	This study
Banff National Park, Alberta, Canada	Upper Devon	51.72°N	116.24°W	2196	6,15,16	12	This study
	McConnell	51.63°N	115.97°W	2300	3,6,11,15,16	13	Hobbs et al., 2011
Beartooth Wilderness, Montana-Wyoming, U.S.A.	Beauty	44.97°N	109.57°W	2874	6,15,16	14	Saros et al., 2003
	Emerald	45.00°N	109.53°W	3292	10,15,16	15	Saros et al., 2003
	Fossil	45.00°N	110.00°W	3018	6,15,16	16	Saros et al., 2005
	Heart	44.98°N	109.54°W	3162	6,15,16	17	Saros et al., 2005
Mount Zirkel Wilderness, Colorado, U.S.A.	Pristine Lake	40.69°N	106.68°W	3366	6,7,10,11,15,16	18	This study
	Hustedt	40.58°N	105.68°W	3350	6,15,16	19	Wolfe et al., 2003
Rocky Mountain National Park, Colorado, U.S.A.	Louise	40.55°N	105.62°W	3360	6,15,16	20	Wolfe et al., 2003
	Snowdrift	40.34°N	105.73°W	3389	6,15,16	21	Wolfe et al., 2003
	Sky Pond	40.28°N	105.67°W	3322	6,7,15,16,17	22	Enders et al., 2008
	Nokoni	40.25°N	105.73°W	3292	6,15,16	23	Wolfe et al., 2003
San Juan Mountains, Colorado, U.S.A.	Big Eldorado	37.71°N	107.54°W	3851	6,15,16	24	Holtgrieve et al., 2011
	White Dome	37.71°N	107.55°W	3853	6,15,16	25	Holtgrieve et al., 2011

denitrifier method for conversion of trace NO_3^- to N_2O (Hastings et al., 2009). Although the chronology and direction of isotopic shifts are broadly parallel between lake regions and Greenland ice (Fig. 6), we note that the amplitude of change is much greater in ice (~10‰) relative to all lakes sampled (i.e., maximum 20th century negative excursion: ~5‰). Glacier ice contains a photochemical signature as well as amplified consequences of kinetic fractionation against ^{15}N during transport, deposition, and firnification. Lake-sediment $\delta^{15}\text{N}$ reflects biotic transformations associated with active Nr cycling, while potentially retaining a legacy of source N-isotopic composition (Wolfe et al., 2003; Enders et al., 2008). The timing and direction of these N-isotopic shifts, however, are remarkably similar (Fig. 6). We surmise that recent negative $\delta^{15}\text{N}$ excursions provide a robust geochemical fingerprint of transformations to the global nitrogen cycle by humans at the hemispheric scale, primarily caused by accelerated fossil fuel combustion and manufacture of agricultural fertilizer. The detailed analysis of Holtgrieve et al. (2011) reveals an early initial

inflection of isotopic ratios at the end of the 19th century (AD 1895 ± 10), followed by a marked acceleration in the second half of the 20th century. The first phase is related to Nr emissions from fossil fuel combustion only, whereas the second integrates both fossil fuel and agricultural Nr sources, in close parallel to the evolution of the “Great Acceleration” as portrayed by Steffen et al. (2007, 2011). At present, humanity fixes more N_2 to Nr than the sum of natural processes (Falkowski et al., 2000), resulting in Nr dissemination to even remote high latitudes (Fig. 2). Because the most volatile forms of anthropogenic Nr are isotopically depleted with respect to ^{15}N (Heaton, 1990; Macko and Ostrom, 1994), the isotopic composition of organic matter deposited in far-field lakes can record anthropogenic Nr inputs despite active post-depositional cycling. The $\delta^{15}\text{N}$ of NO_3^- in Svalbard precipitation illustrates how source signatures may be transmitted, given mean values of $-10.58 \pm 3.70\text{‰}$ ($n=13$; Heaton et al., 2004). Summer rain samples from western Spitsbergen reveal that discrete precipitation events scour polluted air masses, resulting

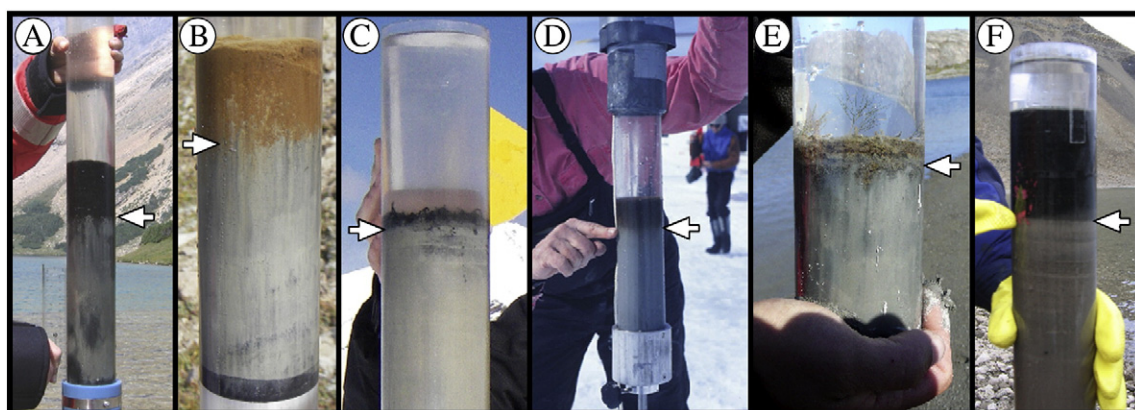


Fig. 3. Lake sediment cores illustrating abrupt lithostratigraphic transitions (marked by white arrows) from proglacial clastic sediments to non-glacial organic deposition up to the sediment–water interface, in each case associated with recent glacial retreat. McConnell (A), Curator (B) and Upper Geraldine (C) lakes are in the Canadian Rockies, Qipisarqo (D) and Goose (E) lakes are in west Greenland, and Kongressvatnet (F) is in west Spitsbergen, Svalbard. Site locations are given in Table 1.

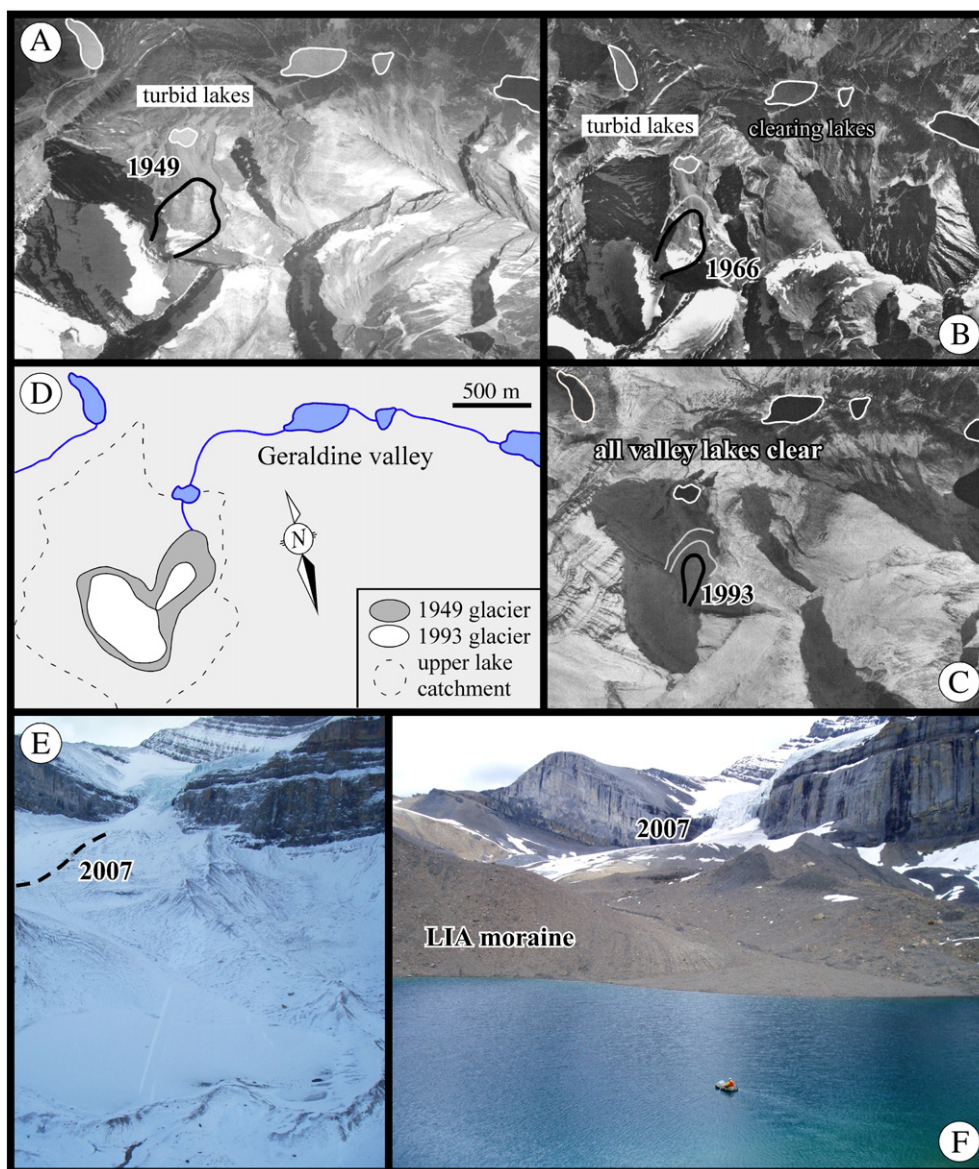


Fig. 4. Air photographs (National Airphoto Library, Ottawa) of the Geraldine lakes in Jasper National Park, Alberta, showing the loss of glacial influence on lake turbidity between AD 1949 (A), AD 1966 (B), and AD 1993 (C). In AD 1949, only the lowest was free of suspended glacial sediment, in AD 1966 only proximal lake were turbid, and by AD 1993, all lakes were clear. The corresponding glacier recession, and the Upper Geraldine Lake catchment, are mapped in (D). The Little Ice Age (LIA) moraine complex, and ice-fall above Upper Geraldine Lake in 2007 (E), viewed from helicopter positioned over the lake. Upper Geraldine Lake, clear of glacial sediment, shown looking south to the glacier terminus (F).

