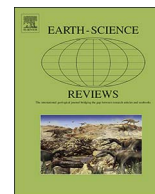


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Global Boundary Stratotype Section and Point (GSSP) for the Anthropocene Series: Where and how to look for potential candidates

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ARTICLE INFO

Keywords:

Anthropocene

Global Boundary Stratotype Sections and

Points

Chronostratigraphy

Palaeoenvironments

ABSTRACT

The Anthropocene as a potential new unit of the International Chronostratigraphic Chart (which serves as the basis of the Geological Time Scale) is assessed in terms of the stratigraphic markers and approximate boundary levels available to define the base of the unit. The task of assessing and selecting potential Global Boundary Stratotype Section and Point (GSSP) candidate sections, a required part of the process in seeking formalisation of the term, is now being actively pursued. Here, we review the suitability of different stratified palaeoenvironmental settings and facies as potential hosts for a candidate GSSP and auxiliary sections, and the relevant stratigraphical markers for correlation. Published examples are evaluated for their strengths and weaknesses in this respect. A marked upturn in abundance of radioisotopes of ^{239}Pu or ^{14}C , approximately in 1952 and 1954 CE respectively, broadly coincident with a downturn in $\delta^{13}\text{C}$ values, is applicable across most environments. Principal palaeoenvironments examined include: settings associated with accumulations of anthropogenic material, marine anoxic basins, coral reefs, estuaries and deltas, lakes at various latitudes, peat bogs, snow/ice layers, speleothems and trees. Together, many of these geographically diverse palaeoenvironments offer annual/subannual laminae that can be counted and independently dated radiometrically (e.g. by ^{210}Pb). Examples of possible sections offer the possibility of correlation with annual/seasonal resolution. From among such examples, a small number of potentially representative sites require the acquisition of more systematic and

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comprehensive datasets, with correlation established between sections, to allow selection of a candidate GSSP and auxiliary stratotypes. The assessments in this paper will help find the optimal locations for these sections.

1. Introduction

The Anthropocene Working Group (AWG), a working group of the Subcommission on Quaternary Stratigraphy (SQS) of the International Commission on Stratigraphy (ICS), is facilitating the process that will lead to the submission of formal proposals to define the Anthropocene as a chronostratigraphic unit. Such a unit comprises a body of strata formed during a specific interval of geological time. Units of the International Chronostratigraphic Chart (upon which the Geological Time Scale is based) are chronostratigraphic units, and each is defined by a synchronous base. The AWG is working towards a definition of the geological Anthropocene based on “the first appearance of a clear synchronous signal of the transformative influence of humans on key physical, chemical, and biological processes at the planetary scale. As such, it stands in contrast to various local or diachronous inscriptions of human influences on the Holocene stratigraphic record” (Zalasiewicz et al., 2017c). This working definition most closely aligns with the mid-20th century “Great Acceleration” in human population, resource consumption, global trade and technological evolution, proxy signals from which produce a distinctive stratigraphical boundary (Steffen et al., 2016). There are alternative interpretations of the definition of the Anthropocene, but these are generally grounded on a non-stratigraphical basis (e.g. the discussion on various geomorphological considerations of the start of the Anthropocene by Brown et al., 2017).

Within the Phanerozoic, the current internationally agreed method for defining chronostratigraphic boundaries is via selection of a Global Boundary Stratotype Section and Point (GSSP) as a physical reference level for a particular, and optimally correlatable, geological time boundary. The process of deciding on a GSSP, outlined by Remane et al. (1996) and Remane (1997, 2003) and summarised by Smith et al. (2014) is a complex process that normally requires: 1) an initial selection of a boundary level characterised by a marker event (the primary marker event) of optimal global correlation potential; 2) selection of a stratotype section from a number of potential sections, with the chosen stratotype section containing the best possible record of the primary marker event as well as other marker events that support global correlation; 3) ideally the selection of some auxiliary stratotypes in which the same level is represented by similar or other proxy signals in different parts of the world (Walker et al., 2009, in defining the Holocene Series, provided five auxiliary stratotypes as well as the GSSP); and, 4) definition of the precise point within stratified rock or sediment (or glacial ice in the case of the Holocene) that fixes the chronostratigraphic boundary with a precise moment of time. Formalisation of a GSSP is a careful procedure as, once ratified, it normally cannot be subsequently revised for at least ten years (Remane et al., 1996).

Table 1 provides a formal and comprehensive listing of the reasonable requirements for establishment of a GSSP, most of which will pertain also to a formal basis for the Anthropocene. It includes the requirement for stratigraphical completeness across the GSSP level, with adequate thickness of strata both above and below the boundary in order to demonstrate the transition. Therefore, the presence of an unconformity, marking a discontinuous succession, at or near the proposed boundary, would render it unsuitable. The selected section should also be accessible for subsequent investigations, ideally with provision for conservation and protection of the site.

The rank currently preferred by the AWG for the Anthropocene is that of series/epoch (Zalasiewicz et al., 2017c). The procedure leading to official acceptance of a GSSP for the Anthropocene Series/Epoch and its corresponding Age/Stage would require: 1) the selection by the AWG of a single GSSP candidate from one or more potential candidates,

based on proposals submitted to it; 2) the recommendation of that proposal by the SQS; 3) its approval by the voting membership of the ICS; and 4) ratification by the Executive Committee of the International Union of Geological Sciences (IUGS). All voting within the ICS, and its constituent subcommissions and their working groups, requires a supermajority of 60% or more for a proposal to be approved.

Here we offer a preliminary assessment of palaeoenvironments and their depositional facies where potential GSSP candidate sections for defining the lower boundary of the Anthropocene may be located, based on the published literature. Few of these example sections were chosen with the specific purpose of defining the Anthropocene as a chronostratigraphic unit. Rather they show a range of proxy signals, analysed in published studies for varied (non-ICS) purposes. The palaeoenvironmental research illustrated in this review demonstrates the timing and processes through which these signals have been imprinted in strata, and the extent to which they allow stratigraphic correlation worldwide. Even with this considerable caveat, the possibilities of correlation are clearly demonstrated, and help constrain the range of potential targets for Anthropocene-specific ICS studies.

2. Key stratigraphic markers

The aim – not always achieved – is for GSSPs to have many guiding criteria to support the primary marker (Remane et al., 1996; Smith et al., 2014) to permit both regional and global correlation. This has been the early focus of the AWG, with the description of potentially suitable markers summarised by Waters et al. (2016), whose recommendations this study follows, and as reported by Zalasiewicz et al. (2017c), concludes that the primary marker for the

Table 1

Requirements for establishing a Global Boundary Stratotype Section and Point (GSSP). Table modified from Gradstein et al. (2012, Table 2.1, p. 36), revised from Remane et al. (1996) according to current procedures and recommendations of the ICS.

1. Name and stratigraphic rank of the boundary
 - Including concise statement of GSSP definition
2. GSSP geographic and physical geology
 - Geographic location, including map coordinates
 - Geological setting (lithostratigraphy, sedimentology, palaeobathymetry, post-depositional tectonics, etc.)
 - Precise location and stratigraphic position of GSSP level and specific point
 - Stratigraphic completeness across the GSSP level
 - Adequate thickness and stratigraphic extent of section above and below
 - Accessibility, including logistics, national politics and property rights
 - Provisions for conservation and protection
3. Primary and secondary markers
 - Primary correlation marker (event) at GSSP level
 - Secondary markers – biostratigraphy, magnetostratigraphy, chemical stratigraphy, sequence stratigraphy, cycle stratigraphy, other event stratigraphy, marine-land correlation potential
 - Potential age dating from volcanic ash and/or orbital tuning
 - Demonstration of regional and global correlation
4. Summary of selection process
 - Relation of the GSSP to historical usage
 - References to historical background and adjacent (stage) units
 - Selected publications
 - Other candidates and reasons for rejection
 - Summary of votes and received comments
 - Other useful reference sections
5. Official publication
 - Summary for official documentation in IUGS journal *Episodes*
 - Digital stratigraphy (litho-, palaeo-, magneto-, and chemo-stratigraphic) images and graphic files submitted to ICS for public archive
 - Full publication in an appropriate journal

Holocene–Anthropocene boundary should be selected and identified in strata with a mid-20th century age. Such a definition makes the Anthropocene so recent that there are more potential archives available to it than for most, if not all, earlier GSSPs.

2.1. Key markers

The lower boundary of a chronostratigraphic unit, at the rank of Stage/Age and above, is defined by a GSSP and is recognised globally by a primary marker within the stratotype section that should be close to or coincident with the GSSP itself. Historically, most chronostratigraphic units use biotic signals, namely the highest or (preferably) lowest occurrence of a single fossil species, as the primary marker (Fig. 1). More recent approaches incorporate physico-chemical markers such as magnetic reversals, isotope excursions and cyclostratigraphy, especially for the Cenozoic (Miller and Wright, 2017). Although not a requirement, the means for numerically dating the succession is considered a strong site advantage.

Of the 65 GSSPs presently ratified by IUGS, all occur within strata deposited in marine sedimentary environments, with the exception of the Holocene GSSP that is placed within a Greenland ice core (Walker et al., 2009). This reflects the fact that marine fossils have proven better for long-distance correlation than non-marine ones; and more generally that proxy signals are readily preserved and continuously recorded in

marine environments. However, a primary marker ideally is traceable globally in marine and non-marine stratigraphic settings, thereby allowing secondary markers to be chosen that will assist correlation in either setting, especially when the primary marker is not detected. These secondary markers should approximate closely to the stratigraphic position of the primary marker. In an Anthropocene context, the choice of markers should not be to provide an indication of the start of anthropogenically driven effects, but to give the most pragmatic marker that will allow global correlation of the chosen boundary, being geographically extensive, temporally abrupt and providing a permanent record. Such primary and secondary markers do not need to be symptomatic of the environmental changes across the Holocene–Anthropocene boundary, rather they need to be amenable to widespread correlation in deposits representing different palaeoenvironments. Key signals considered here are mainly anthropogenic in origin, either being entirely novel, such as the production of novel anthropogenic mineral-like compounds, artificial radionuclides, organic and inorganic chemical species, or through perturbations of natural signals associated with, for example, the carbon and nitrogen cycles, and changes to climate proxies or biotic assemblages (Table 2). However, markers of any kind, including ones with little or no human influence (e.g. palaeomagnetic signals or volcanic ash layers) may serve just as well to help correlation. These markers are discussed by Waters et al. (2016), and are summarised below in the context of their likely presence in potential candidate palaeoenvironments.

2.1.1. Novel materials

Synthetic solids such as metal alloys, glass, carbide abrasives, gemstones, laser crystals, piezoelectric compounds, semiconductors and cement have proliferated in the environment in recent decades (Hazen et al., 2017), being present in novel “metamorphic rocks” such as porcelain, brick and concrete, the latter being the most abundant novel rock on the planet, with over 90% of the ~500 Gt (gigatonnes or 10^{15} g) produced since the mid-20th century. Complex additives to cement help to chemically and petrographically fingerprint modern concrete (Waters and Zalasiewicz, 2018). These novel anthropogenic mineral-like compounds and rocks often accumulate directly within terrestrial anthropogenic deposits such as landfill sites and road networks. Through subsequent erosion of the widespread human-built landscape, these materials may be reworked and deposited in fluvial, lacustrine, coastal and marine sediments, with a variable time lag between formation and accumulation.

Synthetic organic compounds have become abundant, including plastics that are exceptionally mobile and durable within river and ocean currents. Plastics have, since ~1950 CE, become increasingly widely dispersed, including as abundant microfibres and microbeads, forming signals found extensively in lake, estuarine and marine sediments, even within remote abyssal environments (Zalasiewicz et al., 2016a). Most are the effluents from waste water treatment works and transported through river systems to accumulate in lakes or oceans. The recent recognition of synthetic microfibre fallout accumulations in urban Paris (Dris et al., 2016) raises the possibility of rapid and widespread airborne dispersal across environments.

The first development (in effect first occurrence) and annual production figures of novel anthropogenic mineral-like compounds and organic polymers are known from historical records. However, few studies have investigated them in their stratigraphic context, in order to determine the pattern of their accumulation in sedimentary environments. Matsuguma et al. (2017) record the first influx of microplastics into canal sediments in Tokyo in the 1950s, whereas in the Gulf of Thailand a significant increase in abundance occurs in the 1990s; in all cases there is an upward increase in microplastic abundance and polymer types. Because of their novelty, little is known of the longevity and potential fragmentation of plastics in buried sediments. Also of significance in lakes and marine settings is the settling rate within the water column. In the case of microplastic organic polymers (< 5 mm

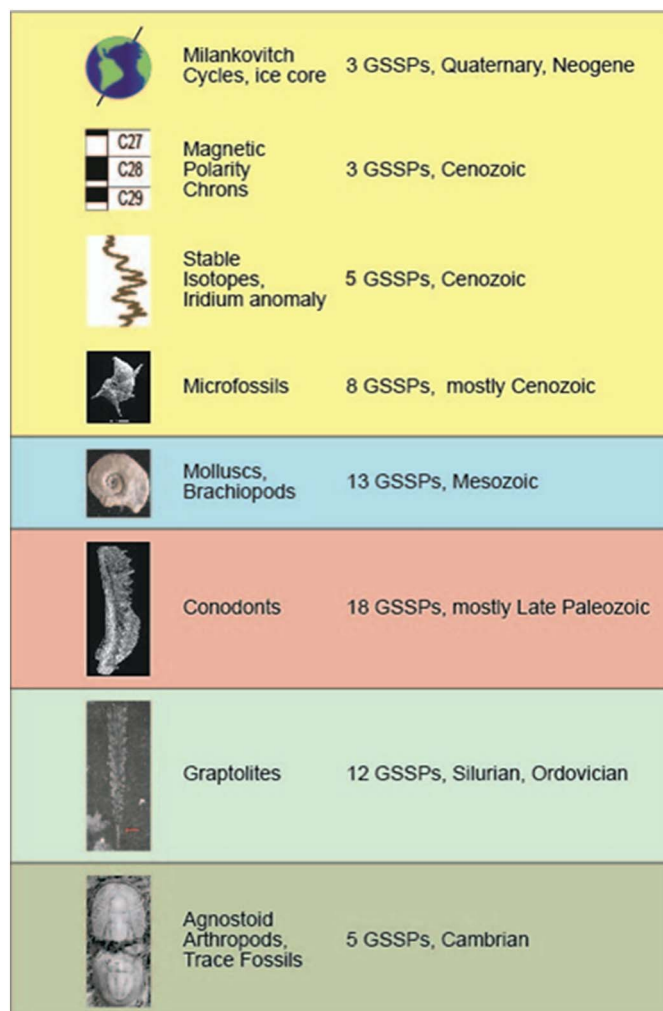


Fig. 1. Primary and some secondary markers for the 65 GSSPs that have been ratified currently by IUGS. The four in the Cenozoic that deal with stable isotope events (base Eocene, Quaternary, Calabrian and Holocene) and the iridium anomaly (base Paleocene) provide models for choosing markers for the base of the Anthropocene.

Table 2

Potential palaeoenvironmental archives and facies and their stratigraphic markers for a candidate GSSP.

Marker Environment	Annual laminae	Novel materials		Geochemical markers									Biotic markers Extinctions/ neobiota
		Plastics	Fly ash	$\delta^{18}\text{O}$	Deuterium &/or dust	CO_2 & CH_4	$\delta^{13}\text{C}$	NO_3^- & $\delta^{15}\text{N}$	S & SO_4^{2-}	Heavy metals	Organic compounds	Radiogenics	
Anthropogenic deposits		✓	✓				✓	✓		✓	✓	✓	✓
Marine anoxic basin deposits	✓	✓	✓	✓			✓	✓		✓	✓	✓	✓
Coral bioherms & marine bivalve shells	✓	✓	?	✓			✓	✓		✓	✓	✓	
Estuarine & deltaic deposits	✓	✓	✓	✓			✓	✓		✓	✓	✓	✓
Lake deposits	✓	✓	✓	✓			✓	✓		✓	✓	✓	✓
Peat & peatlands (mires)			✓		✓		✓	✓	✓	✓	✓	✓	✓
Ice	✓		?	✓	✓	✓	✓	✓	✓	✓	✓	✓	
Speleothems	✓			✓					✓	✓	✓	✓	
Trees	✓			✓	✓		✓	✓	✓	✓		✓	

? – uncertain suitability of the specified signal.

diameter) sinking velocities have been estimated as $6\text{--}91\text{ mm s}^{-1}$, which is more rapid than the settling of clay minerals and organic-rich aggregates (Kowalski et al., 2016). Particle density (variable between distinct polymers) and shape, fluid density (with slower settling in saline water), temperature and ingestion by microplankton all can influence settling rates, and these parameters have different effects on different polymers (Kowalski et al., 2016).

Fly ash, the unburned particulates sourced from fossil fuel combustion, includes inorganic ash spheres (IAS), mainly residues from coal burning, and spheroidal carbonaceous particles (SCPs) principally from both coal burning and oil combustion. These particulates are disseminated aerally, and typically show a marked upturn in abundance from about 1950 CE (Oldfield, 2014; Rose, 2015; Swindles et al., 2015). The upturn in abundance of SCPs should serve as a likely permanent stratigraphic marker for the Anthropocene in lake sediments across all continents. Work is ongoing to assess their likely presence in other environments, such as glacial ice, peat sequences, marine basin sediment and terrestrial anthropogenic deposits such as industrial dumps and landfills. Black carbon (BC), present as char or soot, is generated from both anthropogenic combustion and natural biomass burning (Bond et al., 2007; Han et al., 2017). BC is commonly identified in soil, lacustrine and marine sediments, and glacial ice (Xu et al., 2009).

2.1.2. Geochemical markers

Oxygen and hydrogen isotopes and dust signals are important climate change proxies and hence tools for Quaternary correlation. All three proxies are recognised in glacial ice, within the ice fraction, and complement the air bubble record. Deuterium (^2H or D) is an important isotopic tracer for precipitation, and the relationship between δD and $\delta^{18}\text{O}$ signatures defines a “deuterium excess” marker that is a proxy for sea surface temperature. The deuterium excess ratio is used as the primary marker for the base of the Holocene Series in an ice core GSSP (Walker et al., 2009), where a 2–3‰ decrease in the ratio is associated with a rapid ocean surface temperature decline of $2\text{--}4\text{ }^\circ\text{C}$. An abrupt dip in δD and deuterium excess reflecting cooling is also seen in a Greenland ice core at 8.2 ka (Walker et al., 2012), and suggested to

mark the base of the proposed Middle Holocene Subseries and Northgrippian Stage. Cold stages are commonly associated with increased aeolian dust deposition in ice cores. Decreases in $\delta^{18}\text{O}$ mark cooling events, and in Greenland ice the overall trend since the Middle to Late Holocene has been of gradual cooling culminating in the Little Ice Age from about 1200 to 1850 CE. The shift to warming induced by increased emissions of anthropogenic greenhouse gases is first indicated by a slight change to less negative $\delta^{18}\text{O}$ values in Greenland ice from ~1850 through to the 1870s CE and again from 1979 to 2007 CE (Masson-Delmotte et al., 2015; see Fig. 24 below), although the magnitude of shift is small compared with variations across the Pleistocene–Holocene boundary. Warming is also apparent in the Antarctic Peninsula from about 1950 CE, e.g. on James Ross Island (Abram et al., 2013), with lakes on Signy Island from the South Orkney Islands having experienced some of the most rapid air temperature increases of $2\text{ }^\circ\text{C}$ since the mid-20th century, with concomitant marked responses by lake biota and increased nutrient levels, linked to deglaciation and reductions in lake snow and ice cover (Quayle et al., 2002). Variations in $\delta^{18}\text{O}$ are observable in carbonate and phosphate minerals in speleothems, shells, corals, teeth, and bones and also in tree rings. They can be used as a palaeotemperature proxy in marine settings or within terrestrial environments, primarily as a proxy for $\delta^{18}\text{O}$ composition in atmospheric precipitation.

About 50% of anthropogenic carbon dioxide emissions, the major product of fossil fuel combustion, have accumulated in the atmosphere, with the lifetime of much of the atmospheric fossil fuel-derived CO_2 being ~200 to 2000 years (Archer et al., 2009). The remainder is absorbed approximately equally by land and ocean sinks (Le Quéré et al., 2016). The atmospheric CO_2 record from ~7 ka shows a very slow rise to ~280 ppm by ~1850 CE. Subsequently, the CO_2 concentration shows a marked rate of increase (Etheridge et al., 1996; MacFarling Meure et al., 2006; Rubino et al., 2013), attaining ~310 ppm by 1950 CE and with a further sharp acceleration to its current mean value of > 400 ppm, reached in 2015 CE. The rate of increase between 1950 and 2015 CE is ~100 times greater than that of the Late Pleistocene to Early Holocene rise, itself considered rapid in geological terms (Wolff, 2014).

Anthropogenic methane emissions arise primarily from agriculture (wet rice cultivation and ruminants), with a smaller fraction from biomass burning and fossil fuel combustion. Some large terrestrial anthropogenic deposits in the form of landfills also emit significant amounts of methane. Unlike CO₂, methane is chemically reactive in the atmosphere, with the dominant sink being oxidation by OH radicals. This process removes about 90% of all emissions of CH₄ each year, or about 9% of the total burden of CH₄ – the atmospheric lifetime is 7–11 years (Ciais et al., 2013). Throughout the Quaternary Period, CH₄ concentrations range from ~400 to 800 ppb, with the higher values marking interglacials. In marked contrast, from 1875 CE there has been a sharp climb to the current levels of ~1800 ppb (Nisbet et al., 2016).

Both CO₂ and CH₄ provide a direct stratigraphic marker where they are preserved within air bubbles in glacial ice, and an indirect record via carbon isotope variations. $\delta^{13}\text{C}$ values in CO₂ from ice cores varied by < 0.4‰ over the duration of the pre-industrial Holocene, with the principal change being a slow enrichment in ^{13}C during the Early Holocene. The subsequent combustion of hydrocarbons, which are depleted in the ^{13}C isotope, has resulted in a sharp reduction in $\delta^{13}\text{C}$ values in atmospheric CO₂ of about 2‰ since the late 18th century, with a pronounced inflection at about 1960 CE (Rubino et al., 2013). This Suess effect is recorded in diverse materials, including plants (tree rings), shells, bones, limestones, corals, and in the atmospheric composition of air bubbles trapped in ice, although speleothems are comparatively insensitive to this signal. An abrupt shift towards less negative $\delta^{13}\text{C}$ values in atmospheric CH₄ of ~2.5‰ since ~1875 CE in response to these pyrogenic emissions (Ferretti et al., 2005) has been succeeded over the last 12 years by a shift of –0.17‰ of $\delta^{13}\text{C}$ in the increasing volume of atmospheric CH₄, attributed to an expansion of tropical wetlands (Nisbet et al., 2016).

Nitrate concentrations have been perturbed mainly through the production of agricultural fertilisers, initially through mined nitrate deposits in the 19th century, but most notably through the industrial production of ammonia from atmospheric nitrogen by the Haber-Bosch process, which started in 1913 CE (Fowler et al., 2013). The nitrate associated with artificial fertilisers shows global airborne distribution. $\delta^{15}\text{N}$ values have been acquired from sediment organic matter, commonly from lake deposits, although determinations from anthropogenic deposits, marine anoxic basins, estuaries, deltas, tree rings, and amino acids in corals are also available. Nitrogen compounds and isotopes are not typically recorded in carbonates, including speleothems. Nitrate, total nitrogen oxides (NO_x) and $\delta^{15}\text{N}$ values are determined from solid ice in ice cores, providing data at the time of snow accumulation. Both Northern Hemisphere lake sediments (Holtgrieve et al., 2011; Wolfe et al., 2013) and glacial ice (Hastings et al., 2009) show elevated nitrate concentrations and a fall in $\delta^{15}\text{N}$, which start towards the end of the 19th century but become appreciably more pronounced during the mid-20th century. A marked increase in NO_x sourced from elevated combustion of hydrocarbons, particularly by automobiles and power stations, is also evident in Greenland ice sheets from the mid-20th century (Erisman et al., 2013).

Natural sources of sulphate include volcanic eruptions that produce high-magnitude, short-duration spikes evident in glacial ice over past millennia. There has been an overall trend of increased sulphate content in glacial ice, trees and speleothems since the start of the Industrial Revolution, peaking in the mid- to late-20th century, reflecting patterns of coal consumption; sulphate has widely declined over recent decades, most likely due to legislation requiring the fitting of sulphur scrubbers to power station chimneys to combat acid rain. The signals tend to be associated with local-to-regional industrial changes rather than representing global signals. Stable sulphur isotopes show lower $\delta^{34}\text{S}$ ratios in tree rings (e.g. Fairchild et al., 2009; Wynn et al., 2014) and speleothems (e.g. Frisia et al., 2005; Wynn et al., 2010) during the second half of the 20th century.