in episodic high-nitrate deposition, typified by $[\text{NO}_3^-] > 1 \text{ mg L}^{-1}$ (Holmgren et al., 2010). We envisage that nitrate-laden air masses originate in western Europe or Russia, and are delivered by southerly synoptic types associated with ocean circulation. In the alpine National Parks of the western U.S.A., the $\delta^{15}\text{N}$ of precipitation NO_3^- has a mean value of $-3.66 \pm 1.33\%$ ($n=7$), which is not as depleted as in Svalbard or the Greenland Ice Sheet, but nonetheless retains a pronounced anthropogenic influence (Nanus et al., 2008). Variability in contributions from ^{15}N -enriched sources such as NH_3 volatilized from animal waste are likely implicated in these regional differences, given that most of the alpine lakes considered here have airsheds where agricultural Nr is present. Transport and deposition from these sources is typically associated with episodic upslope (easterly) wind patterns generated by convective heating on the plains, a secondary circulation pattern relative to dominant Pacific (westerly) flow.

Individual lakes also exist that record the opposite trend in 20th-century sediment $\delta^{15}\text{N}$ values; that is, a progressive enrichment

in the order of $\sim 1.5\%$ (Briner et al., 2006). We predict this situation to arise in lakes that approach, but never overcome, N-limitation, resulting in a diminishing ability to fractionate strongly against ^{15}N during assimilation, and a progressive isotopic enrichment of sediment organic matter (Hodell and Schelske, 1998). As a corollary, because this pattern occurs in isolated examples (one lake on Baffin Island and another in Spitsbergen), we conclude that most lakes now receive sufficient anthropogenic Nr deposition to overcome N-limitation, allowing strong fractionation under N-luxuriant conditions. In the American Rockies, this excess Nr is sufficient to result in widespread N saturation of the aquatic environment, leading to net ecosystem export (Williams et al., 1996). Furthermore, as N limitation becomes alleviated, enhanced phosphorus (P) limitation is predicted to develop (Elser et al., 2009). As lakes undergo the transition from N- to P-limitation, strong physiological fractionation against ^{15}N can develop when Nr is replete, ultimately influencing the $\delta^{15}\text{N}$ of sediment organic matter (Goericke et al., 1994). Therefore, it is the coupled effects of isotopic fractionation during Nr assimilation and

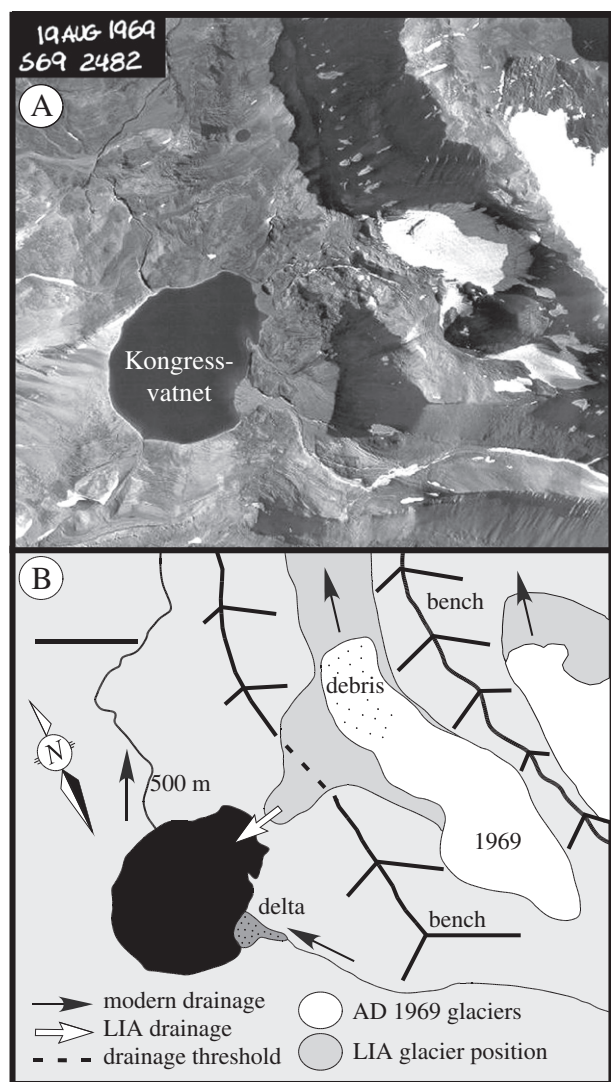


Fig. 5. Air photograph (Norsk Polarinstitutt, Longyearbyen) of Kongressvatnet, West Spitsbergen, in 1969 (A). By this time, the local glacier east of the lake had retreated behind a prominent bedrock bench, redirecting glacial drainage from the lake to the north (B). Approximately 10 cm of non-glacial sediment has accumulated in the profundal region of the lake since the avulsion of proglacial drainage (Fig. 2F).

source isotopic composition that account for the prevailing sediment $\delta^{15}\text{N}$ trend reported here, assuming that post-depositional effects on sediment isotopic composition are minimal, or at least tractable (Enders et al., 2008).

3.2.2. Assessing diagenetic effects on sediment $\delta^{15}\text{N}$

It is evident that bulk organic matter N-isotopes can be overprinted by ecosystem-scale biogeochemical processes, and this explains why ice and snow preserve a greater amplitude $\delta^{15}\text{N}$ depletion than lake sediments. Notwithstanding the results from Greenland ice (Fig. 6B), which might only be relevant to lakes in the Arctic, we have struggled with the possibility that bulk-sediment $\delta^{15}\text{N}$ measurements reflect diagenetic and not primary biogeochemical processes. Three strategies have been used to evaluate this possibility. Firstly, in the Mount Zirkel Wilderness of northern Colorado, tree-ring $\delta^{15}\text{N}$ measurements from *Picea engelmannii* (Engelmann spruce) parallel those from sediments in Pristine Lake, situated immediately above the local tree-line (Fig. 7A–B). This region receives elevated Nr deposition derived largely from down-wind coal-fired power plants in the Yampa River valley. Secondly, in Rocky Mountain

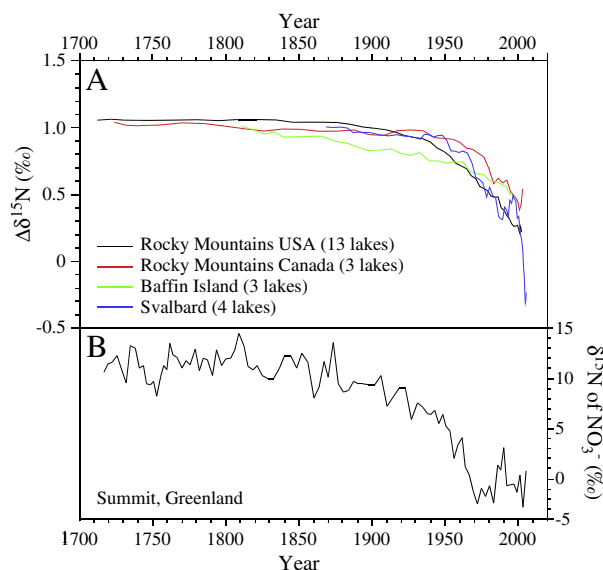


Fig. 6. Compiled nitrogen isotopic records from dated lake sediments in four geographic regions, expressed as relative departures ($\Delta\delta^{15}\text{N}$) from mean pre-AD 1900 $\delta^{15}\text{N}$ values (A). Lakes included in the compilation are listed in Table 1. The general trend observed in arctic and alpine lake sediments is mirrored by that of $\delta^{15}\text{N}$ in ice-core nitrate from the summit of Greenland (B; Hastings et al., 2009). All records show accelerated isotopic depletion after AD 1950.

National Park, sediments from Sky Pond have been studied exhaustively, including extraction, purification, and isotopic characterization of sediment algal chlorin (Enders et al., 2008). The N-isotopic composition of this primary photosynthate expresses the same trend observed in bulk sediments, but with twice the amplitude of late-20th century isotopic depletion (Fig. 7C–D). The biomarker fraction is assumed to be free of diagenetic influences. Therefore the isotopic difference between these fractions provides a first approximation of the potential magnitude of post-depositional overprinting of the bulk-sediment $\delta^{15}\text{N}$ signal. Bulk sediment $\delta^{15}\text{N}$ values are enriched up to 3‰ relative to the biomarker fraction at the mud-water interface, but this difference becomes reduced with depth in the sediment. On these grounds, we believe that regional compilations based on bulk-sediment $\delta^{15}\text{N}$ are conservative measures of actual changes in the isotopic composition of new Nr added to these ecosystems by atmospheric deposition. The most likely mechanism for the isotopic enrichment of bulk sediment is microbial denitrification under episodic sediment anoxia, although this remains to be explored fully. While we are sensitive to the potential effects of early diagenesis on lake-sediment $\delta^{15}\text{N}$ (Lehmann et al., 2002; Gälman et al., 2008), we have failed to demonstrate that these overshadow regional environmental signals that are faithfully recorded (Fig. 6).

Finally, the $\delta^{15}\text{N}$ of modern leaves of *Salix polaris* (polar willow) was measured in collections from two locations in western Spitsbergen (Adventdalen and Skardtjørna) and compared to intact Holocene subfossil leaves from several cores from the latter site (Birks, 1991; Holmgren et al., 2010). Modern collections comprised dead, brown leaves from tundra litter, thus approximating the condition of leaves that enter the lake and become part of the sediment record. Modern leaves produced $\delta^{15}\text{N} = -4.18 \pm 0.64\text{‰}$ ($n=8$) whereas Holocene fossils, ranging in age from 7000 to 4500 cal.yr BP, have an average $\delta^{15}\text{N} = -1.33 \pm 0.95\text{‰}$ ($n=6$), with no overlap of values (Fig. 8). Once again, the direction and amplitude of these isotopic differences corroborate regional lake-sediment $\delta^{15}\text{N}$ records.

Summarily, lake-sediment N isotopes record atmospheric inputs of anthropogenic Nr to these remote lake ecosystems, and therefore have considerable utility for defining the Holocene–Anthropocene transition. Because anthropogenic Nr is historically linked to the

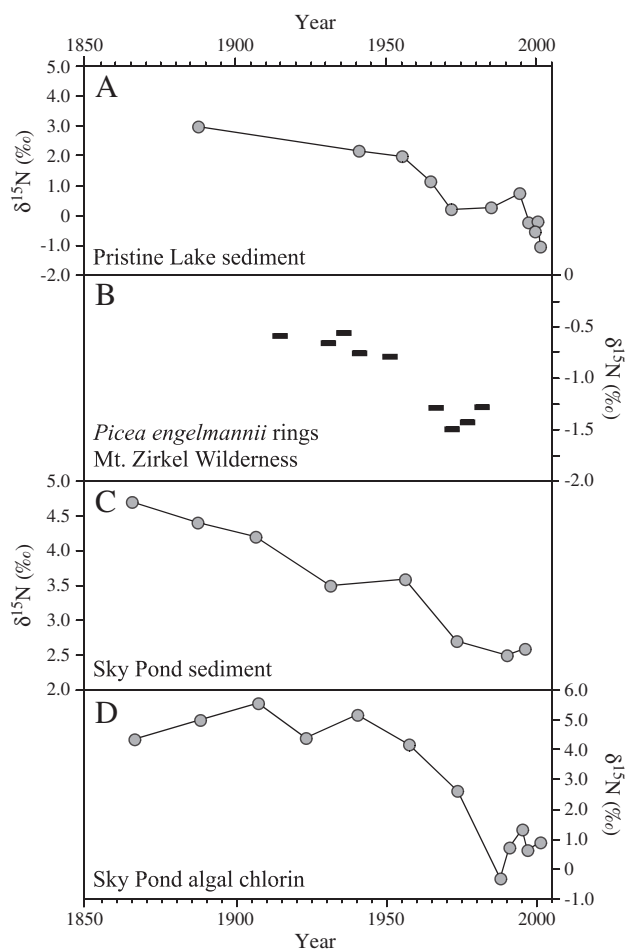


Fig. 7. Comparison of bulk-sediment $\delta^{15}\text{N}$ with independent N isotope records from two alpine lakes. In the Mount Zirkel Wilderness Area of northern Colorado, the isotopic pattern in sediments from Pristine Lake (A); 40.69°N, 106.68°W, 3366 m a.s.l.) is closely matched by that expressed in 5-year tree-ring increments from Engelmann spruce growing at slightly lower elevation (B; 40.33°N, 106.75°W, 3100 m a.s.l.). Similar trends are observed in both bulk sediments from Sky Pond, in Rocky Mountain National Park (C), as well as the purified algal chlorin fraction from the same sediments (D; Enders et al., 2008).

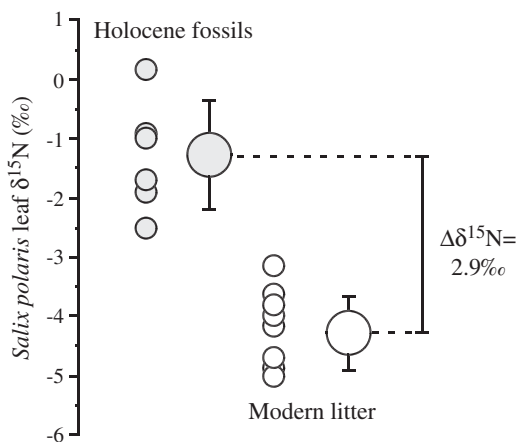


Fig. 8. Nitrogen isotopic measurements from Holocene and modern *Salix polaris* leaves from Svalbard. Holocene leaves were obtained from various pre-industrial stratigraphic levels in sediments from Skardtjørna, West Spitsbergen, whereas modern litter samples were collected from two localities in western Spitsbergen (Nordenskiöldkysten and Adventdalen).

invention of the Haber-Bosch process for industrial ammonia production (AD 1913) and subsequent accelerations of global agricultural and industrial activities (Galloway and Cowling, 2002), tracers of these events do not suffer the complication of pre-industrial anthropogenic contributions, which may influence a number of inorganic chemical markers, including the trace metals Pb and Hg (Renberg et al., 1994; Cooke et al., 2009). Moreover, the biogeochemical role of anthropogenic N_r as an essential nutrient, and the potential it has to release lakes from long-standing N-limitation, imply that N_r deposition alters ecosystem structure and function from the bottom up (Bergström et al., 2005; Elser et al., 2009).

3.3. Biostratigraphy

Siliceous microfossil assemblages (diatoms and chrysophytes) deposited over the last century reveal pervasive changes across the investigated regions, underscoring ecological reorganizations driven by the climatic and biogeochemical changes alluded to above. Diatoms are particularly useful palaeoecological indicators in lakes, owing to their good preservation potential in sediments and considerable sensitivity to a range of limnological perturbations (Stoermer and Smol, 1999). We illustrate results from lakes that are representative of the regional variability observed in our study regions, noting that many additional examples exist in the literature (Douglas et al., 1994; Smol et al., 2005; Hobbs et al., 2010).

3.3.1. Alpine lakes

In sediments from Curator Lake in Jasper National Park, the planktonic diatom *Cyclotella comensis* has increased dramatically since AD 1975, at the expense of benthic taxa, primarily small colonial Fragilariaceae (Fig. 9). Concentrations of pigments reflecting total algal biomass (chlorophyll *a*) and diatom production (the carotenoid fucoxanthin) register parallel upward inflections. Together, these proxies indicate a rapid increase of lake primary production, prolongation of the ice-free season available to planktonic diatoms, and probably enhanced thermal stratification of the water column (Hobbs et al., 2011). Directly comparable patterns have been reported in sediments from other regions (Lotter and Bigler, 2000; Rühland et al., 2008). The widespread expression of compositionally similar changes, involving increased representations of various planktonic cyclotelloid diatoms (including the genera *Cyclotella*, *Discostella*, and *Puncticulata*), appear to be directly associated with the limnological consequences of climate warming (Peeters et al., 2007; Winder et al., 2009). At Curator Lake, this biological shift is evident after AD 1975, which appears typical of lakes in the Canadian Rockies (Vinebrooke et al., 2010).