Pure metals such as gold, silver, lead, copper and iron have a long history of extraction (and blending to obtain alloys) to produce tools or

ornaments. The early historical mining of these metals provides local and highly diachronous signals evident in lakes, estuaries, peats, sea-floor sediments, and carbonates including corals and speleothems, and they are also present in glacial ice records. The heavy metals are either air-borne as aerosols, dissolved in water, or carried as water-transported particulates. Mercury (Hg) is also transported in gaseous form (RGM – reactive gaseous Hg; GEM – gaseous elemental Hg) (Yang et al., 2010). Within sediments, metal compounds are commonly stabilized by being adsorbed to clay mineral surfaces or organic matter; locally they may be diagenetically mobilized as conditions become reducing with burial, and redeposited in overlying oxidizing environments, thus decreasing their effectiveness as stratigraphic markers. Elevated heavy metal concentrations in tree rings have been suggested as unsuitable for temporal studies of pollution events (Bindler et al., 2004). For most metals and their compounds, the 20th century environmental signals continue to be localised, related to production from local industries. Some metals, such as elemental aluminium and many trace elements, have seen marked upturns in production in the second half of the 20th century as new uses are found for them, and these may be expected to leave signals in sediments. For at least two millennia prior to the Industrial Revolution, Pb concentration in sediments increased and Pb isotope ratios declined as a result of European metallurgy (Renberg et al., 2000). Perhaps the most widespread and abrupt signal is that associated with Pb isotope ratios related to the use of tetraethyl-lead in gasoline from the 1920s, although the isotopic fingerprint reflects the sources of lead used to manufacture this additive, which differ between countries (Reuer and Weiss, 2002). Current global atmospheric concentrations of Hg are about two to four times higher than pre-industrial levels as a result of refining ore and burning coal and waste (Lindberg et al., 2007; Yang et al., 2010). Hg has an atmospheric lifetime of 0.75 years so can be dispersed globally before deposition, whereas the larger and more slowly cycling marine reservoir can produce a lagged response of 1 to 2 centuries (Lindberg et al., 2007).

A huge variety of novel organic compounds have been generated and used in industrial processes. Persistent organic pollutants (POPs), those recalcitrant to decomposition, which readily adsorb onto clay and organic particles and are poorly soluble, provide potentially suitable signals. Key potential POP signals include those from organochlorine pesticides (OCPs) such as DDT, aldrin and dieldrin, and various polychlorinated biphenyls (PCBs), along with brominated flame retardants (e.g. polybrominated diphenyl ethers, PBDEs). Most of these have only been generated on industrial scales and released into the environment in notable quantities since the mid-20th century, particularly concentrating in sediments in lakes (Muir and Rose, 2005, 2007), estuaries (Vane et al., 2011) and canyons on continental shelf margins (Paull et al., 2006).

Large amounts of anthropogenically-sourced radionuclides have been released by 543 atmospheric nuclear weapons tests, with a total yield of about 440 Mt (i.e. energy released in TNT equivalent), and subsequent fallout (UNSCEAR, 2000). Although atmospheric tests and military use began in 1945 CE, it was only with the testing of the large thermonuclear (hydrogen) devices from 1952 CE that fallout was dispersed globally and became recorded in most environments. The signal is almost instantaneous in geological terms, with residence times in the stratosphere of some 15 to 18 months following thermonuclear detonations. Typically, the Pu signal shows an abrupt rise in 1952 CE, a peak in 1963–1964 CE and rapid decline since 1963 CE in response to a Partial Test Ban Treaty, which resulted in most testing moving underground (Waters et al., 2015). Radionuclides that are absent or rare in nature, such as plutonium-239 (^{239}Pu), americium-241 (^{241}Am), caesium-137 (^{137}Cs) and strontium-90 (^{90}Sr) show clear signals in soils, peats, lake and sea-bed deposits, coral and tree rings, and glacial ice. However, the caesium and strontium isotopes have half-lives of about 30 years, so their sedimentary signal is short-lived, while caesium and plutonium may be mobile in anoxic conditions (Jeandel, 1981). Additional brief spikes of certain isotopes can be used to aid correlation. For

instance, an artificial satellite re-entry in 1964 CE provided a distinctive global ^{238}Pu signal, and, more regionally, the Chernobyl and Fukushima reactor accidents produced spikes in 1986 and 2011 CE, respectively. The common natural radioisotope ^{14}C shows greatly elevated values associated with nuclear tests and provides a further clear and long-lasting signal in most growing organic matter such as wood, bone and carbonates including corals, speleothems and also in glacial ice. ^{239}Pu , because of its longevity (half-life of 24,110 years), is a strong candidate as a primary marker of radionuclide deposition as it will remain detectable for $\sim 100,000$ years (Hancock et al., 2014) and beyond as the decay product ^{235}U . Alpha-spectrometry carried out during early studies does not differentiate between ^{239}Pu and ^{240}Pu (itself not an issue, as the combined signal is clearly expressed), but is only applicable where Pu concentrations are high – the detection limit is approximately 0.05 mBq of $^{239} + ^{240}\text{Pu}$ (Hancock et al., 2014). Modern ICP-MS techniques can discriminate between these isotopes, although it is necessary to remove interferences from $^{238}\text{U}^+$ and other species, and can be used at low Pu concentrations – the absolute limit of quantification is 0.2 μBq and 0.9 μBq for ^{239}Pu and ^{240}Pu respectively (Lindahl et al., 2011). Other mass spectrometric methods, such as Accelerator Mass Spectrometry (AMS) and Thermal Ionisation Mass Spectrometry (TIMS) also allow the simultaneous determination of ^{239}Pu and ^{240}Pu .

2.1.3. Biotic markers

Changes in fossil assemblages are used to demarcate most chronostratigraphic boundaries in the Phanerozoic, mainly through recognition of the lowest stratigraphic occurrence of a fossil taxon, which may approximate to its evolutionary origin, although within any individual stratal section the stratigraphic appearance of a taxon more specifically reflects its local immigration to that place. Such stratigraphic appearances (and commonly disappearances too) provide a widely effective tool for correlation. In the Quaternary Period (and especially for the Anthropocene) the time scales involved are generally too short for many evolutionary appearances and disappearances to have taken place, and so immigration/emigration patterns are widely used for biostratigraphic correlation. Prominent Anthropocene biostratigraphic signals are expressed in recently forming sedimentary deposits by the appearance of unique terrestrial and marine species associations of the neobiota ('invasive species'), the result of human transport of tens of thousands of species around the globe. Recent neobiota include examples among common microplankton such as foraminifera, coccoliths, dinoflagellates and radiolarians in the marine realm, of the kind commonly used as microfossils to recognise stage boundaries in pre-Anthropocene deposits of the Cenozoic (Fig. 1). These organisms, as well as macroinvertebrates, such as molluscs, vertebrates and plants (both macroplant remains and pollen) can form useful biostratigraphic tools for correlation because of their geologically preservable hard parts such as skeletons, frustules, shells, or phytoliths.

Human-driven extinctions and local extirpations have become increasingly common over the last 500 years, particularly in island faunas (Ceballos et al., 2015). So far, island floras have been more resilient to extinction than island faunas in spite of the high abundance of introduced plant species in many environments (Ellis et al., 2012), but an accumulating extinction debt is likely (Sax and Gaines, 2008). Their provincial nature and low abundances mean that island floras provide regional rather than global stratigraphic markers. Selective breeding of domesticated floras and faunas tends to be associated with rapid changes to morphology, in effect forming distinct morphospecies that can be preserved in sedimentary successions. Moreover, species beneficial to humans, such as domestic maize or chickens, tend to be abundant and geographically widespread. The development of factory farming led to massive increases in amounts of faunal remains of domesticated species being deposited in landfills and other terrestrial stratigraphic contexts from the mid-20th century on.

Another extensive stratigraphic signature over recent centuries, but

especially since the mid-20th century, reflects neobiota reaching far beyond their original geographical ranges, through deliberate or accidental human actions. Local arrival of these neobiota is well documented through observation, and is often linked with the extinction or extirpation of local fauna and flora, as observed on Kauai, Hawaii (see Section 3.5.5), or among benthic foraminifera preserved in the coastal sedimentary record (McGann et al., 2000; Calvo-Marcilese and Langer, 2010).

Biostratigraphic signals may also relate to ecological degradation (Willis et al., 2010; Dearing et al., 2012). In marine environments, a long lead-in of low-level and provincialized ecological degradation gave way to accelerating global change in the early 1950s (Wilkinson et al., 2014). The driver was expansion of industrial pollution, expanding human populations associated with increased raw sewage dispersal offshore, and increased use of fertilisers during agricultural intensification. In Northern Hemisphere lakes, an increase in reactive nitrogen from artificial fertilisers has been well-documented (Holtgrieve et al., 2011; Wolfe et al., 2013), giving rise to increases in primary productivity and marked changes to diatom populations and species assemblages, with surface water acidification as an additional factor. The stratigraphic records collated by Smol (2008) commonly show clear mid-20th century inflections in such lacustrine proxies. Runoff contaminated with these fertilisers has also resulted in seasonally oxygen-starved 'dead zones' in lakes and coastal seas, causing – to take just one example – increases in hypoxia-tolerant benthic foraminifera (Wilkinson et al., 2014). Environmental legislation has locally led to a phase of recovery from pollution stressors in the late-20th and early-21st centuries, but increasingly climate change is imposing pole-ward shifts in species distributions and confounding recovery from acidification (e.g. Battarbee et al., 2014).

2.2. Independent dating techniques

Direct numerical dating of a GSSP is not a prerequisite in the selection of a site (Table 1), but is beneficial. For the Anthropocene, annual resolution for any proposed boundary section is highly desirable, giving a level of precision not available for any other boundary within the Geological Time Scale. The best time resolution for previous GSSPs is for that of the Holocene, with a 2σ uncertainty of ± 99 years (Walker et al., 2009).

Counting annual sedimentary laminae in presently accumulating successions, or annual growth layers in living organisms (that are demonstrably complete), is one means of establishing the numerical age, expressed in years, of an Anthropocene base. The age so determined may be additionally constrained by ^{210}Pb , sourced from the atmosphere or directly from sediments (Noller, 2000). This isotope is rapidly scavenged from sea and freshwater onto organic detritus, but is not mobile in the sediment. It has a useful dating range of up to 150 years (the half-life is 22.3 years) with a measurement error of about 10%, and is typically used to determine accumulation rates in organic lake, coastal or marine sediments, peats, or glacial ice; it is fixed in living plants and corals, too, and so can determine their growth rates.

Other short-lived radioisotopes include ^{234}Th , with a half-life of 24 days, which can be used to date changes in speleothems and corals over periods as short as 3 months, but it has little value for laminae straddling the Holocene–Anthropocene boundary. Radiocarbon (^{14}C) has a half-life of ~ 5600 years and has limited use as a chronometer from the mid-20th century, but along with ^{137}Cs (half-life 30 years) and ^{90}Sr (half-life ~ 28 years), it can help to identify specific episodes, such as the 1963–1964 CE peak in atmospheric nuclear tests (the 'bomb-spike') and the 1986 CE Chernobyl accident recorded in sediments, ice, tree rings and corals.

Anthropogenic materials present possibilities for absolute and relative dating not available in earlier geological successions. For example, the date of manufacture of artefacts (technofossils) found as inclusions in strata can often be ascertained with high resolution

through the use of type series, and sometimes the precise date is even stamped, moulded, cast, or indelibly imprinted on objects. Such evidence could be immensely valuable in supplementing more traditional dating methods. More generally, the value of using archaeological techniques for relative dating of stratigraphic sequences of terrestrial anthropogenic ground, in conjunction with geological methods, should be considered.

3. Suitability of palaeoenvironmental archives for hosting potential GSSP candidates

There is a wide range of sedimentary and biological records within which a GSSP is feasible. For any candidate section, high-precision dating is needed, if only to provide confidence that there are no missing surfaces in successions that are merely decades to centuries long. Such high-precision dating, potentially to annual resolution, can be achieved in seasonally layered sediments in anoxic marine basins and hypoxic lakes, in glacially influenced or hypersaline lakes and sinking marine deltas, in seasonally layered ice and annual growth layers in corals, bivalves, and in speleothems and trees. Non-layered strata, including anthropogenic deposits and peats provide additional environments in which the Holocene–Anthropocene transition may be observed. In this section these key potential archives are discussed in turn, describing their global extent and continuity, and the arguments for or against their potential for hosting a candidate GSSP. Examples are provided to show how key signals are recorded.

3.1. Anthropogenic deposits

This category includes sedimentary successions that have accumulated through direct human deposition (artificial ground) or by human influence on natural systems (Ford et al., 2014). Artificial ground can show a continuum from entirely natural through to entirely anthropogenic materials, but with the key requirement that they have been deposited through human action. Conversely, few “natural” systems are these days free from human modification (Barnosky et al., 2017), whether via construction of dams influencing sediment flux in fluvial systems (Syvitski and Kettner, 2011), or trawler fishing disturbing the upper decimetres of sediments on the continental shelf and slope (Martin et al., 2015).

Zalasiewicz et al. (2017b) estimated the spatial extent of human modification of the landscape at about $82 \times 10^6 \text{ km}^2$, or about 16% of the total Earth's surface or 55% of the terrestrial land surface, providing many sections that could be investigated as potential GSSPs. Road surfaces alone, with a mass estimated at 1.3% of the physical technosphere (sensu Zalasiewicz et al., 2017b), comprise an extensive boundary layer in temperate and tropical climates. Anthropogenic deposits may show remarkably high accumulation rates, locally in excess of one metre per year, and rapidly incorporate novel anthropogenic signatures, which may be lithological, geochemical or biotic. This provides a highly resolvable succession in which artefacts and novel materials can be constrained to decadal resolution (Zalasiewicz et al., 2014a).

A potential Anthropocene Series, which would be defined in a unit of strata, and would represent an equivalent Epoch of geological time, is not synonymous with anthropogenic impact. The lower boundary of anthropogenically-modified deposits (or archaeosphere in an archaeological context) may commonly be a marked unconformity, with the overlying first expression of the archaeosphere often being markedly diachronous over millennia (Edgeworth et al., 2015). Such deposits represent a lithostratigraphic – and when one considers technofossils – also a biostratigraphic unit, but not a chronostratigraphic subdivision with a necessarily isochronous basal boundary. The Anthropocene, if it is to be distinguished within these deposits, requires continuous stratiform accumulation of sediments within which a clear signal can be recognised that can distinguish Anthropocene anthropogenic deposits

from underlying Holocene anthropogenic deposits. Although such distinguishing signals can be readily found within these deposits, e.g. novel minerals, aluminium, concrete, plastics, organic compounds (Waters et al., 2016; Zalasiewicz et al., 2016 and references therein), the successions overall are typically marked by numerous internal erosion/hiatus surfaces; anthropogenic deposits are mainly formed by intermittent depositional events, lack simple vertical accretionary patterns, and are commonly disturbed by ‘anthroturbation’. The pattern of such complex erosional/non-depositional surfaces is of great value for local correlation, and also for correlation over wider areas where exact synchronicity is not required, but hinders precise regional to global correlation (Edgeworth et al., 2015). The typical stratigraphic incompleteness of these deposits also makes them generally unsuitable for hosting a GSSP. Furthermore, they are also mostly terrestrial (although increasingly they are also becoming a feature of the marine realm), making some of them prone to erosion. However, artificial deposits have potential for direct preservation over geological time scales when located on subsiding environments such as major deltas (see Section 3.4) and within subsurface excavations (Zalasiewicz et al., 2014b).

Table 3 summarises the key advantages and disadvantages of using anthropogenic deposits and examples of these extraordinarily diverse successions are provided: the Fresh Kills Landfill (New York), Teufelsberg (Berlin) and in Vienna, all of which are examples of artificial ground; and the Gorrondatxe-Tunelboca beachrock (northern Spain), a naturally formed succession comprising mainly artificial materials (Fig. 2).

3.1.1. Fresh Kills Landfill, New York

Fresh Kills Landfill on Staten Island (USA) is a former repository for New York City refuse. Covering some 8.9 million m^2 , up to 70 m high, and containing ~150 million tons of municipal waste, it may be the largest human-engineered formation in the world (Melosi, 2016). It was opened in 1948 CE (Melosi, 2016), with a peak influx of garbage reaching 29,000 tons per day (Nagle, 2008) and closed in 2002 CE, the last debris being from the 9/11 event in 2001 CE. The landfill deposit rests upon Holocene estuarine marsh deposits, with few intervening pre-mid-20th century anthropogenic deposits, has complex internal patterns, having accreted laterally as the site expanded, and may even have been overturned in places. Its internal complexity precludes it from hosting a functional GSSP.

3.1.2. Teufelsberg, Berlin

Rubble generated in Berlin during World War II was later redistributed in mounds within the city boundary, the largest being the Teufelsberg. The rubble, including concrete, brick, clinker, rock, fly ash, slag and solid chemical waste, was deposited between 1950 and 1972 CE and is up to 80 m thick, with an area of 1.1 million m^2 , and an original rubble volume of 26.2 million m^3 (Mielke, 2011; Cocroft and Schofield, 2012). The Teufelsberg deposit mainly rests unconformably upon Upper Pleistocene sand and till, and these areas would be

Table 3

Reasons for and against using an anthropogenic deposit as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> • Abundant, mainly terrestrial and notably in urban settings • Comparatively thick; rapid accumulation rates • Independent dating using ^{210}Pb and ^{14}C and historical and archaeological context • Highly resolvable technofossil stratigraphy • Rapid incorporation of novel materials and geochemical signals 	<ul style="list-style-type: none"> • No annual lamination • Limited extent in oceans • Limited lateral continuity • Highly variable deposition rates • Numerous omission surfaces • Prone to erosion and organic degradation • No guarantee of normal superposition

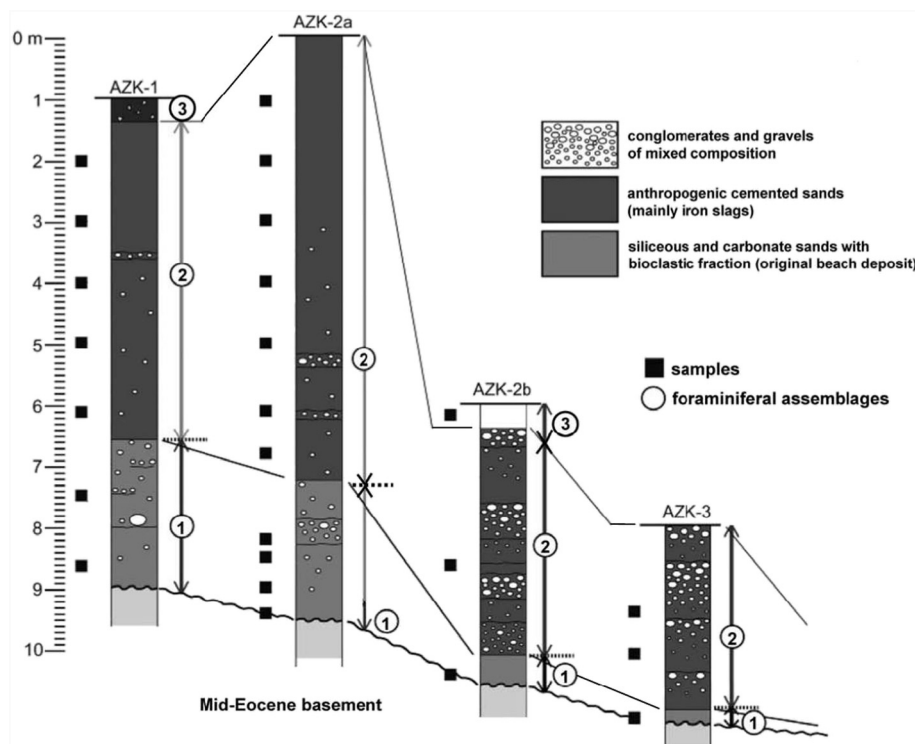


Fig. 2. Cores through the Gorrondatxe-Tunelboca beachrock, Spain. A natural high-energy open beach sand deposit, pre-dating the discharges of iron slag, is overlain by cemented sandy and coarse beach deposits that are commonly conglomeratic with abundant slag clasts and incorporating littoral foraminifera. Three distinct foraminiferal assemblages are recognised (ranges for assemblages 1, 2 and 3 marked by arrows). Modified from Martínez-García et al., 2013.

unsuitable for hosting a GSSP for that reason. But locally, thin Holocene deposits occur between the two units, and here the placing of a GSSP might be considered. The Teufelsberg comprises a distinct, mappable lithostratigraphic unit with a base approximating to that of the Anthropocene, which also extends lens-like across other rubble heaps of the Berlin area (cf. Zalasiewicz et al., 2016a), and these can be correlated through technofossil and novel material inclusions with other such deposits elsewhere. Similar lens-like post-World War II rubble heaps in Germany occur also at Cologne, Dresden, Frankfurt am Main, Hannover, Leipzig, Stuttgart and Munich, and in England in heavily bombed cities such as Coventry.

3.1.3. Vienna, Austria

In the 19th century, Eduard Suess (1862, 1897) mapped urban strata in Vienna as a geological unit (Schuttdecke), which comprised anthropogenic strata formed during development of the Ringstrasse. The deposits typically range from 4 to 5 m in thickness, locally reaching 10–12 m where infilling the former 17th century city moat, and include brick fragments, broken glass, coins, clay jars, and bones of humans and domestic animals. Subsequently, these artificial deposits extended laterally and aggraded vertically, with the newer Anthropocene materials being of distinctively different composition to those described by Suess. The succession of anthropogenic strata below Vienna dates back to the Neolithic, and has been more or less continuous since Roman times, but many unconformities and much reworking make them typically unsuitable for hosting a GSSP.

3.1.4. Gorrondatxe-Tunelboca beachrock, Spain

The Gorrondatxe-Tunelboca beachrock (Fig. 2) comprises 1 million tonnes of sediments accumulated since the 1940s. Its source is the 30 million tonnes of blast furnace waste that was tipped offshore between 1902 and 1995 CE (Irbien et al., 2015). Wave activity transported the waste and it re-sedimented as a beachrock deposit about 1.8 km long and 50 to 100 m wide. The succession comprises up to 7 m of deposits of anthropogenic origin, resting on up to 2.5 m of bioclastic beach sands (Martínez-García et al., 2013). The lowermost high-energy open beach deposit pre-dates the discharges of iron slag and has a distinct

foraminiferal assemblage (Fig. 2). This is overlain by sands with an anthropogenic component, where the foraminifera decrease upwards in abundance (assemblage 2 of Fig. 2), and then decrease in diversity to become largely monospecific (assemblage 3). The decrease in foraminiferal abundance and diversity coincides with the dumping of the iron foundry deposits offshore. The upper deposits comprise a foreshore succession of cemented conglomerate of furnace bricks, foundry slag with an upper sandy deposit that includes plastics (Zalasiewicz et al., 2016a). Internal erosion surfaces are present throughout, as in any beach deposit. Access to the section is good, the site coinciding with the existing GSSP for the Lutetian Stage (Middle Eocene), while the beachrock deposits are recognised as Geosite 96 of the Basque Region. The supply of anthropogenic materials to the shelf ceased in the 1990s, and the beach deposits have started to be eroded by waves, and so have limited potential for long-term preservation. The presence of internal erosion surfaces, the lack of clear lateral continuity, absence of annually laminated deposition, and poor preservation potential of the site make it unsuitable for hosting a candidate GSSP.

3.2. Marine anoxic basin deposits

The oceans where water is > 200 m deep comprise about 66% of the Earth's surface, and the sediments that accumulate there might be expected to preserve relatively complete and less disturbed successions compared with those on land or on the continental shelf. However, for deep-ocean deposits to form outcrops on land, rapid rates of uplift are needed and the prospects of such outcrops diminish with increasingly young deposits. Deep marine deposits of the Anthropocene, therefore, almost invariably need to be cored in situ. Cores are not ordinarily used for GSSPs because of their limited material and access, although exceptions were made for the base of the Holocene (Walker et al., 2009), and the proposed base of the Middle Holocene (Walker et al., 2012), both in ice cores. A case might therefore be made for an Anthropocene GSSP in a deep-ocean core. However, deep-ocean deposits often have low sedimentation rates limiting time-stratigraphic resolution, possess few terrestrial fossils, commonly show reduced preservation of calcareous microfossils, and are generally bioturbated, blurring the

stratigraphic record in detail. Nevertheless, GSSPs have been located in relatively deep-water facies now preserved in terrestrial cliff sections, as in the Gelasian and Calabrian GSSPs of the Quaternary (Head and Gibbard, 2015). Zalasiewicz et al. (2014c) recognised two distinct deep marine systems, the clastic wedges of turbidite fans and contourites that occur adjacent to the continents, and the slowly accumulating deep-sea oozes that lie beyond. In the modern deep ocean, both of these environments may be affected by human activity, directly or indirectly. Anthropogenic modification of river systems, notably through dam construction on major rivers (e.g. Syvitski and Kettner, 2011), and trawling on the upper continental slope (e.g. Martin et al., 2015), directly affect sediment supply, as does, indirectly, global warming and resultant rising sea-level. In the abyssal depths the main impacts are: litter, especially plastics (Ivar do Sul and Costa, 2014; Zalasiewicz et al., 2016a); organic and inorganic chemical contamination, including radiogenic fallout signals; effects associated with atmospheric CO₂ increase and warming such as variations in pH, dissolved oxygen content and $\delta^{13}\text{C}$; and resultant biological changes (Zalasiewicz et al., 2014c, Table 2). Sediment accumulation rates in the deep oceans are typically on the scale of 1–4 mm kyr⁻¹ (Tyson and Pearson, 1991), with the lowest rates at abyssal depths farther from shore. In such distal settings, the Anthropocene would be sub-millimetric in thickness, with the Holocene represented by just ~10–50 mm.