In other alpine lakes the pattern of diatom succession is more complicated, involving diatoms that appear to respond successively to climate warming and enhanced N_r availability. Emerald and Pristine lakes are situated 700 km apart in distinct ranges of the American Rocky Mountains (Table 1), yet reveal strikingly similar patterns of diatom succession (Fig. 10). In both lakes, the 20th-century rise of planktonic diatoms is first manifested by *Discostella stelligera*, but later replaced by *Asterionella formosa*, a taxon that expresses a strong positive response to available nitrogen (Saros et al., 2005). These diatoms, and examples of the taxa they have replaced, are illustrated (Fig. 11). Although diatom concentrations rise markedly in sediments deposited during the late 20th century, species richness declines precipitously in both lakes, with 60–80% fewer diatom taxa at present than prior to AD 1950. Moreover, chrysophyte cysts, produced by primarily oligotrophic taxa (Smol, 1985), undergo concurrent declines. Sediment C:N molar ratios record tell-tale declines associated with enhanced contributions from aquatic sources (Wolfe et al., 2002). Taken together, these records present a coherent picture of changes in the ecology and biogeochemistry of alpine lakes in the central American Rocky Mountains: an initial diatom response to climate

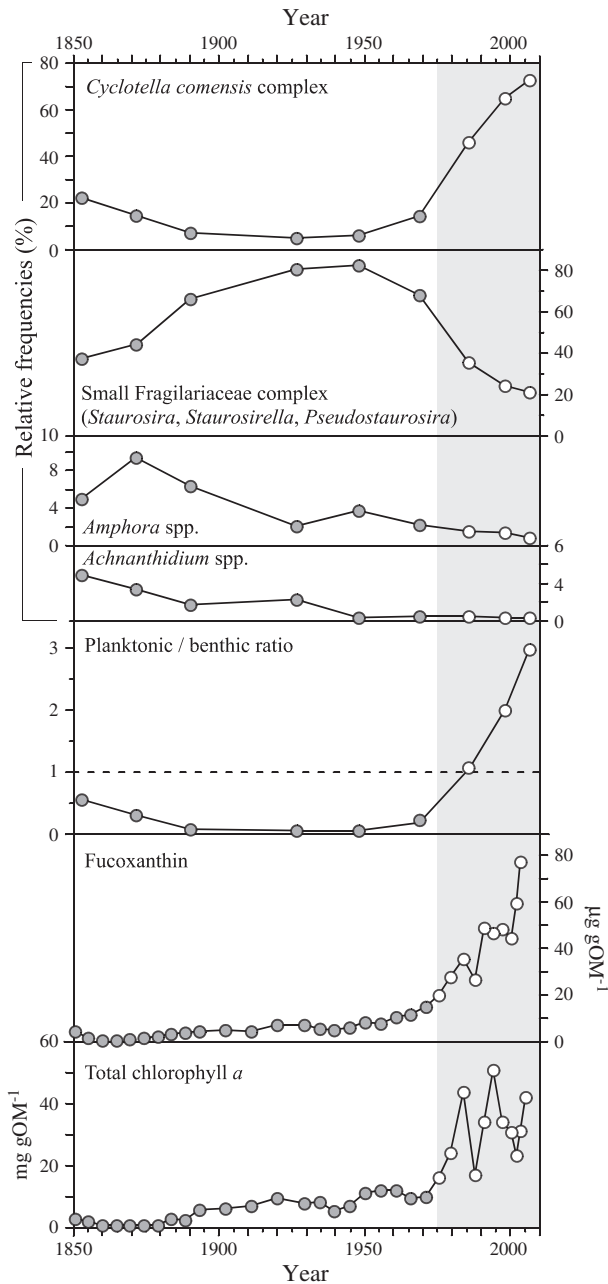


Fig. 9. Summary stratigraphy of sediment diatoms and pigments from Curator Lake (Jasper National Park, Alberta, Canada), showing striking changes since AD 1975 (shaded area) that are interpreted as, foremost, a response to recent climate warming. Pigment concentrations are normalized to sediment organic matter (i.e., mg gOM⁻¹) in order to eliminate the influence of dilution by clastic sediment constituents.

warming, initiated asynchronously during the first half of the 20th century, is followed by an interval recording the compounding influences of increased Nr availability from atmospheric deposition. Although surface-water acidification can arise from chronic Nr deposition in this region (Williams et al., 1996), to date not a single lake provides diatom evidence for recent lake acidification. It is possible that increased autochthonous organic matter production and sedimentation, coupled with warming of the water column, has enhanced processes that contribute to in-lake alkalinity, namely microbial nitrate and sulfate reduction (Schindler, 1986). Whether these lakes will remain buffered in this way indefinitely, despite widespread low-pH precipitation (Turk et al., 2001), remains an open question.

3.3.2. Arctic lakes

Recent changes in siliceous microfossils from arctic lake sediments reveal subtle differences with alpine counterparts, as discussed in considerable detail elsewhere (Douglas et al., 1994; Smol et al., 2005; Hobbs et al., 2010). The diatom stratigraphy from Skardtjørna on western Spitsbergen exemplifies the type of response commonly observed in small and shallow high arctic lakes that lack planktonic diatom floras. Here, small colonial fragilaroid genera (*Staurosira*, *Staurosirella*, and *Pseudostaurosira* spp.) that formed stable communities for millennia are replaced in the 20th century by much more diverse benthic communities of larger raphid diatoms, including stalked and tube-dwelling forms (Fig. 12). Diatom taxonomic richness has more than quadrupled in the last century and diatom fluxes have also increased markedly, in particular for the non-fragilaroid taxa (Fig. 13). The initial inflections in the Skardtjørna profile occur around AD 1920, in close association with a rapid warming step in the Svalbard temperature record (Fig. 2C). Although increased diatom production and community richness are entirely consistent with anticipated diatom responses to a warming climate, the correspondence of diatom changes to the $\delta^{15}\text{N}$ excursion implies that changing nutrient availability may also be implicated, particularly in the last two decades (Fig. 13).

The capstone example is from Kongressvatnet, a larger (0.82 km²) and deeper (>50 m) lake situated 6 km to the northeast of Skardtjørna (Fig. 5). Sediments deposited in this lake during the late 19th and 20th centuries exhibit the full range of observations pertaining to the Holocene–Anthropocene transition (Fig. 14). The lithostratigraphic boundary associated with glacier retreat occurs in sediments that contain trace activities of unsupported ²¹⁰Pb and therefore dates to the final decades of the 19th century. Within overlying black organic-rich sediments, $\delta^{15}\text{N}$ declines by 2‰ after AD 1950, while diatom assemblages change dramatically: an assemblage comprising colonial *Staurosira* and *Staurosirella* spp., occasional benthic raphid forms, and abundant chrysophyte cysts is replaced by a planktonic assemblage dominated by *Cyclotella* spp. and *Fragilaria nanana*. The compilation of available monitoring data suggest that both air and water temperatures at the site have risen by ~2 °C since 1962 (Holm et al., 2012). Together, the palaeolimnological results from Kongressvatn suggest that both climate change and nitrogen deposition have impacted the lake, inducing synergistic responses that are particularly evident within planktonic diatom taxa.

4. Timing of the Holocene–Anthropocene transition

We adopted the following strategy to address objectively the dating of the Holocene–Anthropocene transition as expressed in the lake sediment records reported here (Table 1). Due to chronological difficulties inherent to lakes with variable sediment lithologies (Fig. 3), we concentrate on the best-dated and lithologically uniform sediment $\delta^{15}\text{N}$ and diatom records. To summarize the diatom data for each lake, the leading axis of detrended correspondence analyses (DCA) was used as a synthetic gradient showing the dominant pattern of down-core assemblage turnover (Fig. 15). The first DCA axis captures between 28.8 and 80.5% of species variance (mean = $52.6 \pm 15.8\%$, $n=21$). For both the DCA axis 1 and raw $\delta^{15}\text{N}$ time-series, the slope, or first derivative, was estimated by central-point approximation (Smith, 1985). The steepest slope, indicating the highest rate of between-sample change for either proxy, was ascertained to represent the Holocene–Anthropocene transition in each of 42 records, and this point was assigned its respective ²¹⁰Pb date and associated error (Fig. 15). This approach does not reveal the initiation of changes in lake biology (diatoms) or biogeochemistry ($\delta^{15}\text{N}$), but rather highlights intervals having the greatest rates of change in either proxy. These are thus conservative, or minimum, suggested ages for the Holocene–Anthropocene boundary as expressed in these records. In general, both diatoms and $\delta^{15}\text{N}$ form bimodal

clusters of dates in this analysis, with the greatest changes in diatom assemblages occurring slightly earlier than those expressed by N isotopes (Fig. 15C). The most rapid changes occurred in two discrete intervals: AD 1950–1970 and again after AD 1980.

The individual results (^{210}Pb dates of maximum change in either proxy and corresponding dating error) can be viewed statistically as Gaussian distributions and compiled to generate summed relative probability curves (Long and Rippeteau, 1974; Lowell, 1995). These results (Fig. 16) summarize efficiently the timing of the greatest changes between proxies and across regions. In contrast to the raw data (Fig. 15C), this analysis reveals that the AD 1980–2000 interval has the highest probability of containing the greatest stratigraphic changes observed in almost all of the records considered: only alpine diatom assemblages produce higher peak relative probabilities in the earlier interval of accelerated change, AD 1950–1970 (Fig. 16). This earlier period is better expressed in alpine lakes relative to arctic counterparts, indicating that, on the whole, changes were initiated earlier in mid-latitude high-elevation ecosystems than in the Arctic, presumably due to their closer proximity to Nr emission sources. This observation is consistent with the conclusion that recent changes in alpine lakes are more closely linked to nitrogen deposition, while arctic lakes are more sensitive to climate warming (Hobbs et al., 2010).

5. Uniqueness of current ecological and biogeochemical states

An essential question that remains is how unique are the present conditions with respect to longer time-scales? For lakes where the recent record can be compared directly to older sediments for the same proxies, the notion of uniqueness in the states that characterize the Anthropocene is fully supported. The Late-glacial and Holocene diatom and $\delta^{15}\text{N}$ records from Sky Pond in the Colorado Front Range illustrate this situation (Fig. 17). Although large changes in diatoms and N stable isotopes occurred early in the lake's history during the initial stabilization of the basin following Late Pinedale glaciation, the character and direction of these were fundamentally different from changes in the late 20th century. For example, the indicator diatom *Asterionella formosa* is never present at more than trace abundances before AD 1950. Similarly, the depleted $\delta^{15}\text{N}$ values in sediments deposited in recent decades have no precedent in the lake's 14,000-year history. The DCA ordination of the compiled Sky Pond diatom data reveal the nature of these recent excursions: recent floras plot clearly outside the lake's natural ontogenetic trajectory when viewed as a biplot of the two leading axes (Fig. 17C).