A suitable marine environment for a GSSP would include rhythmic varves associated with seasonal variation in sediment input that provides exceptionally high temporal resolution (Schimmelmann et al., 2016). For GSSP consideration, such varved successions would need to show little or no disturbance through storm or current activity, bioturbation or trawler fishing, but at the same time be characterised by sufficiently high rates of sedimentation to accumulate an adequate thickness of strata to represent the Anthropocene. Marine anoxic basins appear to be potentially suitable environments, where varve

preservation is favoured by oxygen-deficient bottom waters, associated with thermal/or chemical density stratification and enhanced organic accumulation at depth through increased primary productivity in the photic zone, either through coastal upwelling or eutrophication (Schimmelmann et al., 2016).

Varved marine anoxic basin successions can be correlated laterally over tens to thousands of kilometres and can record changes to large-scale ocean and atmospheric circulation patterns (Schimmelmann et al., 2016). Marine varved successions are logistically difficult to collect and as a consequence few modern records have received detailed multiproxy studies including varve counting in coordination with radiometric dating (Schimmelmann et al., 2016).

Marine anoxic/dysoxic basins are mainly semi-enclosed basins where the water mass is isolated from the main shelf, typically below ~100 m and the reach of storm mixing of shelf waters. This can occur in depressions on the continental shelf (e.g. the Santa Barbara and Cariaco basins), in fjords, (e.g. Saanich Inlet and Saguenay Fjord) or in enclosed seas, (e.g. the Black and Baltic seas), but areas with periodic anoxia can also occur on an open shelf (e.g. Middle Atlantic Bight, northern Gulf of Mexico, Namibia, Peru) (Tyson and Pearson, 1991, Fig. 3), or on a continental slope on which impinges a well-developed anoxic oxygen-minimum zone (southeastern Pacific off Peru, north-western Arabian Sea).

De-oxygenated dead zones currently occur over ~245,000 km² of the coastal oceans (Diaz and Rosenberg, 2008), or about 0.7% of the sea floor. The number of these dead zones has approximately doubled each decade since the 1960s (Diaz and Rosenberg, 2008). About half of them are the consequence of eutrophication through runoff of fertilisers into the sea, increased primary productivity and enhanced organic flux to the sea bed (Diaz and Rosenberg, 2008), and so they tend to occur offshore or downriver from populated areas (Fig. 3). Most dead zones are currently seasonal, tending to occur in the summer after spring

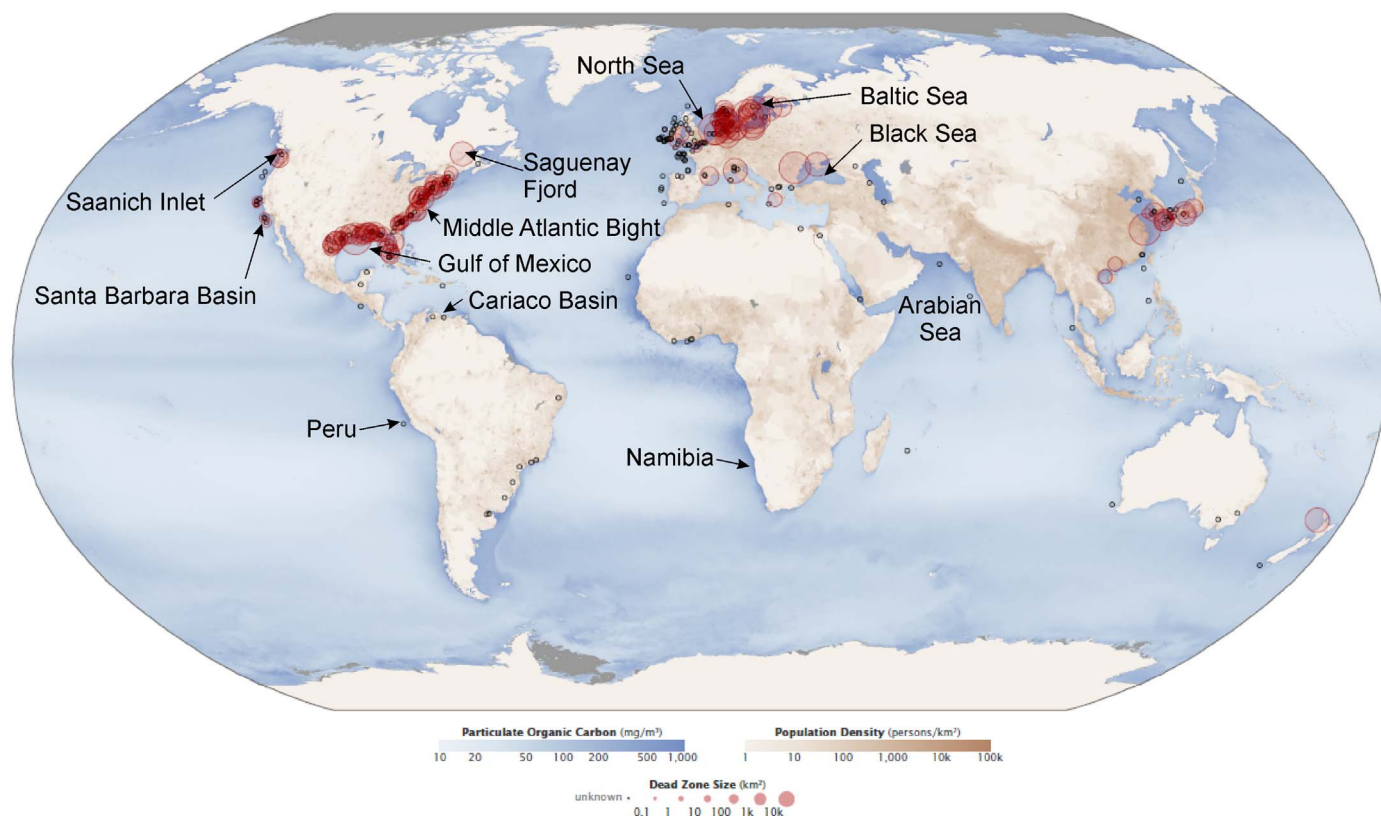


Fig. 3. Location of marine dead zones (from NASA Earth Observatory <https://earthobservatory.nasa.gov/IOTD/view.php?id=44677>; Aquatic Dead Zones generated 17th July 2010). Red circles show the location and size of the dead zones. Black dots show where dead zones of unknown size have been observed. The distribution commonly occurs adjacent to populous land areas (shown by the brown scale), but not to upwelling zones (shown by concentration of particulate organic carbon, in blue scale).

blooms, when the water is warmest and stratification is strongest (Diaz and Rosenberg, 2008). Autumn storms bring a temporary return of benthos, disrupting seabed stratification. Only ~8% of dead zones are persistent (Diaz and Rosenberg, 2008), these areas experiencing the least sediment bioturbation, though in open oceans these seem not yet to have developed varves. Sediments of modern dead zones are a local expression of Anthropocene-related conditions, and identifiable by changes in geochemical character and fauna (Diaz and Rosenberg, 2008), but they seem unlikely to provide good potential for candidate GSSPs, as they typically rest on the bioturbated deposits of the earlier more consistently oxygenated sea floor.

Litter from ships is now seen in most sea floor surveys (Ramirez-Llodra et al., 2011). Clunker, from coal burnt to power steam-ships, was extensively dumped on the sea floor from ~1800 to ~1950 CE (Ramirez-Llodra et al., 2011), and may form a useful signal for latest Holocene strata. Much power station fly-ash and clinker has also been dumped at sea, e.g. Blyth power station ash into the North Sea, UK (Bamber, 1980). Plastics and aluminium largely date from after 1950 CE, with land-derived microplastic beads and plastic fibers now a near-universal component of low-energy modern marine sediments (Ivar do Sul and Costa, 2014; Zalasiewicz et al., 2016a). With about 4.9 billion tons discarded (which is 60% of all plastics produced since 1950) in geologically unstable waste dumps to uncontrolled deposition in terrestrial, aquatic and marine environments (Geyer et al., 2017), the quantity of plastic debris entering the oceans was estimated at 4.8–12.7 million tonnes in 2010 CE (Jambeck et al., 2015) and might strongly increase in the future. However, little work has been done to profile its stratigraphic distribution in marine sediment cores.

Table 4 summarises the key advantages and disadvantages of marine anoxic basin deposits for hosting a potential GSSP. Well-studied examples (explored below) include the Santa Barbara Basin (California), Black Sea, Saanich Inlet (Canada), Saguenay Fjord (Canada) and Cariaco Basin (Venezuela).

3.2.1. Santa Barbara Basin, California

The Santa Barbara Basin lies 100 km west of Los Angeles, USA (Fig. 3), at depths of up to 590 m. Varves comprise dark terrigenous laminae derived from greater winter river runoff alternating with paler summer bloom laminae with diatoms (mainly *Chaetoceros*), planktonic foraminifera, radiolaria and silicoflagellates (Koide et al., 1975; Field et al., 2006; Barron et al., 2015), reflecting seasonal NW winds that cause upwelling and increase biogenic productivity (Barron et al., 2015). Varves have been dated using ^{210}Pb (Krishnaswami et al., 1973; Koide et al., 1975) and radiocarbon AMS (Schimmelmann et al., 2013). Before 1700 CE some drought years did not produce varves, with intermittent oxygenation events causing bioturbation, but from the 18th century to the present the lamina have been shown to be consistently annual and without gaps (Hendy et al., 2013; Schimmelmann et al., 2013). Strata below the upper ~1 cm are highly anaerobic so there is little bioturbation of the laminae or leaching of Pu (Koide et al., 1975).

Table 4

Reasons for and against using a marine anoxic basin deposit as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> • Undisturbed laterally extensive varves (below storm base) correlatable within basins • Independent ^{210}Pb and ^{14}C dating • Anoxic/dysoxic/hypoxic, no/little benthic bio/anthroturbation • Modification of fluvial input to oceanic basins • Clunker ash from steam-powered ships ~1850–1950 CE; microplastics signal post-1950 CE • Coastal eutrophication since mid-20th century • High organic and clay components scavenge contaminants from water column e.g. metals, radionuclides and POPs • Global radionuclide signal (e.g. ^{239}Pu, ^{14}C) • Planktonic foraminifera and diatoms as bio/geochemical signals 	<ul style="list-style-type: none"> • Restricted extent (~0.7% of oceans) • Possibility of missing/additional laminae in near-coastal settings due to turbidite events • Thin strata, e.g. last 75 years < 20 cm for the Santa Barbara Basin and 75 cm for Saanich Inlet • Potential decadal time delay for Pu/metal contaminants to reach sea-bed (but settling velocity of ~2000 m yr^{-1} if present as grains, see Kowalski et al., 2016) • Difficult and costly environment to collect samples without disturbance of youngest laminae

A distinct 1841 CE bioturbated layer is associated with a flood and turbidite event (Hendy et al., 2013). There are no flood event deposits younger than 1930 CE (Krishnaswami et al., 1973), as dam construction along rivers feeding into the basin began in 1912 CE, limiting sediment supply. Despite this, there has been substantial sediment flux since 1928 CE (Krishnaswami et al., 1973, Fig. 4b), with a post-compactional thickness of ~0.5–3 mm per lamina, the diatom ooze lamina component being up to 2 mm thick (Schimmelmann et al., 2016).

The high organic carbon (3–5%) and clay fractions of the sediments scavenge heavy metals, including Pu, from seawater. The $^{239} + ^{240}\text{Pu}$ concentrations initially rise in 1945–1949 CE varves, exceeding background values in the 1950–1954 CE varves (Fig. 4a), and continuing to rise into the 1970s (latest data available), with no sign of a distinct 1963–1964 CE bomb spike (Koide et al., 1975), perhaps reflecting the slow settling of Pu in deep water. Radiogenic particles settle only slowly through the water column, at 12.5 m yr^{-1} in the western Pacific (Livingston et al., 2001). But when such particles are bound within marine snow/faecal pellet aggregates they sink more quickly, at an estimated 74–39 m day^{-1} in the Santa Barbara Basin (Alldredge and Gotschalk, 1988), allowing rapid transmission to the sea-floor. The continued rise of the Pu signal after the 1963–1964 CE bomb spike likely reflects reworking of terrigenous fallout into the basin (Koide et al., 1975), suggesting the first $^{239} + ^{240}\text{Pu}$ signal above background, rather than a peak value, as a possible primary marker for the Anthropocene (Zalasiewicz et al., 2015). An ancillary radiogenic signal of ^{238}Pu , sourced from the Snap 9A satellite burnup in 1964 CE, does show a 1964 CE peak (Koide et al., 1975, Fig. 4a), suggesting both little mobility of Pu within the sediment and little reworking of that radioisotope.

Heavy metal signals in the basin are complex. A rise in Pb concentrations from 1931 CE (Fig. 4c) is due to airborne particulates from leaded gasoline, derived from increased automobile use in Los Angeles, with levelling off from the 1970s representing introduction of unleaded gasoline (Schmidt and Reimers, 1991). Cu values increase notably from about 1960 CE, probably industrially sourced via sewage outfalls (Schmidt and Reimers, 1991). Reduced anthropogenic input, in the case of Cd since the 1970s, may reflect improved sewage treatment (Schmidt and Reimers, 1991).

Among organic geochemical signals, a significant oil spill from a production well in 1969 CE increased Total Organic Carbon (TOC) values, generating a distinctive hydrocarbon signature (Hendy et al., 2015). Polychlorinated biphenyls (PCBs) appeared in the basin sediments about 1945 CE, the DDT pesticide breakdown product DDE [1,1-dichloro-2,2-bis(p-chlorophenyl) ethylene] first appearing ~1952 CE. Concentrations of both show a progressive increase through to 1967 CE (Hom et al., 1974); they may be sourced from sewage outfalls and surface runoff, while DDE can also arrive via atmospheric fallout (Hom et al., 1974).

The basin is located close to the boundary between the subarctic and subtropical gyres, the planktonic foraminiferal assemblages

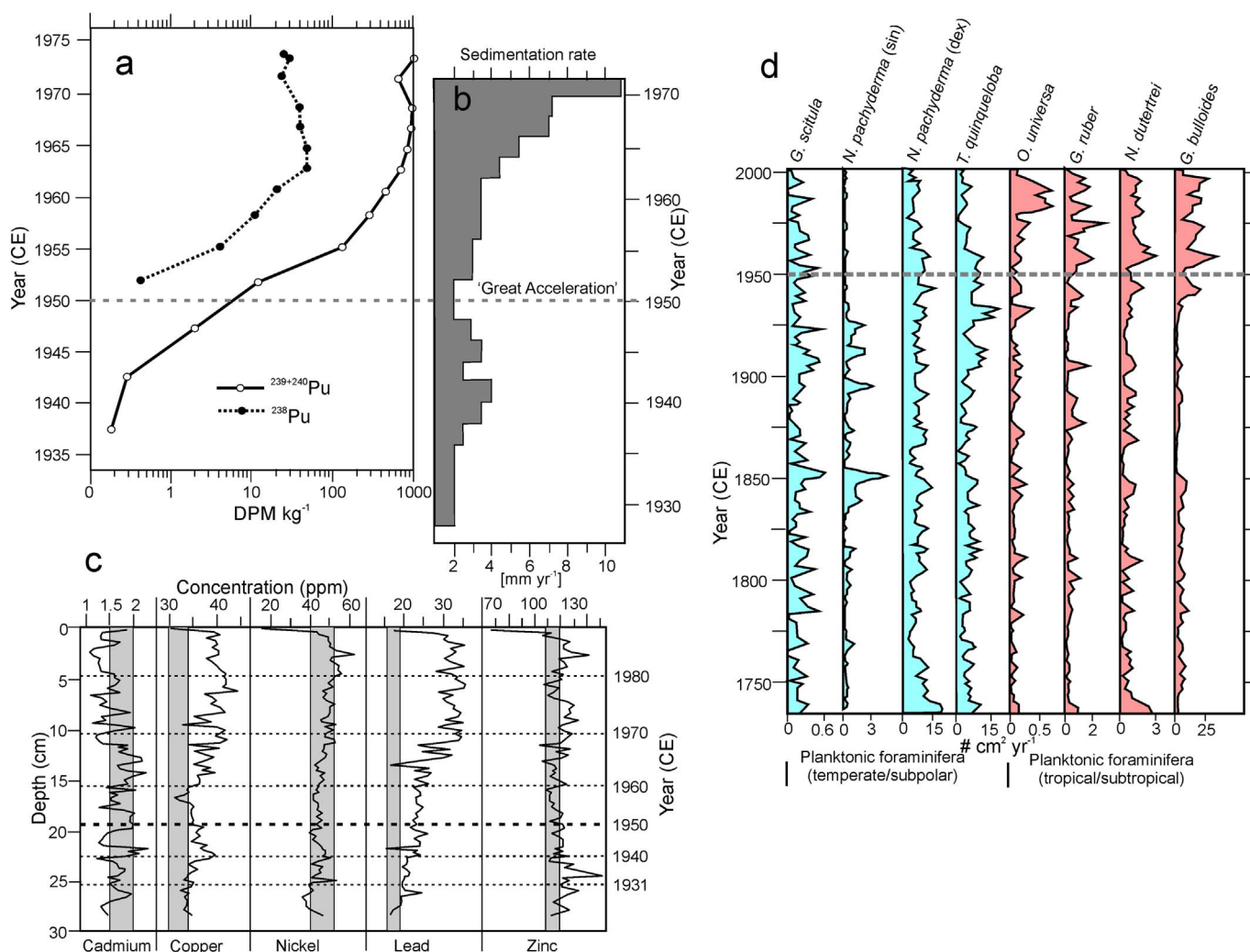


Fig. 4. Key signals in marine cores from the Santa Barbara Basin, with a) Pu signal (Koide et al., 1975), b) variations in sedimentation rates (Krishnaswami et al., 1973), c) selected heavy metals (Schmidt and Reimers, 1991) and d) planktonic foraminifera (Field et al., 2006).

reflecting the dominance of one or other of these gyres (Barron et al., 2015). In a 1400-year dataset, decadal-scale fluctuations (the Pacific Decadal Oscillation) dominate, the two assemblages co-varying until 1959 CE (Fig. 4d), after which there is increasing abundance of tropical/subtropical species, especially after the mid-1970s (Field et al., 2006). A complementary decrease in temperate/polar species in the late-20th century indicates the penetration of deep warm waters (Field et al., 2006). Evidence of such pronounced warming is not seen earlier in the record, not even during the Medieval Climate Anomaly (Barron et al., 2015). This warming is thought to reflect widespread change in the Northern Pacific in the mid-1970s, with greater cyclonic activity of the Aleutian Low Pressure System and warming in the eastern North Pacific (Field et al., 2006). Observed decreases in algae, zooplankton (which show 80% decline between 1951 and 1993 CE), fish and seabird abundance have also been related to this temperature increase (Roemmich and McGowan, 1995).

3.2.2. Black Sea

The Black Sea is an enclosed silled basin with an area of 422,000 km² and a maximum depth of 2212 m (Florea et al., 2011). The water column is stratified, with the pycnocline and halocline both occurring at about 75 m depth in the centre of the basin and 200 m at the margins, and marking the boundary between oxygenated water above and euxinic, H₂S-saturated and saline deeper waters below (Arthur et al., 1994; Florea et al., 2011). At the seafloor, there is a

topmost gelatinous flocculent surface layer about 2 cm thick overlying a further 2 cm of unlaminated grey/white carbonate-rich sediment (Pilskaln and Pike, 2001). The underlying unit (Unit I) is 25–45 cm thick and represents 1633 ± 100 years of sedimentation, based upon varve chronology (Arthur et al., 1994). This unit, present in the deeper parts of the basin, contains varves averaging 0.2 mm thick, consisting of thicker white carbonate-rich summer/autumn laminae with abundant *Emiliania huxleyi* coccoliths alternating with thinner grey clastic-rich laminae deposited from spring and winter river runoff (Arthur et al., 1994; Pilskaln and Pike, 2001; Schimmelmänn et al., 2016). Black organic-rich laminae with biogenic metal sulphides occur between the carbonate and terrigenous laminae (Pilskaln and Pike, 2001; Schimmelmänn et al., 2016). The top flocculent layer is thought to be a permanent transition zone that hydraulically sorts particles by density (Pilskaln and Pike, 2001).

The varves have been independently dated using ²¹⁰Pb, ¹³⁷Cs and ¹⁴C AMS, although the latter commonly gives older ages due either to the presence of “dead” carbon in Black Sea sediments or missing varves (Arthur et al., 1994; Pilskaln and Pike, 2001; Florea et al., 2011). Although the highest countable varves beneath the flocculent layer date to about 1918 CE (Arthur et al., 1994; Florea et al., 2011) discerned the 1986 CE Chernobyl ¹³⁷Cs signal at between 5 and 10 mm depth (in samples collected in 2004 CE), which agrees with a ²¹⁰Pb-dated accumulation rate of 0.49 ± 0.03 mm yr⁻¹. Groups of varves in Unit I have been correlated across the entire deeper basin (Arthur et al.,

1994), with no sign of erosion by interbedded cm- to dm-scale fine-grained turbidite layers (Arthur et al., 1994).

3.2.3. Saanich Inlet, Canada

Saanich Inlet is a 24 km long, up to a ~240 m deep, silled fjord in Vancouver Island, Canada, in which the water column shows stratification by density, salinity and temperature, with deep anoxic water containing hydrogen sulphide (Schimmelmann et al., 2016). Bottom water renewal is restricted to brief episodes of coastal upwelling during late summer, with H₂S otherwise present in water as little as 120 m deep (Anderson et al., 1989). Maximum river discharge and minimum salinity occur during summer months (McQuoid and Hobson, 1997). Modern sediments include varves in which pale diatomaceous laminae (1.5–17 mm thick), with distinct seasonal assemblages, alternate with terrigenous silty clay laminae (0.5–6 mm thick) deposited in winter (Dean et al., 2001; Dean and Kemp, 2004). The varves have been independently dated using ²¹⁰Pb, ¹³⁷Cs and ¹⁴C (Matsumoto and Wong, 1977; McQuoid and Hobson, 1997). Laminated intervals during the Holocene are punctuated by massive debris flow event beds (Dean et al., 2001). Laminae dated from 1950 CE have been recognised at a depth of about 75 cm (McQuoid and Hobson, 1997).

Biostratigraphic indicators include the invasive diatom *Rhizosolenia setigera*, abundant in autumn blooms (Dean et al., 2001), and introduced in the early 1940s through the arrival of aquaculture species and/or from ship ballast waters, both potential sources increasing in the 1930s and 1940s (McQuoid and Hobson, 1997). This invader partly supplanted the species *Thalassiosira gravida* and *Chaetoceros diadema*, which were more abundant prior to the 1930s to 1940s, the assemblages also reflecting increased logging, farming and urbanization, especially from the 1930s (McQuoid and Hobson, 1997). Sinking rates of these organisms, when present in aggregates, may reach 50 to 100 m day⁻¹ (McQuoid and Hobson, 1997 and references therein) meaning that the biotic signal in varves is highly responsive to seasonal changes in abundance in the surface waters. The stratigraphic record shows an abrupt regional climate shift in 1976/1977 CE, with the modern Pacific Decadal Oscillation changing to 22-year periodicity, in contrast to ~15-year periodicity recorded over the previous 2100 years (Dean and Kemp, 2004).

Heavy metal residence times in the water column of the inlet are an estimated 0.08 years for ²¹⁰Pb, 0.07 years for stable Pb, 0.3 years for Cu, 3 years for Cd, and 2 years for Hg (Matsumoto and Wong, 1977). This indicates a near-synchronicity of the influx of Pb into the inlet and its subsequent accumulation in the sediments.

3.2.4. Saguenay Fjord, Canada

The Saguenay Fjord, Quebec, Canada, is a submarine valley, 93 km long and 1–6 km wide which links to the St. Lawrence Seaway via a shallow, 20 m deep sill (Gagnon et al., 1997). The fjord comprises two distinct basins with a deeper and larger landward basin up to 275 m deep and a seaward basin up to 250 m deep (Gagnon et al., 1997). The succession is dated with ²¹⁰Pb and ¹³⁷Cs (Loring et al., 1983; Smith and Levy, 1990). Sediment accumulation rates in the landward basin are 0.2–0.4 cm yr⁻¹ (Smith and Walton, 1980). The deeper waters are generally saline, well-mixed and oxygenated with the redox boundary present a few centimetres below the sediment surface (Gagnon et al., 1997), but organic-rich effluent from pulp and paper mills produces anoxic, unbioturbated sediments at the head of the fjord (Smith and Levy, 1990), making that location more suitable as a potential GSSP target locality.