The results illustrated from Sky Pond are not unique. For example, on eastern Baffin Island it has been demonstrated that 20th-century biostratigraphic changes are unique in the context of multiple late Quaternary interglaciations (Axford et al., 2009). On Ellesmere Island in the Canadian High Arctic, the recent diversification of diatom floras is also unique in the context of the Holocene (Douglas et al., 1994). Some ponds in this region now become dry in summer as a consequence of warming temperatures, a condition that again appears unprecedented and has been dubbed "crossing the final ecological threshold" (Smol and Douglas, 2007b). These results are perhaps not surprising, given glaciological evidence that both summer temperatures of the last 25 years and rates of mass loss since the 1960s are unique in the last 4200 years (Fisher et al., 2012). Melting of arctic glaciers is rapidly approaching conditions that have not been experienced since the Early Holocene thermal maximum.

With respect to even deeper time, current atmospheric CO_2 concentrations have probably not been reached since the early Pliocene

(Pagani et al., 2009), and perhaps even since the middle Miocene (Kürschner et al., 2008; Tripati et al., 2009). Future climate warming already "in the pipeline", which refers to the equilibrium climate state reached for current total greenhouse gas forcing when slow feedbacks are included, is likely of sufficient magnitude to attain global temperatures similar to those of the Pliocene (Hansen et al., 2008). Moreover, we emphasize that climate warming is but one dimension of environmental changes currently apace, implying that ultimately no single geological interval may provide an exact analog for the Anthropocene.

A second and related question is how long will the Anthropocene last? If anthropogenic CO_2 continues to be emitted according to known hydrocarbon reserves and current economic forces, and barring major geopolitical conflicts or immediate shifts to low-carbon economies, the associated climate repercussions have the potential to persist for millennia into the future (Walker and Kasting, 1992; Archer et al., 2009). Even with the most optimistic, and hence unlikely, reductions of anthropogenic CO_2 emissions involving the phase-out of coal-fired power generation by AD 2030 and other societal changes, it will be more than a century before atmospheric CO_2 concentrations stabilize to pre-industrial Holocene levels (Hansen et al., 2008). With respect to atmospheric Nr emissions, these are unlikely to decouple from human population growth, and both continue to rise exponentially (Galloway et al., 2004; Steffen et al., 2011). It therefore seems highly unlikely that the changes reported here are merely transient and reversible stratigraphic events. Rather, they foreshadow the reality that planetary dynamics in the Anthropocene have the capacity to pervade the stratigraphic record far beyond the more immediate reaches of direct human geological activities (Hooke, 2000).

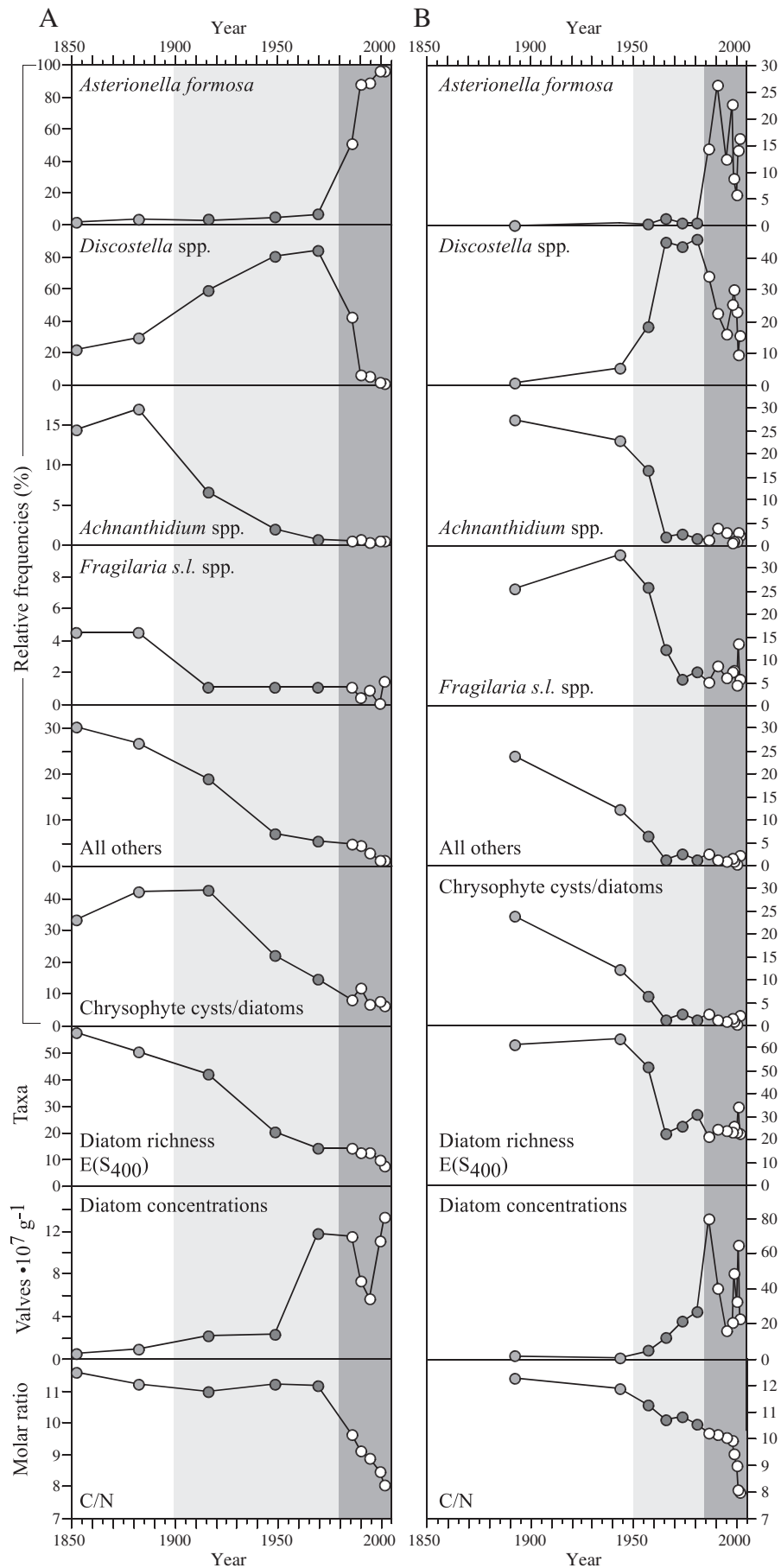
6. Synthesis and recommendations

6.1. A global expression

Collectively, these results provide defining characteristics for the Holocene–Anthropocene transition as expressed in sediments from remote arctic and alpine lakes: lithological, geochemical, and biological changes are expressed reproducibly across broad geographical expanses. Limnological processes influenced by the combined effects of climate warming and biogeochemical imbalances have conspired to produce a recognizable stratigraphic record in sediments from remote lakes of the northern hemisphere, characterized by shifts in algal biodiversity and deviations from historical patterns of nutrient limitation.

Although data coverage outside the northern hemisphere is more fragmentary, there is nonetheless compelling evidence that recent climate warming has induced pronounced limnological changes in regions such as the Peruvian Andes (Bird et al., 2011), east Antarctica (Hodgson et al., 2006), and the Antarctic Peninsula (Appleby et al., 1995; Quayle et al., 2002). Fly-ash particles, primarily sourced from coal combustion, confirm that atmospheric pollution leaves a clear record in sediments from the subantarctic islands and the coastal margin of Antarctica (Rose et al., 2012). However, glaciochemistry of the continent's interior does not support significant anthropogenic modifications to the N cycle, due to the combined influences of photochemical and stratospheric processes (Legrand and Delmas, 1986; Jones et al., 2011). This is in sharp contrast with results from Greenland (Hastings et al., 2009), in part because snow accumulation rates are significantly higher, in part because source areas are more proximal and produce greater emissions. But we do not consider Antarctica immune to anthropogenic Nr deposition, given that

Fig. 10. Siliceous microfossil stratigraphies from Emerald (A) and Pristine (B) lakes, two alpine lakes in the mid-continental American Rocky Mountains (Table 1). The rise of planktonic diatoms is first associated with *Discostella stelligera*, after which *Asterionella formosa* becomes dominant. This succession is attributed to the combined influences of climate warming and, subsequently, Nr availability from atmospheric deposition. The changes are initially time-transgressive (light shading), but become temporally more harmonized after AD 1980 (dark shading).