The Saguenay Fjord has been affected by industrial and municipal waste discharges since the 1930s (Loring et al., 1983; Smith and Levy, 1990). Very high Hg concentrations, and significant Zn, Pb, V and Cd contamination are evident in fjord sediments dating from 1948 CE, with peak values in the early 1960s and early 1970s (Loring et al., 1983; Gagnon et al., 1997). Most of the Hg is bound to organic matter (Loring et al., 1983), limiting Hg diffusion through the sediment (Gagnon et al.,

1997). The oxidized surface sediments inhibit transfer of Hg to the overlying bottom waters, although surface bioturbation may reduce the efficiency of this barrier (Gagnon et al., 1997). Sustained high Hg values suggest resuspension of the Hg by spring runoff or frequent submarine mass flow events (Gagnon et al., 1997). A time lag of a few months at the head of the fjord and 10–15 years at the seaward end (Loring et al., 1983) makes the Hg signal markedly diachronous.

Polycyclic aromatic hydrocarbon (PAH) concentrations gradually increase in 1926 CE above pre-industrial levels, with dramatic increases in the 1940s after major expansion of aluminium production (Smith and Levy, 1990). The PAHs in sediments are mainly scavenged from the water column by organic matter (Smith and Levy, 1990).

3.2.5. Cariaco Basin, Venezuela

The Cariaco Basin off Venezuela comprises two euxinic sub-basins of > 1400 m depth separated by a 900 m-deep saddle. It contains a stratified water column, with a thermocline and a chemocline between 250 and 375 m deep separating oxic waters above from anoxic waters below (Schimmelmann et al., 2016). The basin has been anoxic since about 12.6 ka BP (Hughen et al., 1996). Euxinic bottom waters are associated with sediments including varves (~0.9–3 mm thick, average 1.3 mm) of pale, biogenic (mainly diatoms) winter/spring laminae rich in planktonic opal, carbonate and organic matter accumulated during arid conditions when upwelling develops, and dark, siliciclastic summer/autumn laminae deposited during times of increased runoff, when there is enhanced input of Saharan dust and reduced upwelling (Hughen et al., 1996; Schimmelmann et al., 2016). Microbioturbation is ubiquitous, but generally only slightly disrupts the laminae (Hughen et al., 1996). The varves have been independently dated using ²¹⁰Pb, ¹³⁷Cs and ¹⁴C AMS (Hughen et al., 1996). Microturbidites a few mm thick occur throughout, with two distinct turbidites dated to 1897 ± 5 and 1932 ± 4 CE, likely earthquake-triggered, with the older turbidite possibly having eroded a few centimetres of underlying varves (Hughen et al., 1996; Schimmelmann et al., 2016).

3.3. Coral bioherms, calcified sponges and marine bivalve shells

Shallow-water coral reefs extend over only about 0.1–0.2% of the oceans, and are limited to the tropics, but they contain a large proportion of marine biotic diversity (Fig. 5a). They were first impacted by humans via decreasing diversity of large marine animals beginning around 3.5 ka BP, a trend which accelerated during the Industrial Revolution and especially since 1950 CE (Pandolfi et al., 2003; Hoegh-Guldberg, 2014). Widespread bleaching of tropical coral reefs in response to rising sea temperatures started in 1979 CE and has subsequently increased in frequency and severity (Hoegh-Guldberg, 2014; Hughes et al., 2017). Other significant stressors, including greater turbidity of marine waters due to increased runoff, rises in pollutants, eutrophication and acidification, and severe overfishing, including the dynamiting of reefs by fishermen, have contributed to a 50% reduction in the abundance of reef-building corals over the past 40–50 years. Over the next several decades, this may lead to the collapse of whole reef systems, as happened during mass extinction events of the geological past (Hoegh-Guldberg, 2014).

Although reef systems as a whole are undergoing significant change, indicating their importance for recognising anthropogenic impacts, individual reefs are complex and non-stratiform. Consequently, it is not a reef as a whole that should be considered as hosting a potential GSSP, but one of the long-lived coral bioherms within it that shows annual growth bands. In the Great Barrier Reef, fluorescent banding reflects seasonal-interannual variations in fluvial discharge of organic compounds (Isdale, 1984). The annual laminae can be independently dated using ²¹⁰Pb and ¹⁴C. Key correlatory signals include Pu radionuclide fallout, the ¹³C Suess effect and uptake of pollutants, in particular heavy metals (Fig. 6a).

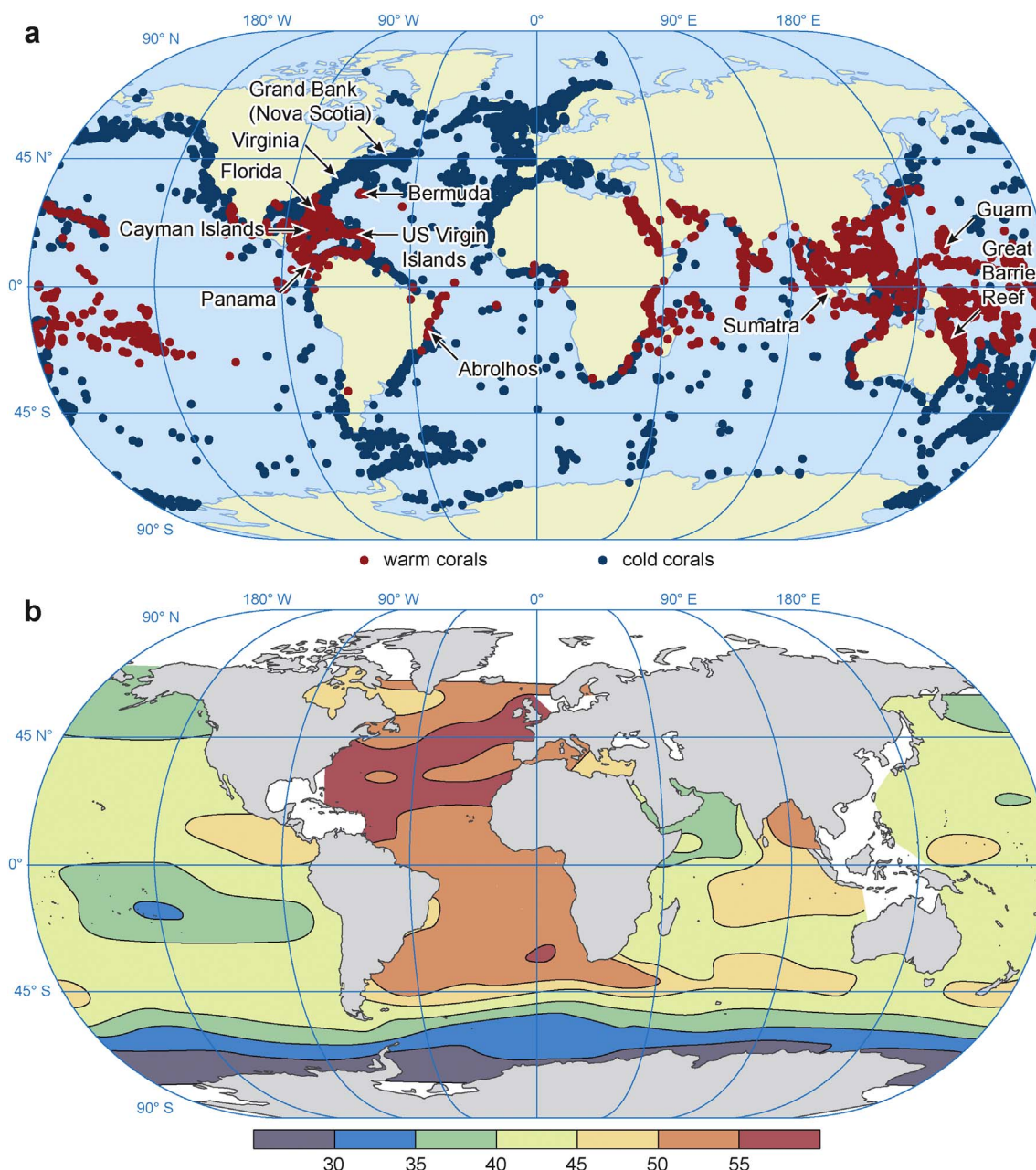


Fig. 5. a) Distribution of shallow-water framework-building coral reefs (from NOAA: Where Are Reef Building Corals Found http://oceanservice.noaa.gov/education/tutorial_corals/coral05_distribution.html and cold-water corals (from Freiwald et al., 2017) and b) the inventory of anthropogenic CO₂ (μmol CO₂/kg) in surface waters (from Swart et al., 2010).

Cold-water growths of aragonitic scleractinian corals and protei-naceous corals are smaller than the tropical reefs, and occupy a range of latitudes and of water depths – from a few metres to abyssal depths (Fig. 5a). Such corals also imprint the Suess effect, radiocarbon and heavy metal signals, and individual cold-water corals can be long-lived. But, those from deep waters contain banding that is not necessarily annual, and ages can only be reliably provided through radiometric dating (Lee et al., 2017). Long-lived bivalves, which also occur in cold waters, also record proxy environmental signals including water temperature and the radiocarbon fallout bomb-spike, though they occupy a smaller depth range (up to ~300 m depth, exceptionally up to 500 m) than cold-water corals.

Calcified sponges (or sclerosponges), found living in water depths up to 145 m, are known to provide good proxy records of water temperature and salinity, secreting their skeletons without some of the vital effects seen in corals and clearly showing the Suess effect on carbon

isotopes (Fig. 6a). They provide records stretching up to 1000 years, much further back in time than many corals (Swart et al., 1998). However, they are very slow growing ($\sim 0.22\text{--}0.27\text{ mm yr}^{-1}$) with living tissue accounting for 3 to 4 years of growth; consequently despite some species showing growth banding, evidence that this is annual is equivocal and ¹⁴C- and ²¹⁰Pb-dating are required to determine ages (Swart et al., 1998). As such, they seem to provide a lower resolution and hence less suitable host for a GSSP than corals or bivalves (see Section 3.3.6).

Table 5 summarises the key advantages and disadvantages of using coral laminae in tropical reef bioherms and cold-water corals and bivalve shells for hosting a potential GSSP. Up to now, research on corals has rarely provided multi-proxy signals, so the examples given below focus on the specific signatures where research has been done.

3.3.1. Caribbean ^{13}C Suess effect and heavy metal concentrations

Scleractinian corals mirror changes in the anthropogenic CO_2 inventory of surface oceans (Fig. 6a). In a global comparison of rates of change of $\delta^{13}\text{C}$ in coral skeletons between 1800 and 2000 CE (Swart et al., 2010), 64% showed a marked decrease from the mid-20th century, attributed to the addition of anthropogenically derived CO_2 (the ^{13}C Suess effect). The decrease was greatest in areas of largest input of anthropogenic CO_2 (Fig. 5b) and, in Florida and elsewhere in the Caribbean, from 1960 to 1990 CE, was comparable to the carbon isotopic decrease in atmospheric CO_2 (Swart et al., 2010, Fig. 6a). A marked inflection of $\delta^{13}\text{C}$ values is seen at about 1955 CE in Atlantic corals and sclerosponges, comparable to the $\delta^{13}\text{C}$ inflection seen in Law Dome ice core from the Antarctic (Fig. 6a). The rate of change is more variable and less pronounced in the Indian and Pacific oceans (Swart et al., 2010). There is an average air–sea equilibration time of about 1 year for the anthropogenically derived CO_2 (Key et al., 2004), resulting in a minimal signal lag in surface waters. However, anthropogenic CO_2 concentrations decrease with depth, and at about 1000 m water depth the signal is only discernible in the North Atlantic (Key et al., 2004), suggesting either a significant lag or lack of transport to these depths. Shallow-water corals thus provide a more effectively correlatable stable carbon isotope signal than do records from deep marine environments such as deep anoxic basins.

Heavy metal concentrations in Panama tend to be higher in near-shore coral reefs and also in certain species such as *Porites furcata*, and

show greater concentrations than in nearby sediments (Berry et al., 2013). Humans have increased the flux of Pb by at least a factor of 10, with rapid and widespread dispersal of Pb into the atmosphere as aerosol particles, and a short residence time of two years within ocean waters (Boyle et al., 2014), which makes Pb potentially suitable as a stratigraphic marker. Of the oceans, the North Atlantic Ocean is most affected by Pb emissions from the Industrial Revolution and subsequently it is this region that has seen most research (Boyle et al., 2014). Corals in Bermuda (Kelly et al., 2009, Fig. 7c) show an initial rise of Pb following exploitation of the Upper Mississippi Valley lead-zinc ore in the USA, with a peak in lead mining in 1840–1848 CE, followed by a clear increase that began in the late 1940s due to greater consumption of leaded gasoline, peaking in the 1970s following US environmental legislation (the US Clean Air Act of 1970, Boyle et al., 2014). Pb in gasoline from the USA typically has $^{206}\text{Pb}/^{207}\text{Pb}$ ratios > 1.17 as compared to European Pb with ratios < 1.15 (Boyle et al., 2014). This is expressed in the Bermuda corals, although the signal is complex, reflecting the different histories of leaded gasoline consumption and legislation in the USA and Europe (Bollhöfer and Rosman, 2001). Higher $^{206}\text{Pb}/^{207}\text{Pb}$ ratios seen in Bermuda until the late-1970s (Fig. 7d) are considered sourced from US gasoline, while the markedly lower ratios in the 1980s and 1990s reflect a European source (Kelly et al., 2009). Pb settles only slowly into deep oceanic waters, as seen off Bermuda (Boyle et al., 2014), but this is not a significant factor in contamination of corals at shallow water depths.

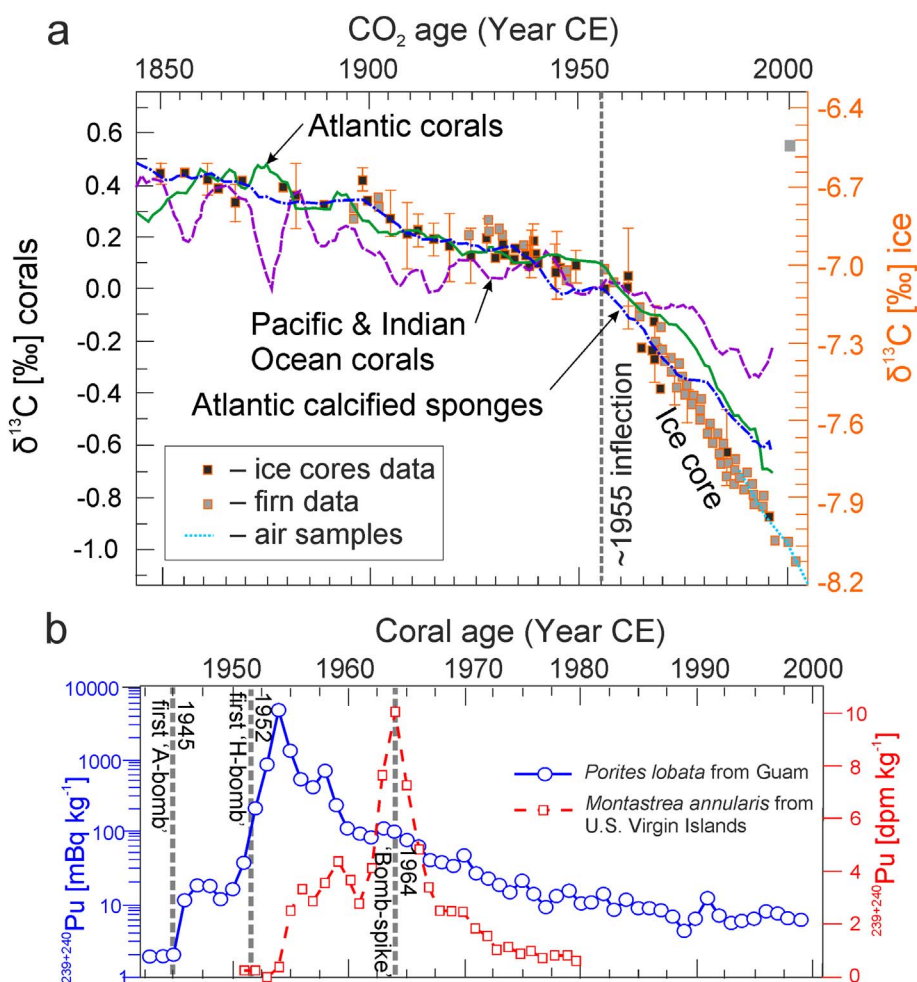


Fig. 6. a) Changes in $\delta^{13}\text{C}$ with respect to age for corals from the Atlantic and the Pacific/Indian oceans and Atlantic calcified sponges, averaged after removing the mean $\delta^{13}\text{C}$ value of the coral skeleton from 1900 to 2000 CE and shown as a five-year running mean (Swart et al., 2010). This is compared with $\delta^{13}\text{C}$ data from Law Dome ice core (Rubino et al., 2013) which show a ~1955 CE inflection; b) $^{240} + ^{239}\text{Pu}$ concentrations in annual growth bands from *Porites lobata* in Guam (Lindahl et al., 2011) and *Orbicella* (*Montastrea*) *annularis* in the U.S. Virgin Islands (Benninger and Dodge, 1986); dpm kg $^{-1}$ = decays per minute per kilogram; mBq kg $^{-1}$ = millibecquerel per kilogram.

Table 5

Reasons for and against using a coral or bivalve shell as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> ● Seasonal growth bands, with large growth rates in shallow-water corals ● Independent dating using ^{210}Pb and ^{14}C ● Microplastic ingestion in corals, e.g. Great Barrier Reef ● Temperature recording ($\delta^{18}\text{O}$ and Sr/Ca) in corals and bivalves; bleaching events in corals beginning 1979 CE ● $\delta^{13}\text{C}$ Suess effect mirrors anthropogenic CO_2 ocean inventory ● Shallow water corals with rapid uptake of radionuclides and contaminants ● Heavy metal enrichment is coral-species specific and more sensitive than adjacent sediments 	<ul style="list-style-type: none"> ● Shallow, warm-water corals limited to tropics; deep-water corals and bivalves provide greater geographical spread ● Reefs are complex structures; signal defined on individual coral or bivalve ● Deep-water corals are affected by a ~ 25 year lag in $\Delta^{14}\text{C}$ and Pb records compared with shallow-water corals ● No GSSP precedent of using living/deceased coral or bivalve

The Pb signal shows global heterogeneity. In the Indian Ocean, significant Pb contamination started in the western Sumatra region around the mid-1970s, with increasing Pb/Ca ratios and decreasing $^{206}\text{Pb}/^{207}\text{Pb}$ and $^{208}\text{Pb}/^{207}\text{Pb}$ ratios around that time, trends that continue into the 21st century (Lee et al., 2014). Pb emissions from China and SE Asia continue to increase (in contrast to the pattern in areas fringing the North Atlantic), sourced from coal consumption (Boyle et al., 2014).

3.3.2. Guam (Pacific Ocean) v. Caribbean Pu radionuclide signals

A natural coral archive in *Porites lobata* from Guam Island (NW Pacific Ocean), dated independently using $\delta^{18}\text{O}$ cycles to produce a biweekly to monthly chronology between 1787 and 2000 CE, was retrieved in a 273 cm core (Lindahl et al., 2011). Between 1943 and 2000 CE the average growth rate was $\sim 15 \text{ mm yr}^{-1}$ (Lindahl et al., 2011), significantly higher than most natural accumulation rates in marine and lacustrine basins. From 1943 to 1945 CE, ^{239}Pu (but not ^{240}Pu) was detected in small concentrations (average of $1.96 \pm 0.11 \text{ mBq kg}^{-1}$, Fig. 6b), though at levels higher than would be expected from an entirely natural source at that time (Lindahl et al., 2011). This suggests some minor mobility of anthropogenic (post-1945 CE) Pu after incorporation into the coral. In 1946 CE, $^{239} + ^{240}\text{Pu}$ activity concentrations increased by an order of magnitude ($11.3 \pm 2.4 \text{ mBq kg}^{-1}$, Fig. 6b) and continued at low levels until 1951 CE, reflecting fallout from locally tested, low-yield, nuclear devices (Lindahl et al., 2011). A marked bomb-nuclide peak is evident from 1952 to 1960 CE, with an apex in 1954 CE ($4540 \pm 70 \text{ mBq kg}^{-1}$, Fig. 6b), the pattern being closely linked to that of large detonations in the local Pacific Proving Grounds (Lindahl et al., 2011). From 1961 to 1980 CE, there was an exponential decrease of $^{239} + ^{240}\text{Pu}$ activity, but despite Nuclear Test Ban treaties, residual Pu remains in the sea water and continues to be incorporated into coral growth laminae, such that by 2000 CE levels were comparable to those in 1946 CE (Fig. 6b).

The Pu signals seen in a short (30-year) duration record from the coral *Orbicella (Montastrea) annularis* in the U.S. Virgin Islands (Benninger and Dodge, 1986), reflect a location that is distant from atmospheric nuclear detonations. That study did not analyse coral bands from 1950 CE or earlier, so it is not possible to determine if a pre-1952 CE global fallout signal is present. But it did show clear 1959 and 1964 CE peaks in $^{239} + ^{240}\text{Pu}$ that closely correspond to global fallout maxima (Benninger and Dodge, 1986; Fig. 6b herein).

These data illustrate the very high-resolution Pu signal present in coral laminae, with only limited evidence of post-growth mobility of Pu. The example in Guam shows an initial rise in Pu between 1946 and 1951 CE and a peak signal in 1954 CE. Such a signal contrasts with areas distant from testing grounds which show a global Pu fallout pattern (e.g. Waters et al., 2015), with an initial 1959 CE peak and a subsequent higher 1964 CE peak (e.g. Benninger and Dodge, 1986). This variability in peak signals suggest the initial rise in global Pu levels would better mark the basal boundary of the Anthropocene than any of the various peak signals (cf. Zalasiewicz et al., 2015). For hosting a potential GSSP, choosing a coral that was distant from nuclear testing

grounds would mean that the global atmospheric signal is resolved without local perturbations.

3.3.3. ^{14}C , Pb and $\delta^{15}\text{N}$ signals in deep-water Atlantic gorgonian corals

Corals show a ^{14}C bomb-spike that has potential as a marker (Sherwood et al., 2005a; Fig. 7a herein). Compared to coral $\Delta^{14}\text{C}$ records from Florida in the North Atlantic, the inception of the Brazil (Abrolhos Archipelago) record is delayed by ~ 1 year during the late 1950s with peak $\Delta^{14}\text{C}$ values reached from 1970 to 1972 CE in Florida and at 1974 CE at Abrolhos (Druffel, 1996), suggesting an appreciable lag for the bomb ^{14}C to transfer from the main source latitudes in the Northern Hemisphere. There is also a $\Delta^{14}\text{C}$ atmosphere–surface ocean equilibration time of approximately 10 years, with some resultant time lag in the signal (Key et al., 2004). However, analysis of growth rings in the deep-water gorgonian coral *Primnoa resedaeformis* at depths of 250–475 m off Nova Scotia (Canada) shows a $\Delta^{14}\text{C}$ peak around 1972 CE (Sherwood et al., 2005a; Fig. 7a herein), similar to that in shallow water corals in Florida (USA), suggesting rapid transfer of the radiocarbon to such moderate depths via plankton. Overall, a mid-1950s inception of the signal is evident in gorgonian corals from these environments, providing a reasonable proxy for the beginning of the Anthropocene.

At much greater depths, $\Delta^{14}\text{C}$ records from the coral *Enalllopsammia rostrata* from 1410 m depth off Bermuda first show bomb radiocarbon at ~ 1980 CE (Lee et al., 2017), a delay of some 25 years compared with the atmospheric signal (Lee et al., 2017). In this record, the introduction of anthropogenic lead, evident as an increased Pb/Ca ratio and decreased $^{206}\text{Pb}/^{207}\text{Pb}$ and $^{208}\text{Pb}/^{207}\text{Pb}$ ratios, occurs in the 1990s and hence shows a similar 25-year delay, and the signal is significantly smaller compared with that from nearby shallow-water corals (Lee et al., 2017). Given the decade-scale lags for $\Delta^{14}\text{C}$ in shallow-water corals, with even greater lags in deep-water corals, Pu is here considered the preferred fallout signal in corals.

Primnoa resedaeformis from Nova Scotia (Canada) to Virginia (USA) records the Suess effect, with $\delta^{13}\text{C}$ values decreasing by at least 2‰ since around 1960 CE (Sherwood et al., 2005b). *Primnoa resedaeformis* off Nova Scotia also records a $\delta^{15}\text{N}$ signal, considered a proxy for nitrogen source, which is correlated with increasing subtropical versus subpolar slope waters over the 20th century (Sherwood et al., 2011); $\delta^{15}\text{N}$ decreased by about 1‰ over this interval, with most of the decline occurring after 1970 CE, coincident with a rise in nitrates (Sherwood et al., 2011). Such coral studies show that the persistence of a warm, nutrient-rich regime since the early 1970s is exceptional in the context of the last ~ 1800 years (Sherwood et al., 2011, Fig. 7b). Coral cores from Dongsha Atoll, western Pacific, show markedly lowered $\delta^{15}\text{N}$ values since ~ 2000 CE in coral skeleton-bound organic matter, probably associated with increased deposition of anthropogenic atmospheric nitrogen (Ren et al., 2017). By contrast, coral studies from some Indo-Pacific reefs affected by untreated sewage outfalls show high $\delta^{15}\text{N}$ (and $\delta^{13}\text{C}$) signals (Heikoop et al., 2000), indicating the complex behaviour of the nitrogen stratigraphic proxy.

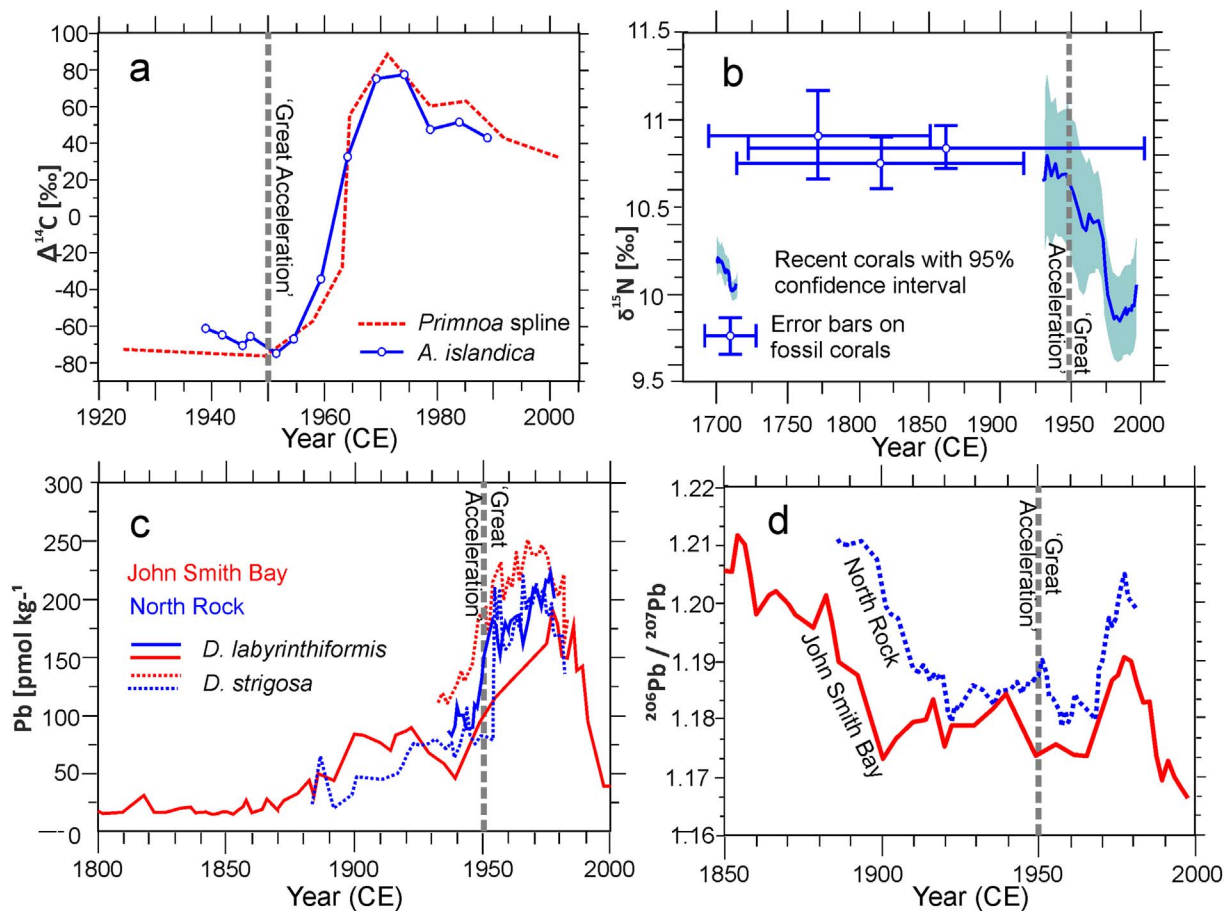


Fig. 7. a) Plot of $\Delta^{14}\text{C}$ vs age for 7 different colonies of the deep-sea gorgonian coral *Primnoa resedaeformis* with spline fit through the data (Sherwood et al., 2005a), and for the bivalve *Arctica islandica* (Weidman and Jones, 1993); b) decreased $\delta^{15}\text{N-AA}$ values in deep-sea gorgonian corals in the NW Atlantic (Sherwood et al., 2011) © (2011) National Academy of Sciences; c) Pb concentration; and d) $^{206}\text{Pb}/^{207}\text{Pb}$ in Bermuda corals (Kelly et al., 2009).

3.3.4. Global temperature and pH proxies from corals in the Great Barrier Reef

In the Great Barrier Reef, Australia, annual band thicknesses in corals equate to growth rate, which can be correlated with sea surface temperatures (SSTs) on a time scale of years to centuries (Lough and Barnes, 2000). During the bleaching events associated with elevated SSTs of the late 20th century, *Porites* colonies from the Great Barrier Reef showed a decline in linear growth of 13.3% between 1990 and 2005 CE (De'ath et al., 2009), with calcification rates falling by 14.2%. Such changes are unprecedented over at least the last 400 years (De'ath et al., 2009).

Corals in the tropical Pacific region have shown a general decrease of $\delta^{18}\text{O}$ values from the late 20th century to the present (Cole, 1996). This reflects large-scale warming in the upper layers of the ocean, or a general freshening of these waters, or a combination of both. $\delta^{18}\text{O}$ signals in scleractinian coral skeletons have been used as SST proxies in other areas, some as long records (e.g. Colonna et al., 1996), or at very high resolution (e.g. Ahmad et al., 2011). However, $\delta^{18}\text{O}$ signals reflect evaporation rates as well as SST, which is why they are frequently used together with other proxies, especially Sr/Ca (e.g. Tierney et al., 2015; Zinke et al., 2016). While some locations show a pronounced temperature increase around 1950 CE (e.g. Hetzinger et al., 2010), coral proxies have also shown that tropical seas might have started to warm as early as the mid-19th century, as an effect of both natural and possibly anthropogenic drivers predating the widely recognised global warming typical of the mid-20th century (Abram et al., 2016). Although temperature reconstructions can be highly refined (DeLong et al., 2016), temperature records from coral skeletons are a mix of local and global factors, and hence better as an ancillary rather than primary

marker for defining any potential coral-based GSSP.

$\delta^{11}\text{B}$ isotopic records in carbonates can provide a proxy for seawater pH. Data from a *Porites* coral in the Great Barrier Reef covering an interval from 1800 to 2004 CE show a recent trend of acidification correlating positively with the $\delta^{13}\text{C}$ Suess effect, and negatively with Mg/Ca ratios (a further proxy for SSTs) from 1940 to 2004 CE, indicating a coincident warming (Wei et al., 2009). Pre-industrial boron isotope signals in the coral show 22- and 10-year cycles strongly controlled by the Pacific Decadal Oscillation (Wei et al., 2009). From 1940 to 2004 CE the $\delta^{11}\text{B}$ signal indicates a decrease in pH of about 0.2–0.3 U, but with marked annual oscillations of 0.5 U around 1940 and 1998 CE (Wei et al., 2009).

3.3.5. Plastics in the Great Barrier Reef

Macroplastics have become noticeable in coral reefs, through the snagging of fishing nets and as plastic debris generally washed into the oceans. Microplastics are increasingly found both within marine waters and in underlying sediments (Zalasiewicz et al., 2016a). These plastics adsorb heavy metals and organic compounds from water, which can bioaccumulate in worms, fish and seabirds (Ivar do Sul and Costa, 2014), and zooplankton (Cole et al., 2013). Microplastic ingestion into scleractinian coral polyps has been recognised recently in the Great Barrier Reef (Hall et al., 2015). Ingestion probably occurs both via microplastic-containing zooplankton and directly from the water. It remains uncertain whether any ingested plastic is transferred into the coral skeleton, as a potentially recognizable feature of future coral growth bands.

3.3.6. North Atlantic marine bivalves

A $\Delta^{14}\text{C}$ bomb signal comparable to that in cold-water corals (Fig. 7a) was recorded in a long-lived bivalve, the ocean quahog (*Arctica islandica*) collected at a depth of 75 m on Georges Bank off New England, USA (Weidman and Jones, 1993), with similar records in haddock otoliths (*Melanogrammus aeglefinus*) from the Grand Banks (Campana, 1997). The $\Delta^{14}\text{C}$ values measured in these and in the gorgonian coral *P. resedaeformis* (see above) (~ -80 to $\sim +80\text{‰}$) are low compared to those in reef corals from Florida (~ -60 to $\sim +160\text{‰}$) where the stratified subtropical waters have a less diluted bomb signal (Sherwood et al., 2005a). *Arctica islandica* from water depths of 3–80 m across the temperate North Atlantic shows a gradient of increasing amplitude of the radiocarbon bomb-spike from Iceland southwards towards the North Sea (Scourse et al., 2012). All localities show a rapid response to the increase in atmospheric $\Delta^{14}\text{C}$ excess, with a lag within the bivalves of as little as 1–2 years, despite the equilibration time between atmosphere and ocean of 7–10 years (Scourse et al., 2012). However, the bivalve data shows a slow response to the decline in the radiocarbon signal, with a diachronous peak that occurs earlier in the south (Scourse et al., 2012).

Annually resolved $\delta^{18}\text{O}$ signals recorded in *A. islandica* collected at a water depth of 80 m north of Iceland show gradually increasing values over the period 953–1891 CE (± 18 years), after which there is a rapid transition to lower values, with the 20th century represented by values that are significantly lower than at any other time in the last 1000 years (Reynolds et al., 2016).

3.4. Estuarine and deltaic deposits

River estuaries and deltas are present on all continents except Antarctica, forming the transition between the fluvial and marine realms (Fig. 8) and representing the zone through which most terrestrial sediments (including those associated with human signals) are transported to the oceans. Estuaries occupy about 10 million km^2 and major deltas 1 million km^2 (Syvitski and Saito, 2007). Most modern estuaries and deltas developed during the Early Holocene sea-level rise, with the flooding of river or glacial valleys. Sea level during the Middle

and Late Holocene was essentially stable, allowing estuaries to silt up and deltas to build seawards (Zalasiewicz et al., 2014c). In addition, early deforestation and new agricultural practices commonly resulted in increased sediment flux to estuaries and deltas (Syvitski and Kettner, 2011). Current and projected absolute sea-level rises are expected to flood both estuarine and deltaic deposits as warming continues to expand ocean water and melt land ice (Giosan et al., 2014). The trapping of sediment behind major dams, constructed on nearly all major rivers in recent decades (Syvitski and Kettner, 2011), has greatly reduced sediment flux to the coast since the 1950s (Walling and Fang, 2003). This change will leave a clear response as a transgressive, diachronous sequence-stratigraphic system with excellent preservation potential (Poirier et al., 2011). One consequence of decreasing fluvial profiles during sea-level rise is lower rates of erosion in the proximal headwaters, resulting in rivers with slower flow regimes, transporting relatively finer sediments to estuaries and deltas and therefore to coastlines. Where rivers have been dammed, even these sediment loads will not reach the coastal ocean. Many urban centres are located on or close to deltas or estuaries, which may therefore experience substantial anthropogenic influence. Deltas and coastal deposits continue to subside under their own weight, the subsidence now being accentuated by the compaction associated with urban development and with the extraction of groundwater and hydrocarbons (Syvitski and Kettner, 2011), while locally drainage has led to the wholesale removal of surface peat layers, further lowering the surface (e.g. Smith et al., 2010) and increasing flooding risk. Twenty four out of the world's 33 major deltas are subsiding, levels of subsidence having reached several metres, as on parts of the Rhine, Nile, Pearl and Yangtze River deltas (Syvitski et al., 2009).

Mud deposition occurs on the continental shelves in front of major estuarine and deltaic areas (pro-deltas) around the world (Fig. 8). They can also form subaqueous deltas disconnected from the source delta, or as scattered mud patches and widespread mud blankets, with a moderate sedimentation rate of $1\text{--}5\text{ mm yr}^{-1}$ (Hanebuth et al., 2015). They represent well-resolved palaeoenvironmental accumulations, commonly initiated during early flooding of continental shelves following deglaciation $\sim 14\text{ ky}$ ago (Hanebuth et al., 2015). One example now forming a sink of anthropogenic pollution, including heavy metals and

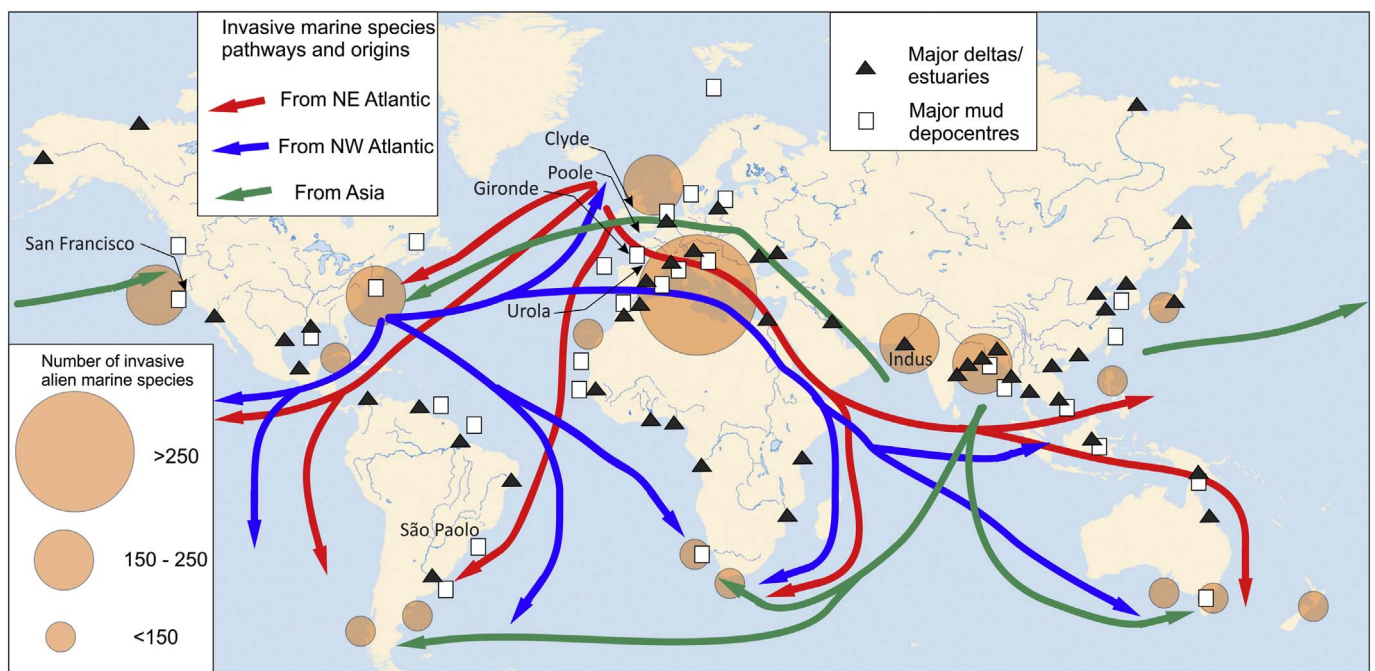


Fig. 8. Distribution of deltas and estuaries (from Tessler et al., 2015), major mud depocentres (from Hanebuth et al., 2015) and areas where neobiota amenable to fossilization have recognizably altered coastal ecosystems (from: Major pathways and origins of invasive species infestations in the marine environment In UNEP/GRID-Arendal Maps and Graphics Library http://www.international-marine.com/invasivespecies/PublishingImages/invasive_vectors_001.png).

Table 6

Reasons for and against using an estuarine or deltaic deposit as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> • May be varved through seasonal variations in sediment flux • Independent dating using ^{210}Pb; ^{14}C more challenging • Radionuclides, no settling delay in water column; clear bomb spike • Micro- and macro-organisms reflect environmental change and commonly preserve as fossils (e.g. foraminifera, diatoms, palynomorphs, molluscs, fish) • Microplastics, direct source from effluents • Heavy metals, close to industrial sources • Persistent organic pollutants signal e.g. chlorinated pesticides, polychlorinated biphenyls • Modification of fluvial input to estuaries related to catchment modification • Samples relatively easy to procure 	<ul style="list-style-type: none"> • Potential of bioturbation • Numerous omission surfaces (natural erosion and anthropogenic dredging) • Limited lateral continuity, but distinct systems may undergo similar histories • Strong modulation of signals by local processes • No GSSP precedent

hydrocarbons, lies off the coast of São Paulo State, SE Brazil (Mahiques et al., 2016). Here, the mud deposits originate from many small river inputs and are mixed with coastal sediments. These deposits form a laterally extensive sedimentary succession, where the mid-shelf section at 50–100 m depth is prone to reworking through bioturbation, storms and anthropogenic disturbance, including trawler fishing and sea-bed dredging (Mahiques et al., 2016). Similar shelf mud deposits occur in the Bay of Biscay, in front of the Gironde estuary, SW France (Lesueur et al., 1996), where three mud-patches, totalling 630 km², occupy palaeo-valleys at depths of 30–75 m. The rhythmically laminated deposits there, which broadly represent storm layers, accumulated to a thickness of 1.8 m over the last 2000 years and the upper 0.2 m accumulated in < 30 years, as shown by ^{14}C and pollen analyses. A marked increase in pollen from *Pinus pinaster* over the last century reflects extensive afforestation (Lesueur et al., 1996).

Estuaries and deltas are particular foci for neobiota (Wilkinson et al., 2014), often associated with marked declines in indigenous species. Widespread translocation of invasive species took place during the 19th and 20th centuries through the establishment of global shipping routes, and resultant transfer of species in ballast water and from hull fouling (Fig. 8). Organisms such as diatoms, dinoflagellates, foraminifera and ostracods, are particularly susceptible to transportation in ballast water and hence are likely to produce novel signals within coastal sediments, especially close to major ports (McGann et al., 2000; Calvo-Marcilese and Langer, 2010). The International Convention for the Control and Management of Ships' Ballast Water and Sediments, which came into force in September 2017, is aimed at reducing the rate of such biotic transfer.

Aquaculture of fish, shellfish and shrimps has become a growing feature of estuarine settings in the late 20th and early 21st century (Martinez-Porchas and Martinez-Cordova, 2012). This has resulted in diverse biotic changes including the displacement of native species, competition for space and food, and the spread of pathogens and parasites, and impacts like these are likely to increase as this kind of fishery develops. In addition, destruction of natural ecosystems, most notably of mangroves along equatorial coastlines, which have seen a 25% decrease in 20 years to 150,000 km² in 2000 CE (Martinez-Porchas and Martinez-Cordova, 2012), has changed species presence/absence and abundance patterns. Aquaculture can also lead to eutrophication causing phytoplankton blooms, including red tides, the resultant anoxia and increased pathogens killing off benthic organisms (Martinez-Porchas and Martinez-Cordova, 2012). The first arrival of many invasive species to local areas is commonly historically documented, but the fossil record of these changes as successive assemblages in recent well-dated sedimentary successions remains poorly studied.

'Bomb spike' calibration and AMS ^{14}C dating have been used to derive a detailed age-depth model for a 1-m long sediment column collected from a salt marsh in Poole Harbour, southern England (Marshall et al., 2007), enabling comparison with chronologies obtained from CRS ^{210}Pb analysis, ^{137}Cs age markers, pollen and

spheroidal carbonaceous particles (SCPs). Little or no agreement was found between the ^{14}C 'bomb spike' dates, the ^{210}Pb chronology, and the ^{137}Cs data, the last of these being affected by local (non-fallout) discharges (Marshall et al., 2007). The study showed an acceleration in sedimentation rates during the last 100 years, and indicated that 'bomb spike' ^{14}C calibration dating may often better constrain salt-marsh sedimentation rates than ^{210}Pb dating (Marshall et al., 2007).

Table 6 summarises key advantages and disadvantages of hosting a potential GSSP within estuarine or deltaic deposits, where research has focused on impacts of anthropogenic pollution through geochemical signatures, especially heavy metals and POPs, and on microbiotic content such as foraminifera, diatoms and palynomorphs (Wilkinson et al., 2014). The examples provided below, for the Clyde Estuary (Scotland), Urola Estuary (northern Spain) and Indus Delta (Pakistan), demonstrate a range of signals. For a variety of reasons (Table 6) these types of environment seem unsuitable for hosting a GSSP.

3.4.1. Clyde Estuary, Scotland (Pb and organic compounds)

Lead concentrations in the Clyde Estuary were initially elevated early in industrial activity, peaking by about 1915 CE, with a prominent decline to historically low concentrations post-1980 CE (Vane et al., 2011, Fig. 9a), as heavy industry declined and Pb use in gasoline was phased out. $^{207/206}\text{Pb}$ isotope ratios of *Sphagnum* moss collected in Scotland from 1838 to 2000 CE record variation of atmospheric Pb aerosols (Farmer et al., 2002): a constant pattern during the 19th century until ~ 1915 CE reflects use of British lead-ore and coal-burning (Fig. 9a); increased ratios from 1920 CE are due to overseas Broken Hill-type Pb additives in gasoline and industrial lead, peaking in the 1980s; and then declining ratios following the banning of lead in gasoline (Vane et al., 2011).

Increased industrialisation of the Clyde Estuary in west Scotland has led to elevated levels of sediment-hosted anthropogenic organic chemicals (Vane et al., 2011), including total petroleum hydrocarbons (TPHs), polycyclic aromatic hydrocarbons (PAHs), polychlorinated biphenyls (PCBs) and brominated flame-retardants (polybrominated diphenyl ethers, PBDEs) (Fig. 9b). The PAHs are largely sourced from combustion of coal, petroleum or wood, with peak concentrations in 1954 CE (Vane et al., 2011, Fig. 9b). Increasing petroleum pollution, mainly from shipping and petroleum refineries, was recorded from about the 1950s by increasing TPH concentrations and a decline in PAHs that reflects decreasing coal use (Vane et al., 2011). This was followed by more modern pollution from the PCBs, which show a prominent onset (1950s), peak (1965–1977 CE) and decline (post-1980s).

The complexity of signals provides a very fine resolution chronometer of pollution in the Clyde Estuary. Although the Pb, TPHs and PAHs provide a local signal, the Pb isotopes and PCBs provide a more widespread and consistent, although not precisely coincident, global signal.

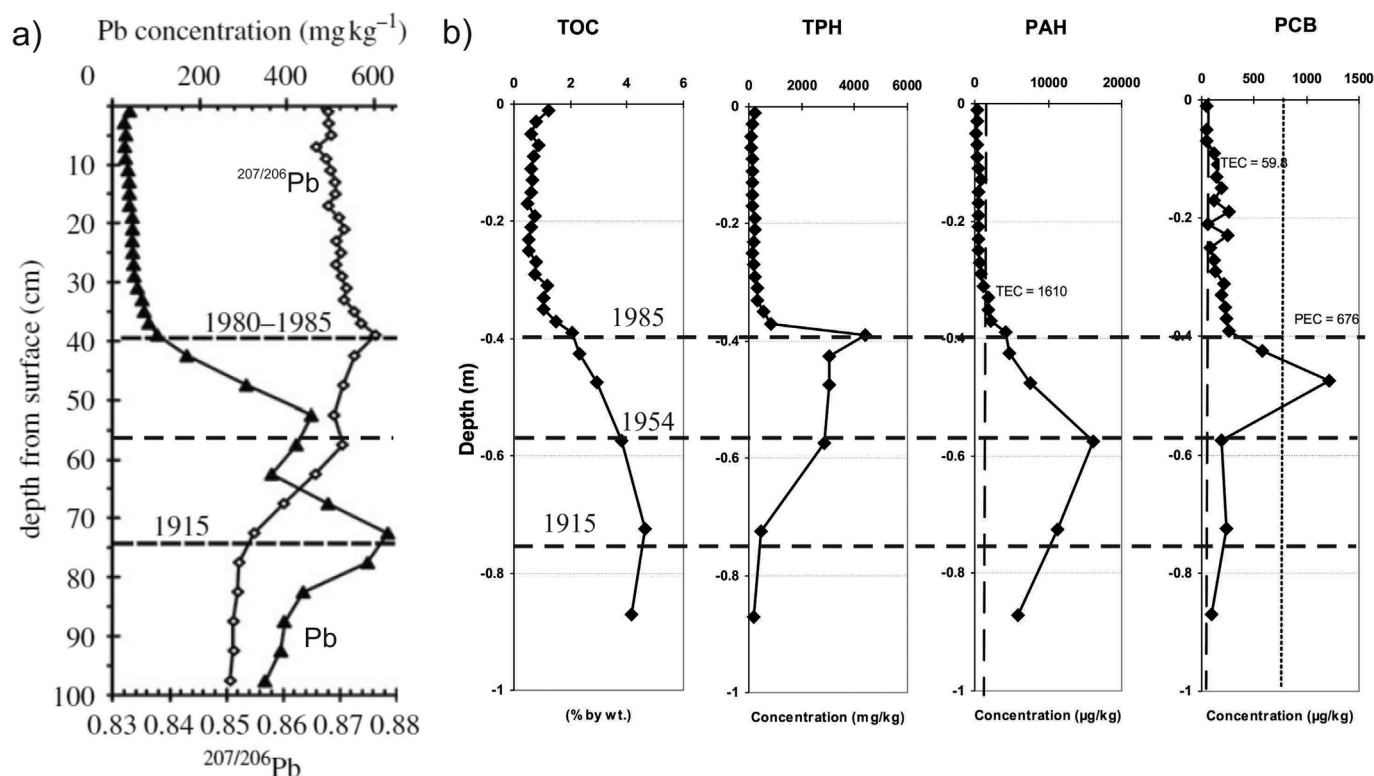


Fig. 9. Example of data from a single core from the Clyde Estuary showing a) Pb concentrations and ^{207/206}Pb isotope ratios, and b) PAH, TPH and PCB organic chemical signatures (Vane et al., 2011). Dates are interpreted.