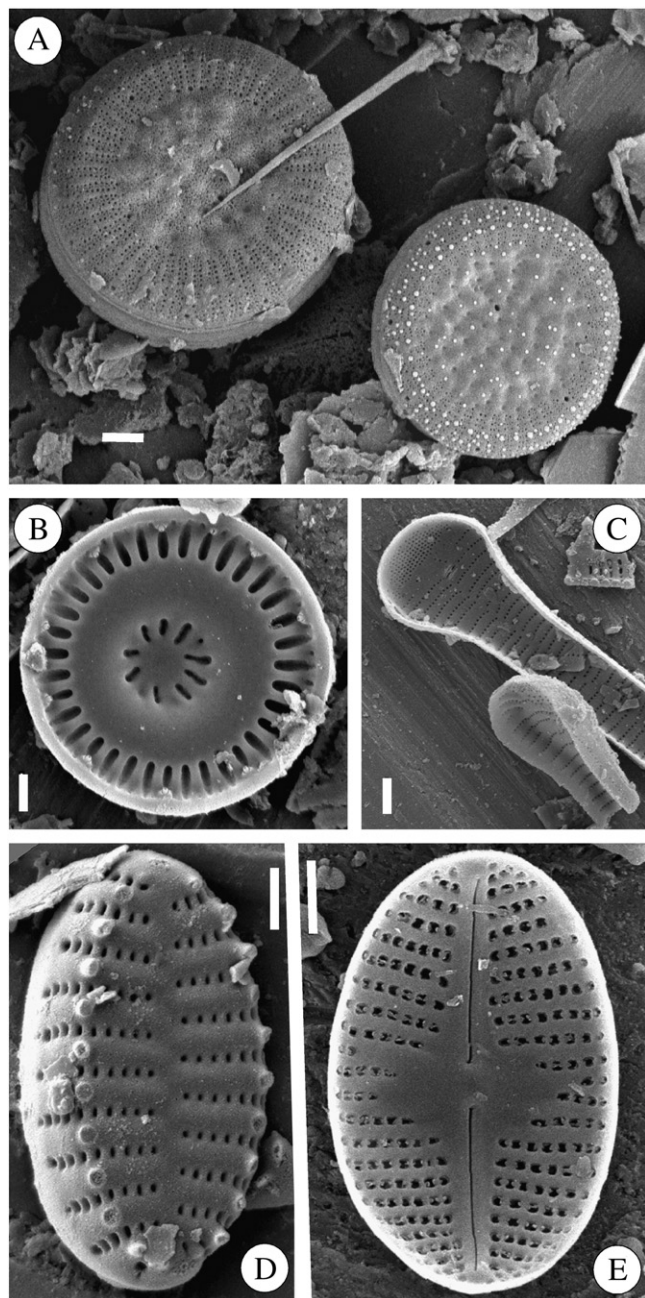


Fig. 11. Diatoms typical of the Holocene–Anthropocene transition under field-emission scanning electron microscopy. (A) *Cyclotella comensis* from Curator Lake and (B) *Discostella stelligera* from McConnell Lake, two taxa associated with the limnological consequences of climate warming. (C) *Asterionella formosa* from Pristine Lake, a taxon more closely associated with increased Nr availability. In most of the alpine lakes investigated, these planktonic diatoms have replaced small benthic forms, including (but not limited to) species of *Staurosira* (D) and *Achnanthes* (E) that were dominant throughout the late Holocene. Scale bars are 1 µm.

atmospheric vectors are demonstrably able to deliver far-field pollution (Rose et al., 2012). With economic and population growth in Australia, Argentina, and Chile, we predict that limnological effects attributable to Nr deposition will soon become discernible in the circum-antarctic regions. As far as the more populated tropical regions,

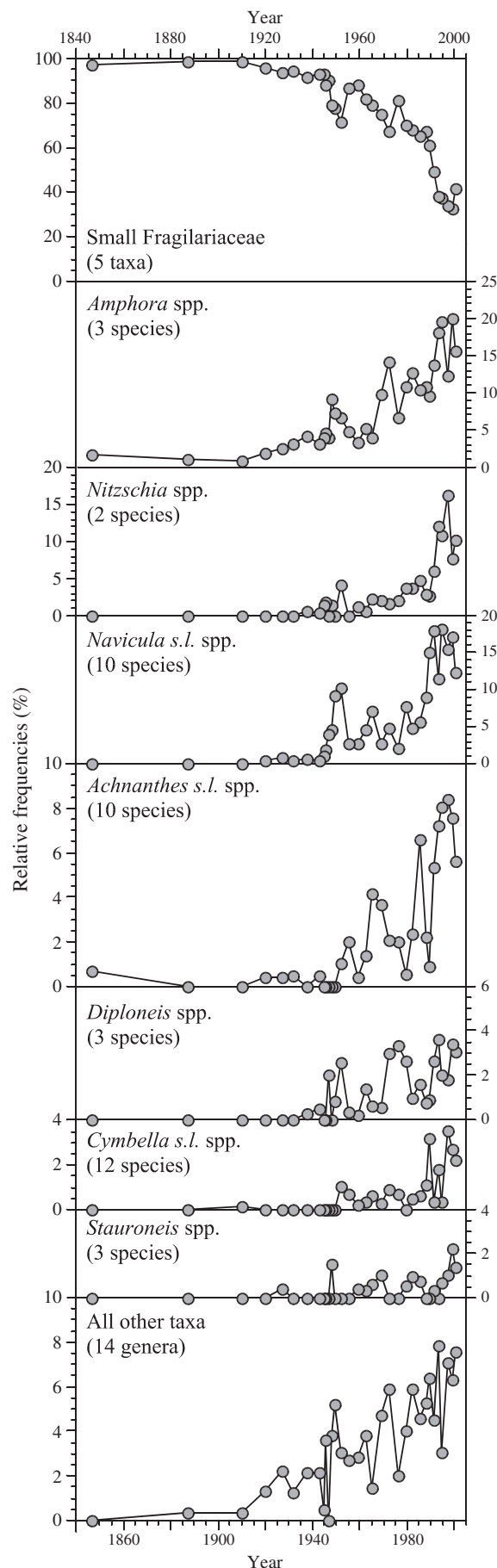


Fig. 12. Summarized diatom stratigraphy from Skardtjøna, West Spitsbergen. In the 20th century, diatom communities became increasingly diverse as sediment assemblages dominated by small colonial fragilarioid taxa were replaced by a range of larger benthic forms. This trend accelerates after ~AD 1980.

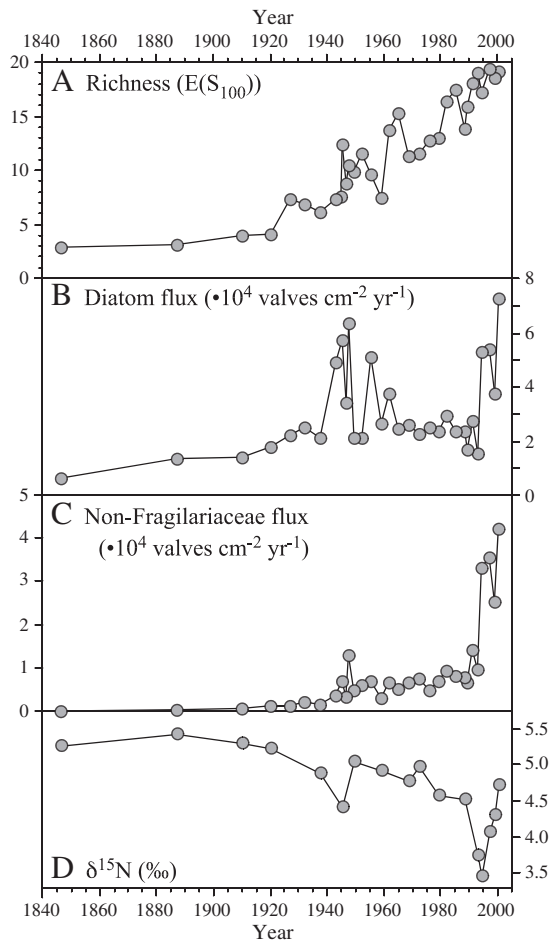


Fig. 13. Derived diatom data from Skardtjørna, including taxonomic richness estimated by rarefaction analysis ((A); Birks and Line, 1992) and diatom fluxes for total (B) as well as non-fragilarioid (C) taxa. These time-series demonstrate the diversification and greater production of diatoms in this lake, both of which accelerated in the 20th century, and especially since AD 1990. Correspondence of these shifts with the bulk-sediment $\delta^{15}\text{N}$ record (D) suggests that both climate change and Nr deposition are involved in driving these unprecedented changes.

we highlight that high-altitude areas experience amplified warming that is somewhat analogous to polar amplification (Thompson et al., 2003), and thus we predict that here too the lake record will ultimately integrate the synergy between climate warming and anthropogenic Nr emission. We encourage further scrutiny of additional archives from these and other regions, coupled to the inclusion of additional proxies, in order to more fully define the global character of sedimentary signatures attributable to the novel environmental states that define the Anthropocene.

While the oceanic realm represents a separate topic altogether (Tyrrell, 2011), it is worth noting that several organism groups with considerable fossilization potential now express range shifts that are consistent with recent warming during the Anthropocene. This includes bivalves (Berge et al., 2005), radiolarians (Björklund et al., 2012), and both benthic (Saher et al., 2012) and planktonic foraminifera (Spielhagen et al., 2011). From the results of this paper, and the observations presented above, we surmise that the Holocene–Anthropocene boundary is already well-expressed in a stratigraphic sense, and that it likely merits formal definition.