3.4.2. Urola Estuary, Spain

The lower intertidal part of the Urola estuary (Basque coast, Spain) was studied in a 50 cm core and dated using ²¹⁰Pb, the basal parts dated to ~1850 CE (Goffard, 2016). From the 1960s to the 1980s the rate of sedimentation increased by > 60% in response to growing industrial activity, with further increase from the mid-1990s coinciding with nearby dredging. Micropalaeontological analysis of benthic foraminiferal assemblages (FAs) (Goffard, 2016) identified three assemblages (Fig. 10) that reflect a local cycle of biological deterioration and reduced diversity followed by recovery. This biostratigraphic succession has been correlated with changes in levels of industrially-derived metals in the sediment, showing that chemical deterioration preceded biological deterioration. Longer-range correlation is provided by: spheroidal carbonaceous particles, which occur mainly in the lower levels; iron spheroids, showing enrichment from the 1960s; and plastic microbeads, which increase from the 1980s (Goffard, 2016).

3.4.3. San Francisco Bay, USA

San Francisco Bay, with a well-documented history of invasive species (neobiota), is typical of the many coastal ecosystems around the world that are near shipping ports (Fig. 8). In a seminal study by Cohen and Carlton (1998), some 234 neobiotic species were recognised, ranging from sponges to mammals. They noted an accelerating trend over 145 years, with about half the invasions occurring since 1960 CE. More recently, studies of shallow subtidal sediments in the bay have identified soft-sediment communities dominated by invasive species (Jimenez and Ruiz, 2016). The presence of so many neobiotic species in San Francisco Bay suggests the potential for a high-resolution biostratigraphy using species invasions (Fig. 11). Building such a stratigraphy would need to resolve the myriad complexities of a sedimentary succession heavily disturbed since the 19th century. Nevertheless, sediment cores with Pleistocene and Holocene micropalaeontology are well known (e.g. Sloan, 1992; McGann et al., 2002), providing a pre-human influenced baseline for local coastal assemblages.

A focus of such analysis could be bivalves, which are commonly fossilized, widespread and readily recognizable neobiotic species. A biostratigraphic subdivision could begin with a pre-neobiotic interval that is characterised only by indigenous bivalve species, these persisting into the mid-19th century. A subsequent early neobiotic stage (see also Cohen and Carlton, 1998) might be characterised by the first appearance of the oyster *Crassostrea virginica*, which entered the Bay in 1869 CE. A native to the Atlantic coast of North America, this species was introduced with the development of oyster fisheries. *Crassostrea virginica* was also introduced into the estuaries of the Hawaiian islands at about the same time (1866 CE), providing a potential biostratigraphic tie with a distant island succession (Fig. 11).

The early phase of invasions into San Francisco Bay is followed by increasing indicators of pan-Pacific connections between North America and East Asia, signalled by, for example, the introduction of the East Asian bivalve *Venerupis philippinarum*, first recognised in San Francisco Bay in 1946 CE (Cohen, 2011, Fig. 11), although there are earlier records elsewhere on the Pacific coast of North America. Locally, this species occurs in abundances of 2000 individuals per m², and it has good preservation potential. It has also been widely introduced, from Hawaii (introduced in the 1880s, established by 1918 CE) to Europe. A later colonisation from East Asia was by the Amur Clam *Potamocorbula amurensis*, native to the waters of China, Siberia, Korea and Japan. It was introduced into San Francisco Bay in 1986 CE, likely through ships' ballast water (Fig. 11) and then spread widely and rapidly. It now occurs in huge numbers in muddy substrates within the Bay: up to 48,000 individuals have been recorded per m² sediment (Cohen, 2011).

3.4.4. Indus Delta

The Indus Delta, Pakistan, and its associated river system, is highly managed, with artificial levees, barrages, irrigation canals, upstream reservoir construction and inter-basin water diversions. It has also been affected by watershed deforestation. As a consequence, the river has changed radically, with a current sediment flux of ~13 Mt yr⁻¹ to the

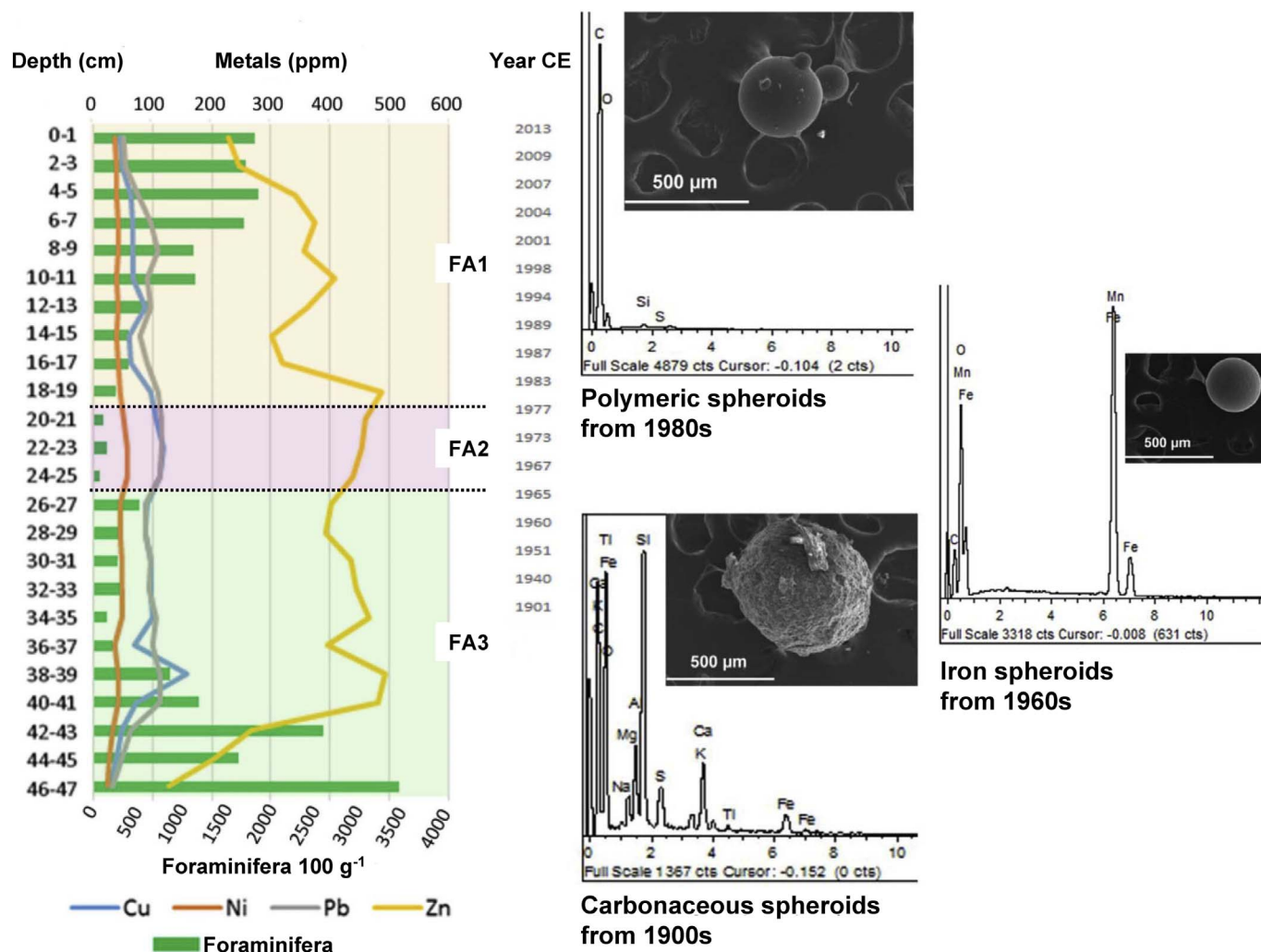


Fig. 10. Metal concentrations (Cu, Ni, Pb, Zn) versus foraminiferal density in the Urola estuary (northern Spain). FA1–FA3 represent three distinct foraminiferal assemblages referred to in the text. Modified from Goffard (2016).

tidal-dominated delta and its single active distributary, in contrast to the much larger flux $> 270 \text{ Mt yr}^{-1}$ of mainly silt that was transported to the fluvial-dominated delta before these management schemes were implemented. The original delta system comprised 17 channels (which still avulsed naturally) in 1861 CE (Syvitski et al., 2014). Modifications started as early as 1762 CE, although constructed dams were all washed away in a flood in 1826 CE. The modern barrage system was initiated in 1859 CE (Syvitski et al., 2014). With rising sea-levels accompanied by tectonic subsidence, the abandoned distributary channels have been tidally reworked since 1944 CE and the delta has lost $12.7 \text{ km}^2 \text{ yr}^{-1}$ of land and $\sim 69 \text{ Mt yr}^{-1}$ of sediment by erosion. The system is highly dynamic due to monsoon-driven floods, storm surges, tsunamis and earthquakes (Syvitski et al., 2014), and hence stratigraphic successions are beginning to emerge as complex backstepping packages of sediment with omission surfaces related to erosive events, in contrast to the pre-1869 CE delta, which was prograding at about 200 m yr^{-1} .

3.5. Lake sediments

The planet has about 117 million lakes $> 2000 \text{ m}^2$, covering about 3.7% of the ice-free land surface, with the highest concentration and area of lakes at boreal and arctic latitudes between 45°N and 75°N (Verpoorter et al., 2014). The most suitable lake environment for hosting a potential GSSP is one where varves are present. Varves tend to occur in lakes in which a flat bottom profile limits sediment flow, and,

ideally, where there is little or no bioturbation, water movement, or gas emission from buried organic material to disturb the laminae (Zolitschka et al., 2015). Many such features are common in glacial lakes, characterised by graded summer silt laminae alternating with winter clays, e.g. Holtgrieve et al. (2011) and Wolfe et al. (2013). Varves can also develop in hypoxic lakes (Fig. 12), typically meromictic lakes with stratified water columns in which sediment is introduced through seasonal input of clastic or biogenic material. Saline lakes can develop varves in response to seasonal precipitation of evaporite minerals, within an environment commonly too hostile for benthic life, allowing preservation of the varves; however, these lakes are more likely to periodically dry out, thus tending to miss annual laminae. Lakes are already recognised as suitable locations for hosting a GSSP, having been used as the auxiliary GSSP locations for the base of the Holocene (Walker et al., 2009).

Hypoxia has been recorded in sediments of 365 lakes worldwide (Fig. 12). Of these, 71 ($\sim 20\%$) show hypoxia developed since the mid-19th century CE, much earlier than the widespread development of hypoxia in coastal zones during the mid-20th century (Jenny et al., 2016). This reflects the sensitivity of lakes to environmental change, where increasing human activities and nutrient release commonly lead to the onset of hypoxia. No correlations were found with changes in precipitation or temperature. There is no evidence for a post-1980s return to well-oxygenated lacustrine conditions in industrialized countries, despite the implementation of restoration programmes

California: San Francisco Bay (marine)		Potential stratigraphical connections?	Hawaii: Maha ulepu, Kauai (terrestrial and marine)	
Anthropocene 1950 ?	American era		United States 1959	Indigenous snails extirpated
			Territory of USA 1898	<i>E. rosea</i> (1957) <i>R. rattus</i> (after 1950) <i>L. fulica</i> (1936)
Holocene (<i>pars</i>)	United States 1846	<i>L. fulica</i> , late 1940s (San Pedro, LA) Terrestrial	Polynesians (latterly with European and Asian Influence from 1778)	<i>E. caballus</i>
	Spanish and Mexican era (beginning 1769) San Francisco 1776	<i>C. virginica</i> 1866 (Pearl Harbor, Oahu) Marine	James Cook 1778	European livestock
	Indigenous People (European visitors from 1542 onwards)	Pre-neobiota Interval in the Bay (but, for example, re-introduction of <i>E. caballus</i> by Spanish to land)	Polynesians	1425–1665 <i>S. scrofa</i> <i>R. exulans</i> 1039 to 1241 AD No clear evidence of human influence at Maha Ulepu
			Pre-human	Pre-neobiota interval

Fig. 11. Chronology of selected invasive mollusc species into San Francisco Bay (for dates of invasion see: Carlton et al., 1990; Cohen, 2004, 2011; Fofonoff et al., 2017, Committee on Nonnative Oysters in the Chesapeake Bay, National Research Council, 2004), and terrestrial invasive species in the Maha'ulepu sinkhole succession of Kauai, Hawaii (Burney et al., 2001). In both successions, cultural human changes are indicated by the left-hand column, and neobiota in the right-hand column. Although the two successions developed nearly 4000 km apart, and in warm temperate and tropical zones respectively, by the mid-19th century some taxa present on the California coast (e.g. the bivalve *Crassostrea virginica*, see DeFelice et al., 2001) were also present in Hawaii, suggesting the possibility of correlation between remote successions.

(Jenny et al., 2016). Hypoxia notably can change the behaviour of trace elements and radionuclides, by either increasing or reducing their mobility (e.g. Pavin Lake, France; Jeandel, 1981).

The radiogenic fallout signature is commonly well expressed in unbioturbated lake sediments, especially as the mid-latitude Northern Hemisphere peak fallout distributions (Hancock et al., 2014; Waters et al., 2015) coincide with the greatest abundance of hypoxic lakes. Shallow lakes are advantageous in that they show little of the settling delay evident in the anoxic marine basins (see Section 3.2). However, they may still be prone to reworking of fallout signals from catchment areas, such that the onset of the signal will probably be unaffected, but the bomb-peak may be broadened although still centred at ~1964 CE. Lake Victoria in Australia is an example of a lake with a large catchment. Despite showing the expected peak of the $^{239} + ^{240}\text{Pu}$ and ^{137}Cs signals in 1964 CE (Fig. 13), the Cs peak is broader and the post-1964 CE Pu decline is prolonged because of sedimentation associated with erosion within that catchment.

Abundant organic and clay components in lake sediments enable sorption of the main long-lived radiogenic elements (plutonium and radiocarbon). Where plutonium concentrations are relatively high, α -spectroscopy effectively measures the combined activity of ^{239}Pu + ^{240}Pu and also measures ^{238}Pu activity (Fig. 13), whereas mass spectrometric (MS) techniques are required where concentrations are

low and to measure distinct ^{239}Pu concentrations (Hancock et al., 2014). However, plutonium (and caesium) can be mobilized under anoxic conditions in meromictic lakes, e.g. Pavin Lake, France (Jeandel, 1981).

The atmospheric transfer of reactive nitrogen (Nr) results in the deposition of NH_4^+ , HNO_3 , and NO_3^- even in remote lakes, as demonstrated in North America and the Arctic regions (Holtgrieve et al., 2011; Wolfe et al., 2013). Stable nitrogen isotope values from lake sediments of those regions show a consistent perturbation of $\delta^{15}\text{N}$ commencing after 1850 CE, typically at $1895 \text{ CE} \pm 10 \text{ years}$ (Holtgrieve et al., 2011, Fig. 14), but with a faster rate of decline from 1950 to 1970 CE and after 1980 CE (Wolfe et al., 2013). This decline is attributed to fossil fuel combustion and the production and use of fertilisers, both with lower isotopic ratios compared with catchment and pre-industrial atmospheric sources (Holtgrieve et al., 2011).

Although spheroidal carbonaceous particles (SCPs) are first recorded from strata dating to about 1830 CE, they show a near-synchronous global mid-20th century increase in abundance. This increase was driven by the introduction of fuel-oil combustion, in addition to coal, and coincided with increased abundance of large-scale power plants, with peak abundance in the 1970s to 1990s (Rose, 2015). Black carbon, produced from the incomplete combustion of fossil fuels and vegetation, can accumulate in lakes from both atmospheric aerosols and

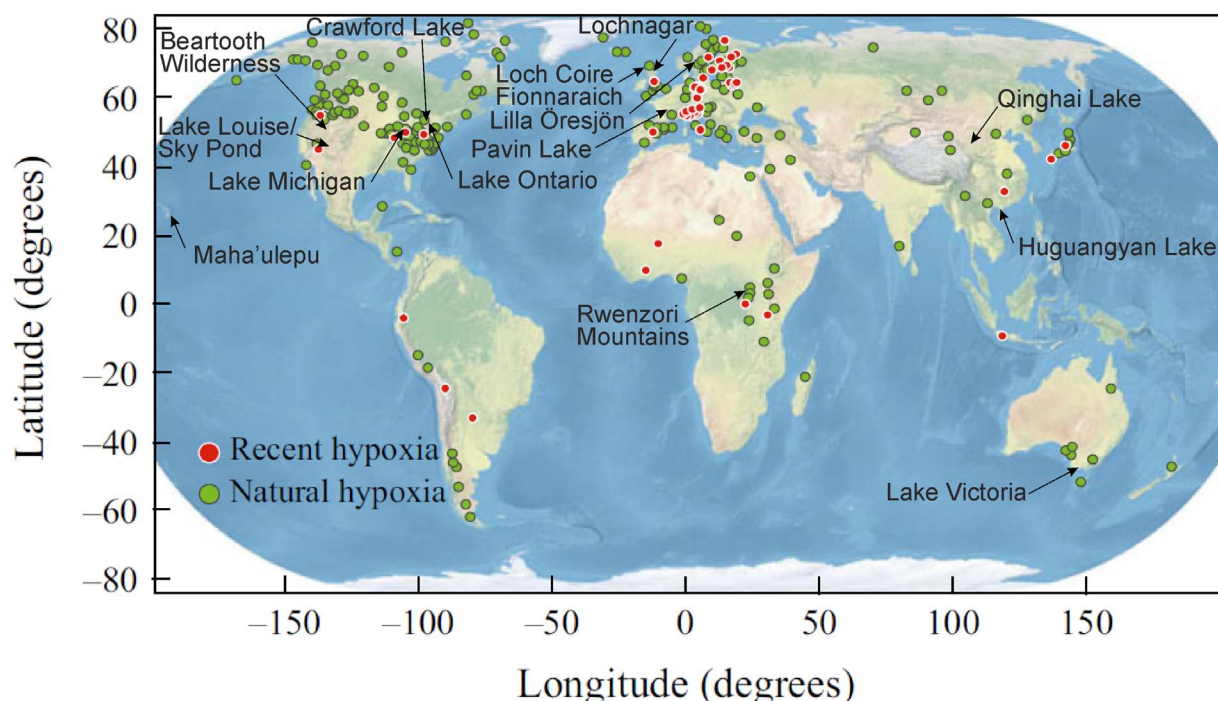


Fig. 12. Location of 365 sites recording lake hypoxia (Jenny et al., 2016). Recent hypoxia records onset of varves after 1700 CE, whereas naturally hypoxic lakes were taken to be those in which laminations persisted for at least 300 years.

riverine input, and is inert and resistant to degradation (Han et al., 2016). The atmospheric component, generally seen as a smaller size fraction (soot), is more widely transported and hence provides a more regional signal than coarser riverine char (Han et al., 2016). Biomass burning typically produces higher char/soot ratios than fossil fuel combustion, especially motor vehicle emissions (Han et al., 2016). Polycyclic aromatic compounds (PACs) tend to be co-produced with black carbon during hydrocarbon combustion. These can include polycyclic aromatic hydrocarbons (PAHs) and derivatives such as oxygenated PAHs (OPAHs) and azaarenes (nitrogen heterocyclic PAHs, AZAs) (Han et al., 2016). The relationship is demonstrated for Huguangyan Lake, eastern China (Section 3.5.4).

Metal concentrations in lake sediments can be affected by variable factors such as the timing of local industrialisation, distinct sources of sediment supplied by river fluxes into the lake, and varying degrees of weathering of the source catchment areas. Anthropogenic Pb deposition in lakes typically shows elevated concentrations in modern sediments in response not only to local industrial sources, but also to regional to global scale Pb atmospheric release. In remote lakes, the supply of anthropogenic Pb is sourced via the atmosphere (e.g. Lake Qinghai, China; Jin et al., 2010), providing a wider and more consistent signal than Pb supplied via river-water influx from nearby industrial sources. In Lake Qinghai, elevated anthropogenic Pb concentrations from the 1960s to at least 2000 CE are considered consistent with other remote

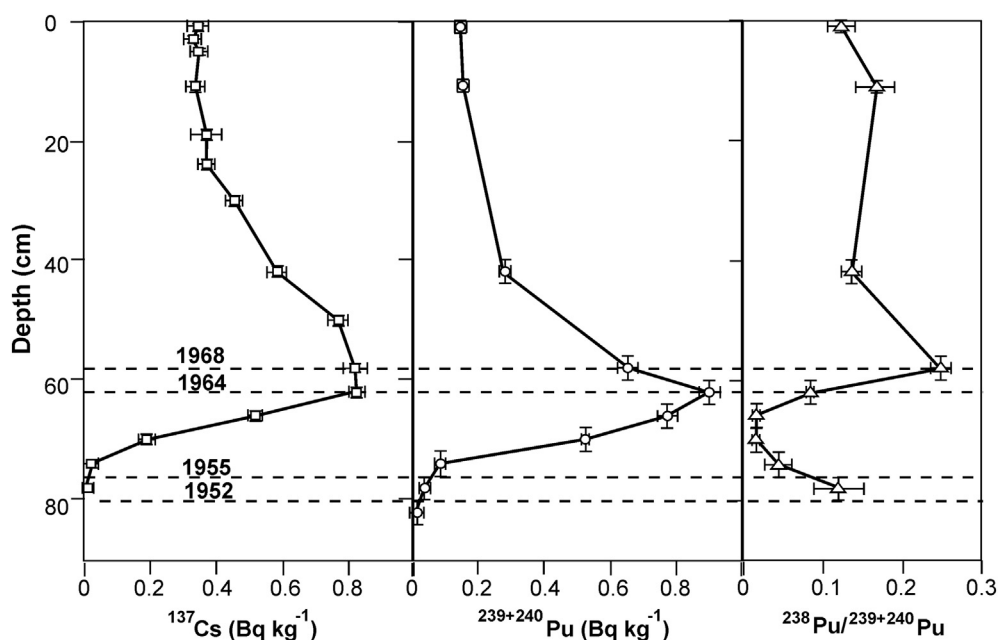


Fig. 13. Radiogenic signature from Lake Victoria, Australia (Hancock et al., 2011). Profiles of ^{137}Cs (closed squares), $^{239} + ^{240}\text{Pu}$ (open circles) and $^{238}\text{Pu}/^{239} + ^{240}\text{Pu}$ (triangles).

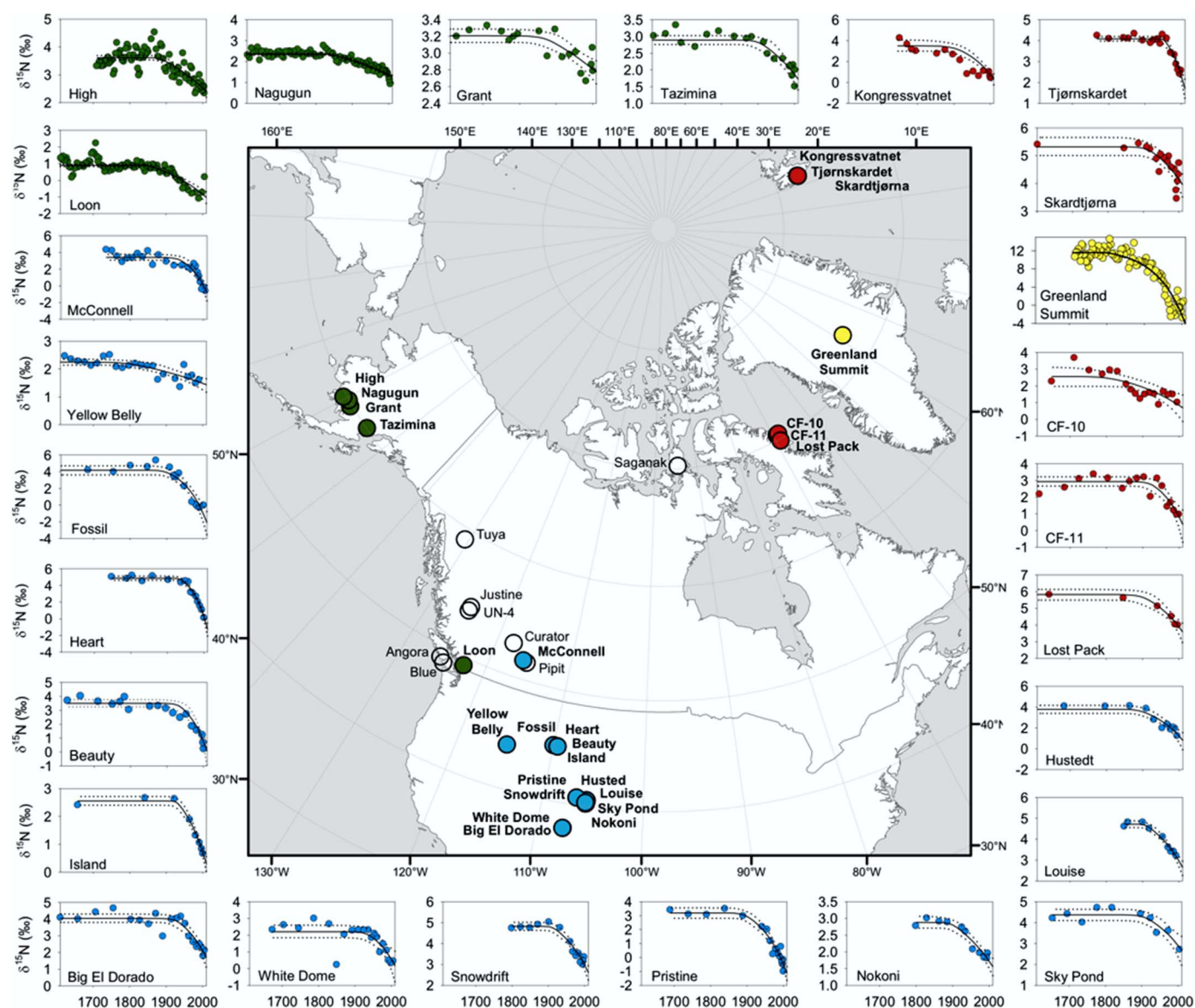


Fig. 14. Sediment $\delta^{15}\text{N}$ profiles from Northern Hemisphere lakes (from Holtgrieve et al., 2011). Lake ecotypes include: temperate/boreal (green circles), alpine (blue circles), and arctic (red circles), and the Greenland Summit ice core is indicated with a yellow circle. The solid lines are the median posterior fits to the observed data using the most parsimonious model, and the dotted lines are the 2.5 and 97.5% credible limits.

Northern Hemisphere lakes (Jin et al., 2010). The lakes of the Rwenzori Mountains of Uganda are distant from direct sources of Pb, and therefore the lake sediments potentially contain reliable archives of long-range atmospheric pollution (Yang et al., 2010). Hg concentrations in these high-altitude lakes show increased Hg burdens from the 1860s, with significantly elevated levels from 1930 ± 6 CE (Yang et al., 2010).

The Great Lakes of North America show widespread microplastic contamination, mainly in surface waters and beach sediments, while plastic pellets have been recorded in bottom sediments of Lake Ontario (Corcoran et al., 2015). Notably, high-density microplastics, including mineral-polyethylene and mineral-polypropylene mixtures, sink to the lake bottom rapidly and have been accumulating since at least 1977 CE (Corcoran et al., 2015). Microplastics tend to concentrate in nearshore sediments, within low-energy environments close to urban and industrial areas at concentrations of up to $\sim 28,000$ particles kg^{-1} dry sediment and sediment depths of up to 15 cm (Ballent et al., 2016). Where microplastics are deposited at water depths of < 40 m, they are prone to reworking by storm events (Ballent et al., 2016).

Persistent organic pollutants (POPs) are characterised by slow rates of environmental degradation (years to decades), low solubility, high absorption to suspended particles and efficient atmospheric transport. Chlorinated pesticides (e.g. DDT and its degradation products), show a consistent worldwide rise from the 1950s (Muir and Rose, 2007), and subsequent decline in the youngest sediments reflecting the US banning of DDT in 1972 CE. Temporal trends of input fluxes of polychlorinated biphenyls (PCBs), and the fire retardants hexabromocyclododecanes (HBCD) and polybrominated diphenyl ethers (PBDEs) in sediment cores from English lakes in urban settings show generally slow rises from ~ 1960 CE, which are less clearly developed in rural settings. In the case of HBCD, the rise may be delayed by some 35 years (Yang et al., 2016).

Invasive zebra mussels (*Dreissena polymorpha*) and quagga mussels (*Dreissena rostriformis bugensis*) represent novel species introduced into the North American Great Lakes in 1988 and 1989 CE respectively (<https://nas.er.usgs.gov/taxgroup/mollusks/zebramussel/>), both originating from southern Russia/Ukraine. As well as providing a novel biogenic signal, they also greatly modify the biotic content of the lakes.

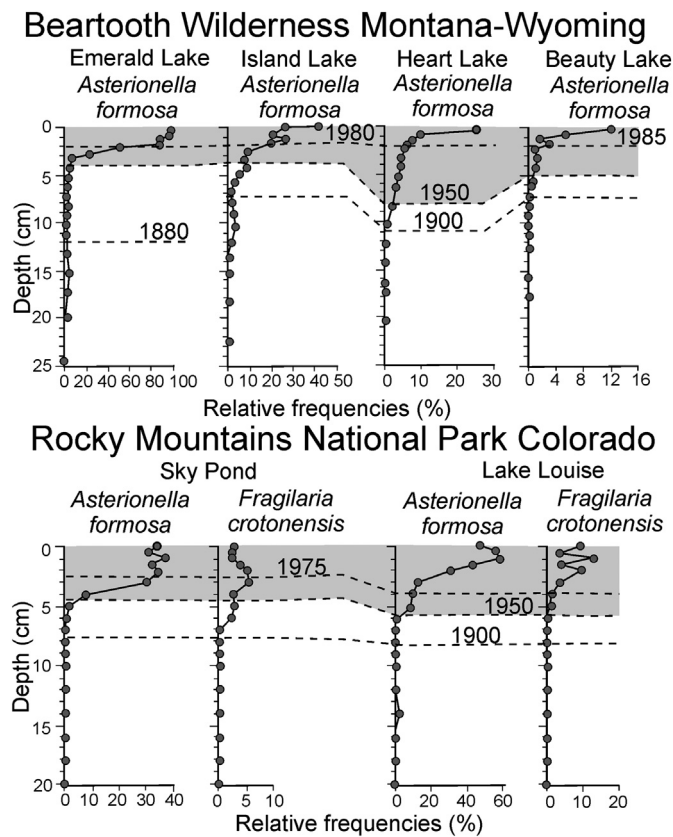


Fig. 15. Replacement of Holocene diatom assemblages by *Asterionella formosa* and/or *Fragilaria crotonensis* mainly in lake cores from US high-altitude sites, since ~1950 CE (Saros et al., 2005). Location of lakes shown on Fig. 14.

The trillions of quagga mussels in Lake Michigan (USA) can filter the equivalent of the entire lake volume in about 1 or 2 days. This has resulted in a shift in diatom composition, with significant reduction in the larger diatom genera, e.g. *Stephanodiscus* and *Aulacoseira*, leaving an impoverished algal community with mainly smaller genera, e.g. *Cyclotella* (Evans et al., 2011). Diatom assemblages in remote Northern Hemisphere lakes commonly show a consistent pattern. Typical Late Holocene benthic diatoms, such as *Staurosira* and *Achnanthisidium* show declines concurrent with the rise of planktonic diatoms, first associated with an initially time-transgressive appearance of *Discostella stelligera*, attributed to climate warming (Wolfe et al., 2013). After 1950 CE, the planktonic diatoms *Asterionella formosa* and *Fragilaria crotonensis* became dominant (Saros et al., 2005, Fig. 15; Wolfe et al., 2013 suggest

1980 CE). This later change may reflect elevated Nr availability (Wolfe et al., 2013, Fig. 14) as well as rising temperatures. A comparable biotic change in the remote oligotrophic Scottish lake of Loch Coire Fionnaraich began in the 19th century following several hundred years of ecological stability (Pla et al., 2009). Changes to diatom (mainly benthic) assemblages were strongly coincident with changes in spheroidal carbonaceous particle concentrations, and may result from atmospheric contamination, possibly Nr enrichment (Pla et al., 2009). Chrysophyte diatom cyst (mainly planktonic) assemblages changed in response to a combination of atmospheric pollution and regional climate warming (Pla et al., 2009).

Table 7 summarises the key advantages and disadvantages of having a potential GSSP within lacustrine environments. Varved lake sediments can be strongly influenced by local variations in the geology, hydrology and land use present within catchments, and may be associated with distinct geochemical signals related to input from local industries. Correlation of varved successions within individual lakes may be limited to the extent of the water body, but widespread airborne contaminants, e.g. radiogenic fallout, nitrates, fly ash (Table 2), can show remarkably consistent patterns in lakes across diverse latitudes and be associated across continents. Crawford Lake (Ontario, Canada), Lochnagar (Scotland), Lilla Öresjön (Sweden), Huguangyan Lake (Guangdong, China) and Maha'ulepu Sinkhole (Kauai, Hawaii) are described as examples of potential trans-continental reference sections.

3.5.1. Crawford Lake, Canada

Crawford Lake (Ontario, Canada) is a small (surface area 2.5 ha), deep (24 m), meromictic lake located close to industrial pollution sources, although it lacks any significant urban development within the lake catchment area. It has a sediment record spanning eight centuries, with a varve chronology dating to 1867 CE and ^{14}C dating for older laminae (Ekdahl et al., 2004). Maize (*Zea mays*) pollen and corn smut spores (*Ustilago maydis*) are irregularly distributed through an interval > 200 years, termed the Iroquoian, reflecting sporadic occupation by pre-European indigenous peoples (Ekdahl et al., 2004, 2007, Fig. 16). This occupation is associated with increased numbers of the planktonic diatoms *Asterionella formosa*, *Fragilaria crotonensis* and *Synedra nana*, depleted numbers of *Cyclotella michiganiana*, and the loss of *Cyclotella bodanica*. These changes are attributed to increased nutrient input, which raised lake productivity, causing bottom-water anoxia and altering diatom assemblages (Ekdahl et al., 2004, 2007). Additional stratigraphic proxies record marked changes in the late-19th century (starting 1867 CE) due to introduction of a new agricultural regime by European settlers, and again in the mid- and late-20th century, driven by regional land-use changes of comparable magnitude to the earlier changes by indigenous peoples.

Table 7

Reasons for and against using a lake deposit as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> • May be varved; relatively few omission surfaces • Spatially extensive with regional-global coherent signals • Independent dating using ^{210}Pb and ^{14}C • May be correlatable with tephra stratigraphy • Microplastics and metals with direct source from effluents • Spheroidal carbonaceous particles show global upturn • $\delta^{15}\text{N}$ marked decrease from mid-20th century • Persistent organic pollutants, e.g. chlorinated pesticides • Radionuclides, no settling delay in water column; clear bomb spike • Microfossils responsive to environmental change e.g. diatoms, ostracods, palynomorphs • Cores relatively easy to procure 	<ul style="list-style-type: none"> • Commonly low sediment accumulation rates (though typically greater than for marine anoxic basins) • Strong modulation of signals by local processes, except for upland or remote lakes • Pu and Cs are mobile under anoxic conditions at the bottom of meromictic lakes • No precedent as GSSP candidate; although 4 of 5 auxiliary stratotypes for the Holocene GSSP are lacustrine

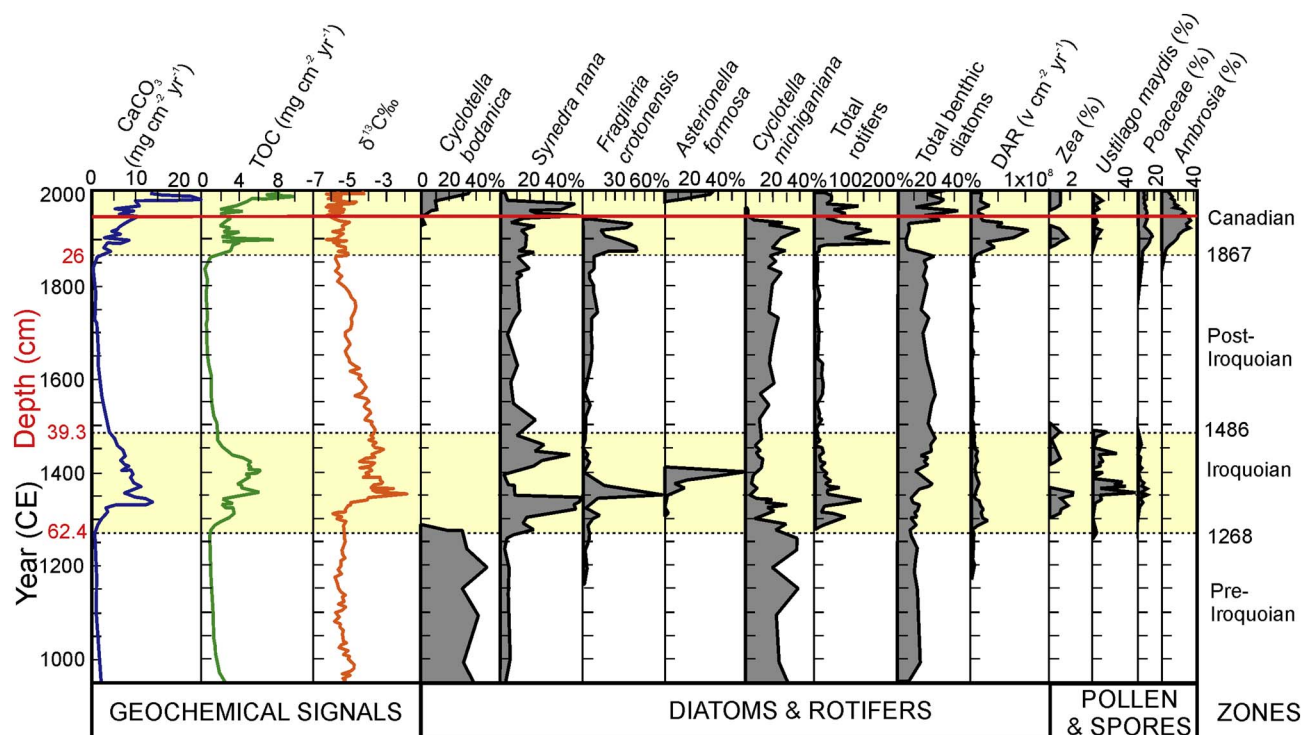


Fig. 16. Stratigraphic signals of the last millennium in Crawford Lake, Canada. The horizontal red line near the top is the ~1950 CE level, marked by both lithological and biostratigraphic changes in available data. From Zalasiewicz et al. (2017a), modified after Ekdahl et al. (2004). DAR is diatom accumulation rate.

3.5.2. Lochnagar, Scotland

Lochnagar is a small (surface area 9.8 ha), deep (26.4 m), high altitude (788 m a.s.l.) corrie lake in the Grampian Mountains of Scotland. As with many upland and mountain lakes, there are no direct sources of contamination in the catchment so all pollutants are originally from atmospheric deposition. However, the legacy of centuries of atmospheric inputs from industrial sources has resulted in a considerable store in the sparse peat soils which are now being remobilised to constitute a major pollution source to the loch (Yang et al., 2002a).

Lochnagar exhibits a full Holocene record contained within a sedimentary succession < 2 m thick. These deposits, neither laminated nor varved but dated by ^{14}C (Dalton et al., 2005), show catchment-driven changes due to the development and degradation of soils (e.g. post-Little Ice Age soil erosion) and human interference on terrestrial vegetation via 'fire management' over the last ~1000 years. Due to the steep morphology of the loch basin, accumulation of sediments varies across the loch with recent accumulations (Rose, 2007) being higher than earlier sedimentation rates, as is the case with many European lakes (Rose et al., 2011). Recent sediments are readily dated by ^{210}Pb , independently supported by 1963 CE nuclear weapons testing signals of ^{137}Cs and ^{241}Am and the Chernobyl ^{137}Cs peak in 1986 CE.

The diatom-inferred pH reconstruction for the last 9500 years shows a stable pH of 6.0–6.5 until the 19th century (Monteith et al., 2007) when pH started to decline with the resultant loss of acid-sensitive diatom taxa (*Achnanthes minutissima*, *Achnanthes scotica* and *Fragilaria virescens* var. *exigua*) and an increase in those which are more acid-tolerant (*Achnanthes marginulata* and *Aulacoseira distans* var. *nivalis*) (Battarbee et al., 2014). The pH eventually reached a low of 5.5 in the 1970s before starting to recover in the 1980s, although the trajectory for diatom recovery is not a straightforward reversal (Battarbee et al., 2014). Alongside the 19th century decline in pH were: a) the start of spheroidal carbonaceous particle deposition, which elevated markedly in the mid-20th century (Rose and Yang, 2007, Fig. 17a); b) increases in the concentrations of trace metals (Fig. 17b) such as Pb (from 1860s) and Hg (from the early-20th century), becoming elevated over long-term base-lines (Yang et al., 2002a), and also an increase in Cd, Cu and

Zn concentrations (Yang et al., 2002b); and c) a decline in $\delta^{15}\text{N}$ starting in the late-19th century (Curtis and Simpson, 2011, Fig. 17c). All of these signals indicate contamination via atmospheric deposition of long-range industrial pollutants.

Mercury, Pb and SCPs show a further and more significant increase in concentration in the mid-20th century (Fig. 17a and b), primarily from fossil fuel combustion. This period is also accompanied by significant increases in a range of POPs such as organochlorine pesticides, including DDT, although the accumulation in Lochnagar sediments lags about a decade behind emission figures (Muir and Rose, 2007, Fig. 17d). The organochlorine pesticide toxaphene was never used or produced in the UK (Muir and Rose, 2007; Rose et al., 2001), but global emission was greatest in the 1970s with high accumulation rates in Lochnagar sediments in the 1980s and 1990s (Fig. 17d). Industrial chemicals including chlorobenzenes and PCBs show elevated values from ~1960 CE and, later, brominated flame retardants such as PBDEs show elevated concentrations from ~1980 CE (Muir and Rose, 2007) (Fig. 17e–g). The lake sediment records of trace metals and SCPs are matched by records of these same contaminants in the catchment peats of Lochnagar (Yang et al., 2001) with both mid-19th and mid-20th century increases observable.

While the sediment and peat cores of Lochnagar faithfully reflect the temporal trends in contaminant deposition through most of the industrial period, there is evidence that this is not now the case. Temporally-resolved full-basin inventories of Hg and Pb (Yang et al., 2002a) show no decline in inputs to lake sediments in recent decades (Fig. 17b) even though emissions of these metals to the atmosphere have declined by 80–90% since the 1970s (NAEI, 2017). This 'additional' input of metals can only come from the store of legacy contaminants in catchment soils as a result of increased soil erosion, possibly in part a result of a changing climate. This hypothesis is supported by work at other upland lakes in Scotland for a suite of trace metals and also for SCPs (Rose et al., 2012). The implication of this is that even while upland lake sediments are no longer reliably recording changes in atmospheric deposition (Yang and Smyntek, 2014), by being a sink for reworked anthropogenic deposition on land, they are concentrating and

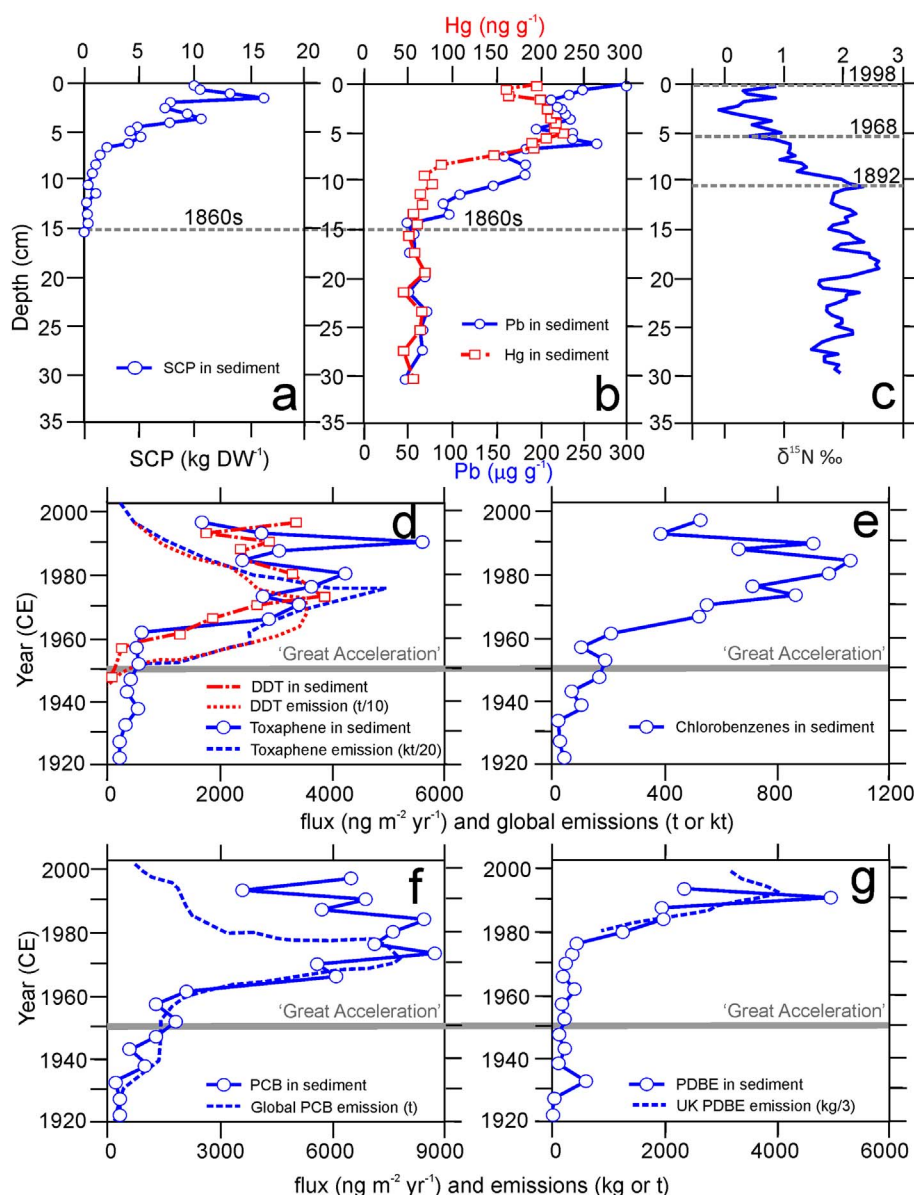


Fig. 17. Contamination record in Lochnagar sediments (Scotland) demonstrating appearances as post-mid-20th century markers: (a) spheroidal carbonaceous particles (SCPs) (from Yang et al., 2002a); (b) Hg and Pb (from Yang et al., 2002a); (c) $\delta^{15}\text{N}$ (from Curtis and Simpson, 2011); (d) the chlorinated pesticides DDT and toxaphene from core collected in 1997 CE related to emission data (from Muir and Rose, 2007); (e) chlorobenzenes (from Muir and Rose, 2007); (f) total PCBs related to global emissions (from Muir and Rose, 2007); and (g) PDBEs related to UK emissions (from Muir and Rose, 2007).

temporally extending an anthropogenic signal. This would not affect a mid-20th century GSSP from among such sites.

3.5.3. Lilla Öresjön, Sweden

A lake core from Lilla Öresjön (SW Sweden), a high-sulphate deposition area, records marked physical, chemical and biological trends since 1800 CE (Renberg and Battarbee, 1990). Areas of acute lake acidification tend to be associated with atmospheric pollutants including heavy metals, sulphur, fly-ash (spheroidal carbonaceous particles) and PAH, related to fossil fuel combustion. However, anthropogenic Pb signals have a much longer record in lake sediments in Sweden, initiating 3500 to 4000 years ago, with a small, but clear Roman Pb peak at about 1 CE, a major and unreversed increase at ~1000 CE, markedly elevated levels during the Industrial Revolution, but most prominently after World War II, with peak concentrations at ~1970 CE (Renberg et al., 2000). The record of increased Pb concentration is associated with concomitant declines in the Pb isotope ratio (Renberg et al., 2000).

At Lilla Öresjön atmospheric pollution is evident with an increase in Pb, Zn, and benzo(a)pyrene at about 1900 CE (dated by ^{210}Pb) and a more pronounced increase of these signals along with sulphur and SCPS during the 1960s (Renberg and Battarbee, 1990, Fig. 18). Heavy metal and sulphur signals peaked in the 1960s and 1970s, and SCPS and PAH in the 1970s. Analysis of diatoms, chrysophytes, cladocerans and pollen show a succession of changes reflecting decreasing lake pH, with a distinct post-mid 20th interval discernable (Fig. 18 herein; Renberg and Battarbee, 1990).