6.2. Stratigraphic considerations

The eventual formalization the Anthropocene, its designated status, and its lower boundary remain the responsibilities of the Anthropocene Working Group of the Subcommittee on Quaternary Stratigraphy, the International Commission on Stratigraphy, and ultimately the International Union of Geological Sciences. We concur with Zalasiewicz et al. (2011) that a discrete GSSP may not be necessary at this stage, largely because the community is still fleshing out the full range of physical, chemical, and biological phenomena associated with the Holocene–Anthropocene transition. The information contained in the northern-hemisphere lake-sediment records presented here, in conjunction with other syntheses (Smol et al., 2005; Hobbs et al., 2010; Holtgrieve et al., 2011), clearly advance this agenda meaningfully. The results of our chronological analysis (Fig. 16) tend to support a date for the Holocene–Anthropocene boundary around AD 1950, immediately following the onset of the “Great Acceleration” (Steffen et al., 2007). This presents the possibil-

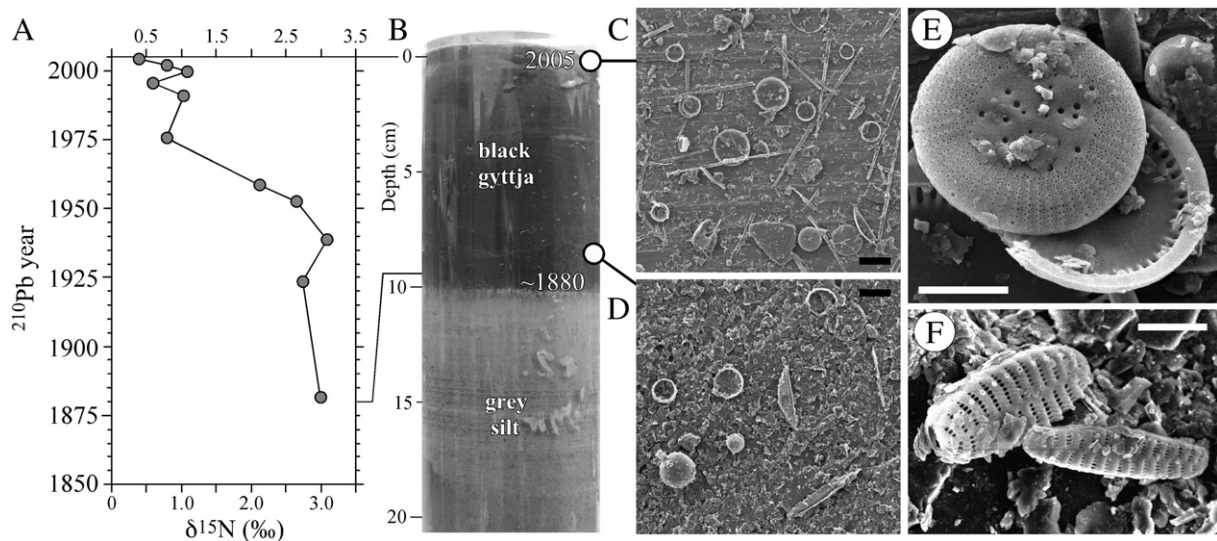
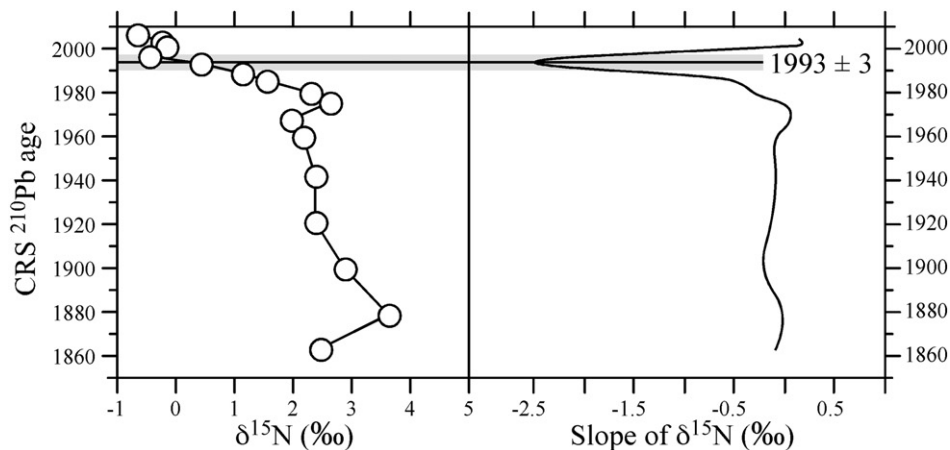
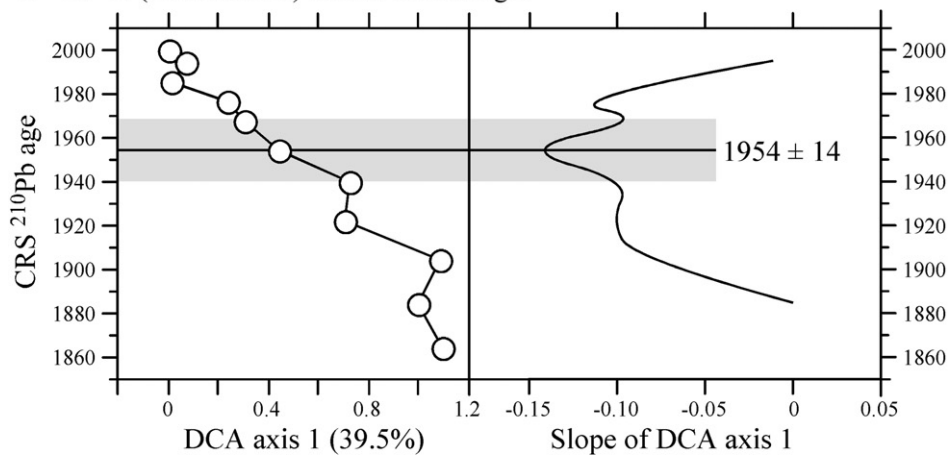


Fig. 14. Kongressvatnet, on western Spitsbergen, Svalbard, preserves all three lines of stratigraphic evidence used to characterize the Holocene–Anthropocene transition: (A) the chemical signature of nitrogen isotopic decline in the late 20th century; (B) lithological changes associated with glacier retreat; and (C)–(F) dramatic shifts in siliceous microfossil assemblages. Low magnification scanning electron micrographs are shown for the upper and lower portions of the organic facies, showing the progression from benthic forms and chrysophyte cysts (D) to dominance by centric *Cyclotella* spp. and spindle-shaped *Fragilaria nanana* (D), both of which are planktonic. Both mounts were equally diluted. High-magnification images are shown for *C. tripartita* (E) and small *Staurosira* spp. (F), which typify Anthropocene and Holocene sediments, respectively. Scale bars are 10 μm (C and D) and 5 μm (E and F).

A McConnell Lake (Banff N.P.) N isotopes



B CF-11 (Baffin Island) diatom assemblages



C Decadal compilation of intervals of greatest change

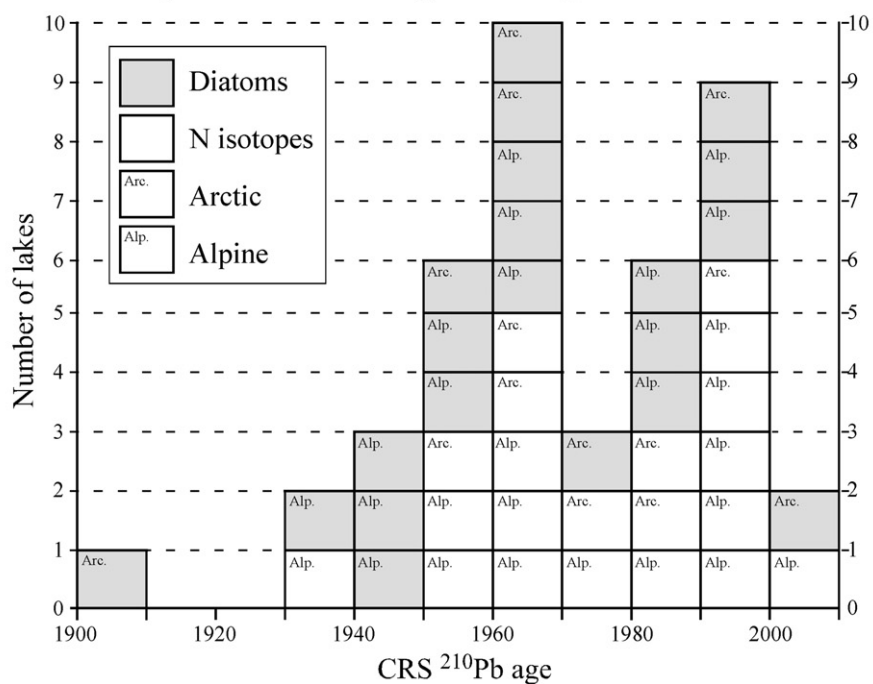


Fig. 15. Examples of the derivation of intervals of maximum stratigraphic change using central-point approximation applied to down-core nitrogen isotope values (A) and diatom DCA first-axis sample scores (B). Intervals of maximum change were obtained in this way for 21 isotope and 21 diatom records, and collated into decadal bins according to region and proxy (C).

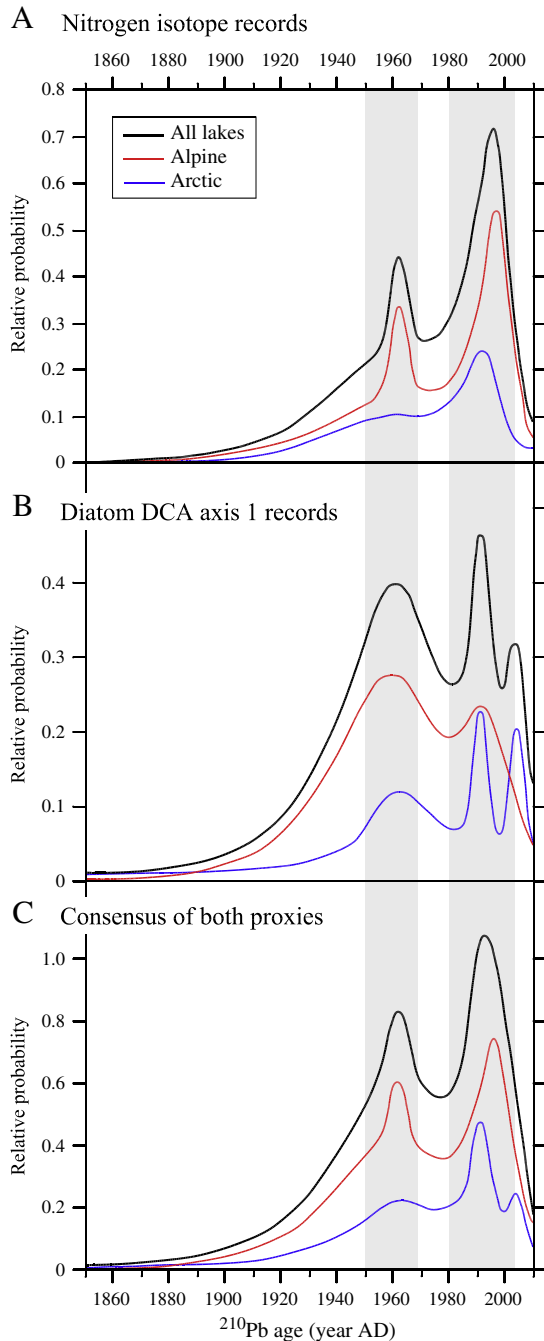


Fig. 16. Relative probabilities for the intervals of maximum chemostratigraphic (A) and biostratigraphic (B) change during the Anthropocene. The consensus of both proxies (i.e., all 42 records considered together) is shown in (C), illustrating the two temporal clusters of accelerated change (shaded areas).

ity that a Global Standard Stratigraphic Age (GSSA) at or near this time, rather than GSSP, may provide the most inclusive and hence parsimonious designation of the Holocene–Anthropocene boundary. Although this postdates the initial upward inflection of atmospheric $[\text{CO}_2]$ from fossil fuel combustion by more than a century, such an age does have several advantages. First, this is the interval during which anthropogenic greenhouse gas forcing gained prevalence over all natural climate forcings as well as all negative anthropogenic forcings (Hansen et al., 2008). Second, as intimated by Zalasiewicz et al. (2011), the global fallout of bomb-test radioisotopes has the potential to create a global marker horizon for the Anthropocene. While some of these isotopes, namely ^{90}Sr and ^{137}Cs , have short

half-lives (~ 30 years) that curtail their long-term utility in this regard, others, such as ^{239}Pu , ^{240}Pu and ^{241}Am , are much longer lived. Furthermore, Pu radioisotopic measurements are becoming increasingly routine with advances in inductively-coupled-plasma mass spectrometry (ICP-MS), and their distribution in sediments better understood (Ketterer et al., 2002, 2004a). Sediment $^{239+240}\text{Pu}$ profiles mirror those of ^{137}Cs activity, and thus provide more lasting fingerprints of cold-war nuclear testing and the Chernobyl accident. Moreover, Pu isotopic ratios ($^{240}\text{Pu}/^{239}\text{Pu}$) effectively differentiate local sources from global fallout (Ketterer et al., 2004b), while sediment Pu appears highly immobile in a range of sedimentary environments (Sanders et al., 2010; Kuehl et al., 2012). For these reasons, we view Pu isotopes as a potentially valuable marker horizon, and encourage further measurements.

In conclusion, a relatively late date for the onset of the Anthropocene, one that straddles the “Great Acceleration”, reconciles the primary data presented in this review with expressions from other regions and proxies, and for which an explicit and lasting anthropogenic radionuclide fingerprint is likely to exist. The lithological, geochemical, and biological changes preserved in lake sediments far removed from direct human impacts are conservative measures that underscore the following global reality: the Holocene has effectively ended and the Anthropocene more aptly describes current planetary dynamics.

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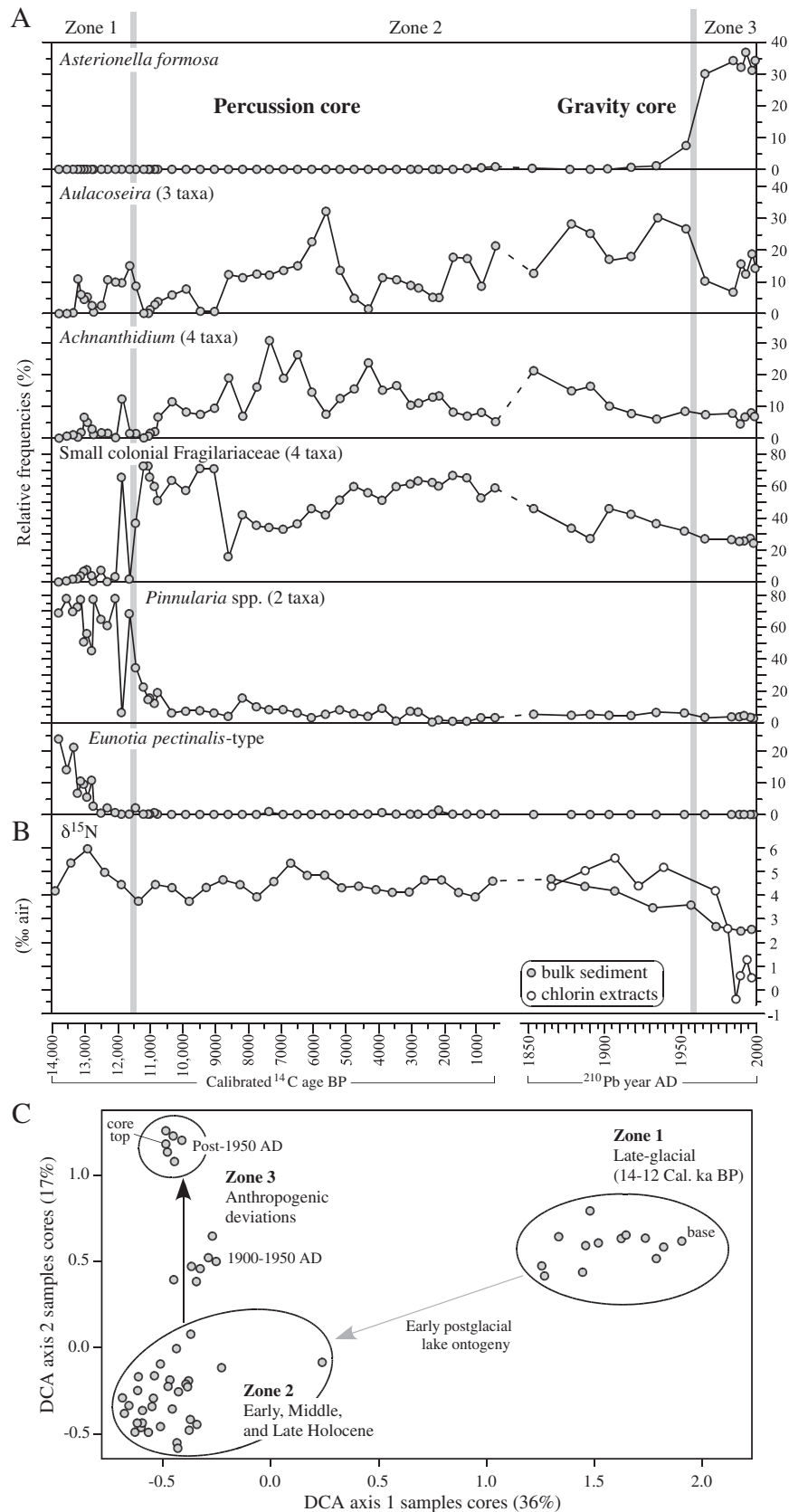


Fig. 17. Millennial-scale sediment stratigraphy from Sky Pond allowing for comparison between naturally-mediated changes and those associated with the Holocene–Anthropocene transition. Diatom relative frequencies (A) and nitrogen stable isotopes (B) are shown for both a percussion core spanning the last 14 cal.ka BP (Menounos and Reasoner, 1997) and gravity cores for the interval of unsupported ^{210}Pb activities (Wolfe et al., 2001; Enders et al., 2008). Diatom zones were established by cluster analysis and the significance of zones assessed with broken-stick methods (Bennett, 1996). A biplot of sample scores for the first two axes of DCA extracted from the diatom data (C) illustrates the unique character of the most recent assemblages in the context of natural variability.

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