3.5.4. Huguangyan Maar Lake, Guangdong, China

Huguangyan Lake, eastern China, receives little sediment input from rivers or from soil erosion, hence its records of black carbon (BC) and polycyclic aromatic compounds (PACs) reflect atmospheric deposition (Han et al., 2016). The sediment record covers the past 350 years and has been dated reliably for the last 150 years using ^{210}Pb and ^{137}Cs activities (Han et al., 2016). Concentrations of BC, PAH, oxygenated PAHs (OPAH) and azaarenes (AZAs) increased sharply

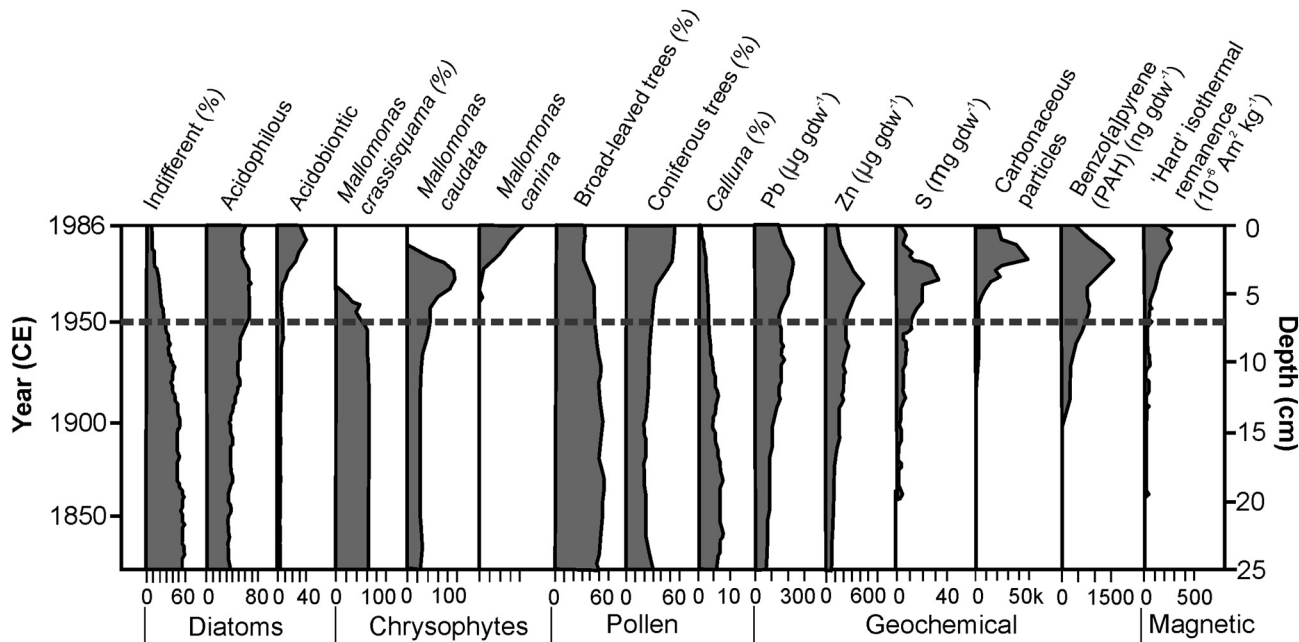


Fig. 18. Physical, chemical and biological trends from Lilla Öresjön (SW Sweden), a high-sulphate deposition area (from Renberg and Battarbee, 1990).

during the late-1940s to early-1950s, and again in the late-1970s (Han et al., 2016, Fig. 19). The BC record peaks from 2004 to 2006 CE, with subsequent rapid decrease, whereas the PAC record has continued to rise (Han et al., 2016).

The sedimentary record of char, soot, and PACs in this lake reflects the start of the rapid industrialisation in China after the 1950s, associated with a change in dominant energy sources from mainly wood burning in the pre-1950s to fossil fuels in the post-1950s (Han et al., 2016).

3.5.5. Maha'ulepu Lake of Kauai, Hawaii

The sedimentary successions of the Maha'ulepu Sinkhole and caves of Kauai, Hawaii (Burney et al., 2001) show a record of extirpation of geographically local taxa and biotic invasions of rats and other fauna. The sinkhole succession extends over the past 9500 years that accumulated in a freshwater to brackish lake, with periodic marine incursions, and which includes diatoms, land snails and bivalves. The diatoms, coupled with ¹⁴C dates, provide a detailed stratigraphy for the sinkhole and cave succession, which records step changes in anthropogenic influence through Polynesian and European colonisation of the island.

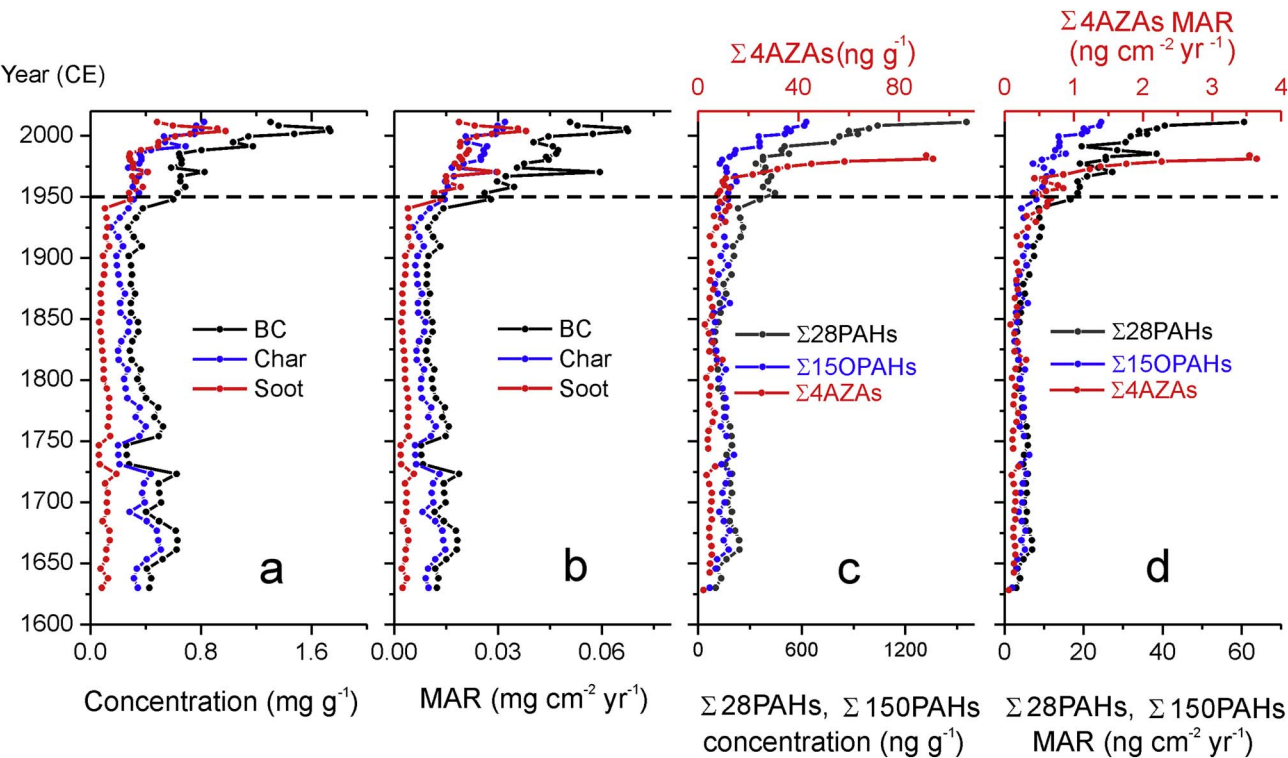


Fig. 19. Historical variations of concentrations and mass accumulation rates (MARs) of black carbon (BC), char, and soot, parent-PAHs, oxygenated PAHs (OPAHs), and azaarenes (AZAs) in the Huguangyan Maar Lake (from Han et al., 2016).

Human impact in the Maha'ulepu Sinkhole (Fig. 11) is first seen as an introduced Pacific rat dated to between 1039 and 1241 CE (Burney et al., 2001). Changes in land snail species indicate the impact firstly of Polynesian and latterly European colonisers and their associated neo-biota. Declines in several species coincide with Polynesian activity, while all indigenous snail taxa, with the exception of *Cookeconcha* cf. *psaunicostrata*, became extinct during the 19th and early-20th centuries. In the 20th century, *Lissachatina fulica* (in 1936 CE) – the Giant African Snail – and *Euglandina rosea* (in 1957 CE) – the Cannibal Snail of Central America – were introduced (Fig. 11). Anecdotal evidence suggests a 20th century extinction for the last indigenous snail population of *C. cf. psauicostrata*. These changes provide a detailed and unfolding biostratigraphical record of human impact on the local biota that is supported by diatom, bird, pollen, and other fossil data (Burney et al., 2001). Furthermore, in the marine succession, neobiotic marine bivalves such as *C. virginica* on Oahu offer potential for inter-regional correlation (Fig. 11).

3.6. Peat and peatlands (mires)

Peatlands (mires) may be defined as wetlands in which at least 40 cm of peat has accumulated; peat is partially-decomposed, sub-fossil plant material containing < 25% by weight mineral matter (Shotyk, 1992). Swamps and fens are “minerotrophic” wetlands, that is, plants growing there receive mineral nutrients primarily from surface waters and groundwaters. Fens are almost always peatlands, but swamps may be either mineral wetlands or peatlands, depending on the thickness of peat accumulation (Shotyk, 1992). In contrast, the surface layers in bogs are beyond the influence of these waters, and the plants growing there receive their nutrients solely from atmospheric inputs (termed “ombrotrophic”). Because bog plants are fed exclusively by rain and dust, the peats which accumulate in bogs have naturally low concentrations of mineral matter and trace metals, providing low background values against which anthropogenic inputs may be compared.

Note that wetlands are not necessarily peatlands, and that most peatlands are not ombrotrophic bogs.

Northern (boreal and subarctic) peatlands are by far the most extensive development of peat, covering some 4 million km², whereas the equivalent southern peatlands, mainly in Patagonia, South America cover only 45,000 km² (Yu et al., 2010, Fig. 20). Tropical peatlands cover an area of ~370,000 km² (Yu et al., 2010), the most extensive area being the ~145,000 km² of the Congo Basin (Dargie et al., 2017). In the case of the Congo Basin, however, peat was defined as having < 35% mineral matter, and a thickness of > 30 cm (Dargie et al., 2017), this differing from the conventional definition cited earlier. Tropical peats mainly occur along major river systems, and in SE Asia are associated with mangrove development along coastal zones (Yu et al., 2010). Peak accumulation rates for these peats are in the Middle Holocene for the tropical peats, Early Holocene for the northern peatlands, and Late Pleistocene for the southern peats (Yu et al., 2010).

Surface waters and groundwaters contain carbonate alkalinity from the chemical weathering of carbonate and silicate minerals, so the waters in fens and swamps tend to be neutral to alkaline in pH. Rainwater, on the other hand, contains very little alkalinity. In consequence, the organic acids generated during the decay of organic matter quickly render bog surface waters acidic (pH 4). Bog surface waters are also oligotrophic, so only specialized plant communities are able to thrive. Finally, the depth to water table in bogs will vary seasonally, resulting in redox conditions varying from oxic to anoxic. Seen from this perspective, the peat and corresponding porewaters in bogs represent a complex chemical milieu for the compounds that are continually being supplied by the atmosphere.

As peat accumulates over time, fens and swamps can evolve into bogs, as the surface layers become increasingly removed from the basal mineral sediments and plants gradually become deprived of nutrients. Thus, it is very common to find ombrotrophic bog peats overlying minerotrophic fen and swamp peats (Shotyk, 1988). To use peat cores from bogs as archives of atmospheric deposition, it is crucial to clearly

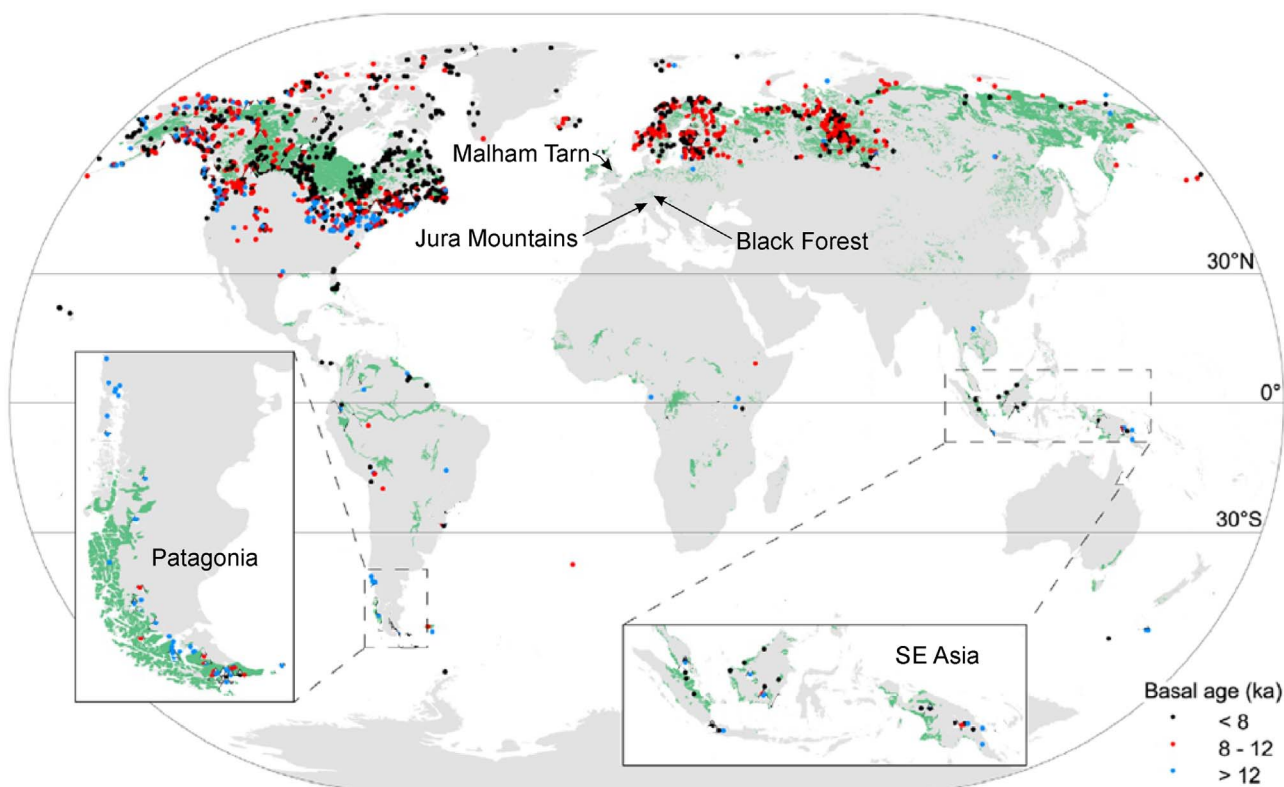


Fig. 20. Global map of peatland regions with basal peat ages (black < 8 ka, red 8–12 ka, and blue > 12 ka) (from Yu et al., 2010).

distinguish between the ombrotrophic versus minerotrophic sections: this can be done using a broad array of chemical indicators in either the solid or aqueous phases, such as ash or Ca content of the peat, or pH and Ca content of the porewaters (Shotyk, 1996).

Peat cores from bogs have been used to study atmospheric change over time using the broad range of parameters listed in Table 8. This list of proxies of environmental change does not imply their fidelity in the peat bog archive. Rather, it should be viewed as an illustration of the broad potential and number of opportunities represented by bogs to study environmental change. While most of these proxies have been impacted by human activities, few provide truly global signals, mainly because of the varying chronology and intensity of anthropogenic emissions throughout the industrial world (mineral dusts, spheroidal particles), biogeochemical transformations within the peatland (N and S compounds) and the size of the host aerosol which is crucial for long-range atmospheric transport (metals and organic contaminants). Radionuclides from nuclear weapons testing are exceptional in that they reached the stratosphere, are found in the sub-micrometre aerosol fraction, and therefore their fallout was dispersed globally (Junge, 1963).

The preservation over time of some of the constituents supplied by the atmosphere may be constrained by the low pH, abundance of complex-forming organic acids, range in redox potentials, and advective flow in response to water table drawdown. The extent of preservation ranges from near-perfect (plant macrofossils are extremely well preserved in anoxic water) to very poor (^{137}Cs is very mobile in ombrotrophic peat, simply because monovalent cations are poorly retained; it is far better retained in minerotrophic peat because of the abundance of phyllosilicates which fix Cs within their siloxane cavities). Some trace metals are very well preserved (e.g. Ti and Zr because of the stability of their host minerals in acidic waters), others not at all (e.g. Mn and Fe which form soluble, mobile cations under anoxic conditions). While there may be hundreds of publications employing bogs as environmental archives, there are remarkably few studies of chemical diagenesis of atmospheric contaminants, and the possible importance of post-depositional migration. Diagenesis of organic contaminants in peat is discussed at length by Thuens et al. (2013), using the example of PAHs, fallout radionuclides using the examples of U and

Pu by Quinto et al. (2013a, 2013b), and trace metals using the example of Pb by Shotyk et al. (2016).

Age-dating of peat profiles presents many challenges; peat bogs are not varved, but can have high (annual) growth rates, although these can be very variable. In Switzerland, 1 m peat cores ranged from ~ 1000 to 7000 years of peat accumulation (Shotyk et al., 2000). The most robust age-depth models use a combination of approaches, including ^{14}C (which is physically incorporated within the organic molecules making up the peat) and ^{210}Pb (which is supplied to the peatland surface by the sub-micrometre aerosol fraction and, apparently, becomes adsorbed to the peat).

Table 9 summarises the key advantages and disadvantages of hosting a potential GSSP within peatlands. Examples of geochemical records from peat bogs are provided from the Jura Mountains of Switzerland and Malham Tarn in England.

3.6.1. Pb, organic contaminants and radionuclide fallout deposition at Etang de la Gruère, Jura Mountains, Switzerland

The protected ombrotrophic *Sphagnum* bog of Etang de la Gruère has the longest record of peat accumulation (6.5 m accumulated in ~15,000 years) in the Northern Hemisphere (Shotyk et al., 1998). Traces of atmospheric lead appear in this Swiss peat bog from 12,370 ^{14}C yr BP (Shotyk et al., 1998). Climate-controlled enhanced Pb fluxes derived from soil dust are recorded with maxima at 10,590 and 8230 ^{14}C yr BP, the latter coinciding with the start of the Middle Holocene (Shotyk et al., 1998). The first anthropogenically-controlled increase in Pb deposition starting at 5320 ^{14}C yr BP is thought to be due to forest clearing and the introduction of agriculture (Shotyk et al., 1998). The first influence of Pb pollution from mining and smelting started at 3000 ^{14}C yr BP and occurs at a depth of 1.4–1.5 m, seen as increased Pb enrichment and decreased $^{206}\text{Pb}/^{207}\text{Pb}$ ratios (Shotyk et al., 1998). However, by far the greatest Pb flux occurred in the late 20th century (Fig. 21a), reaching 1570 times the background value by 1979 CE (Shotyk et al., 1998), associated with greatly decreased $^{206}\text{Pb}/^{207}\text{Pb}$ ratios (Fig. 21b). Subsequently, reduced Pb contents and increasing Pb isotope ratios in the later decades of the 20th century appear to record the introduction of unleaded gasoline and a reduction in industrial sources of Pb in the region (Shotyk et al., 1998). The Pb signals in peat

Table 8

Selected examples of the use of proxies for various types of atmospherically sourced environmental change.

Proxy	Type of environmental change	Example
Plant macrofossils	Climate change, human-induced changes in vegetation caused by impacts to watershed hydrology	Magnan et al. (2014)
Pollen	Landscape evolution in response to climate change, forest clearing and agriculture	Markgraf and Huber (2010)
Soil-derived mineral dusts	Wind erosion of soils due to climate change, or caused by forest clearing for agriculture and soil tillage	Chapman (1964); Vuorela (1983)
Volcanic ash	Natural, episodic inputs that may mask anthropogenic inputs of other inorganic constituents	Zoltai (1989)
Trace elements (Ag, As, Cd, Cr, Cu, Ni, Pb, Sb, Se, Sn, V, Tl, Zn)	Derived mainly from mining, smelting and refining of base metals as well as ferrous metallurgical processing and coal combustion. The chlor-alkali industry was a significant source of atmospheric Hg in the past, and artisanal gold mining is an important source of Hg today	As, Cu, Zn: Küttner et al. (2014); Cr, V; Allan et al. (2013); Hg: Enrico et al. (2016); Pb: Veron et al. (2014); Sb: Rothwell et al. (2010); Sn: Meharg et al. (2012)
Lithophile elements (Al, Sc, Th, Ti, Y and the REE)	Wind erosion of soils due to climate change, or caused by forest clearing for agriculture and soil tillage	Vanneste et al. (2015); Kylander et al. (2016); Pratte et al. (2017)
Platinum group elements (PGE)	From automobile catalytic converters	Rauch et al. (2004, 2010)
Fallout radionuclides (including ^{137}Cs , ^{241}Am ; also isotopes of U and Pu)	From atmospheric nuclear weapons testing, plus accidental releases from nuclear reactors	Quinto et al. (2013a, 2013b)
Organic contaminants	Polycyclic aromatic hydrocarbons (PAHs) from both natural (prairie and forest fires) and anthropogenic (fossil fuel combustion, petroleum refining) sources	Zhang et al. (2016)
Organometallic compounds	Tetraethyl-lead and tetramethyl lead, from leaded gasoline	Shotyk et al. (2002)
Nitrogen and sulphur (including stable isotopes)	Atmospheric emissions of nitrogen oxides from fossil fuel combustion; sulphur oxides from this plus sulphide mineral roasting	Wieder et al. (2016)
Spheroidal carbonaceous particles	From fossil fuel combustion	Rose (2015); Swindles et al. (2015)
Inorganic ash spheres	From fossil fuel combustion and nuclear explosions	Fiałkiewicz-Kozielec et al. (2016)

Table 9

Reasons for and against using a peat deposit as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> Widely distributed and studied in northern and southern latitudes; extensive deposits in the tropics are beginning to receive attention Relatively rapid rates of accumulation of organic matter provide reasonable temporal resolution of atmospheric change Readily amenable to ^{14}C age dating (including the post-1950 CE atmospheric bomb pulse) Dating of recent peat accumulation (past ~150 years) possible using ^{210}Pb Ombrotrophic bogs receive inputs directly and exclusively from the atmosphere with no settling delay Wide variety of indicators of industrial activities are easily measured: soil-derived dust particles, N and S compounds, fallout radionuclides, heavy metals, organic contaminants, spheroidal carbonaceous particles (SCP) and inorganic ash spheres (IAS) 	<ul style="list-style-type: none"> In many industrialized regions most bogs have been damaged, altered or destroyed due to drainage for agriculture, development, forestry, or exploited for fuel or horticultural peat Peat accumulation rates vary over time depending on climate, hydrology, fire and inputs of mineral matter from wind erosion of soils, or volcanic ash; they are not varved ^{210}Pb age dating requires independent confirmation using chronostratigraphic markers including: other fallout radionuclides (such as ^{241}Am), pollen, tephrochronology, or contaminants of known emission history (e.g. DDT) Pb is immobile in undisturbed peat profiles but can be mobilized by bog water acidification (from acid rain) or peatland drainage (which promotes leaching) Minerotrophic peatlands are influenced and commonly dominated by aquatic inputs (surface and groundwater). Atmospheric signals may be discernible in minerotrophic systems provided that aquatic inputs are insignificant (e.g. fallout radionuclides) With acidic bog waters and seasonal variations in water table depth (and therefore varying thickness of the oxic surface layer), each proxy may be subjected to chemical diagenesis; the possible importance of post-depositional migration (or chemical transformation) of each proxy must be evaluated

at this locality are consistent with Pb levels in the Greenland GRIP ice core over the past 3000 years (Shotyk et al., 1998) and can be compared with the Greenland ACT2 core (Fig. 26).

The highest concentrations of PCBs occurred at a depth of 10–15 cm (dating from 1976 to 1960 CE; Fig. 21c) while PAHs reached a maximum at a depth of 20–25 cm (1951 to 1930 CE; Fig. 21d), with post-depositional downward migration of these compounds unlikely (Berset et al., 2001). The maximum PAH values coincide with greatest use of coal in Switzerland, after which fuel oil dominated (Berset et al., 2001).

The same core has records of ^{241}Am activity limited to 12–15 cm depth (Appleby et al., 1997), marking the early 1950s to early 1960s bomb-spike. ^{137}Cs activity is at a maximum in the living part of the profile in the upper 3 cm (Appleby et al., 1997; Fig. 21e), inferred to be sourced by the Chernobyl disaster of 1986 CE. ^{137}Cs is also recorded below the level of the ^{241}Am signal, probably because it has been remobilized downwards.

3.6.2. Spheroidal carbonaceous particles and Pb in Malham Tarn Moss, England

Spheroidal carbonaceous particles (SCPs) first appear in the peat succession adjacent to an upland lake at Malham, England, at a level

deposited in the 1850s, but there is a marked increase in their abundance in the 1950s (represented by < 10 cm of peat), with a peak in the 1970s (Fig. 22, Swindles et al., 2015). Atmospheric lead pollution from local industrial activity and additives in petrol, and increased soil erosion (reflected in the Fe and loss-on-ignition data), show comparable upturns in the mid-20th century.

3.7. Ice

Continental ice sheets including those of Antarctica, covering about 13 million km², and Greenland, covering about 1.7 million km² (UNEP, 2007), are together equivalent to about 9.4% of the total land surface area (Fig. 23). Ice from glaciers occurring both marginal to the ice sheets and in high-altitude regions makes up a further ~0.5% of the Earth's land surface, a coverage of $726,800 \pm 34,000$ km² (Pfeffer et al., 2014) (Davies, B. Mapping the World's Glaciers <http://www.antarcticglaciers.org/glaciers-and-climate/mapping-worlds-glaciers/> last updated 22/06/2017). Sea ice fluctuates in extent seasonally and does not form a permanent ice record for consideration as a potential GSSP site. Although ice sheets and glaciers have significant extents, what little permanent ice remains in Africa and Australasia (Fig. 23) is

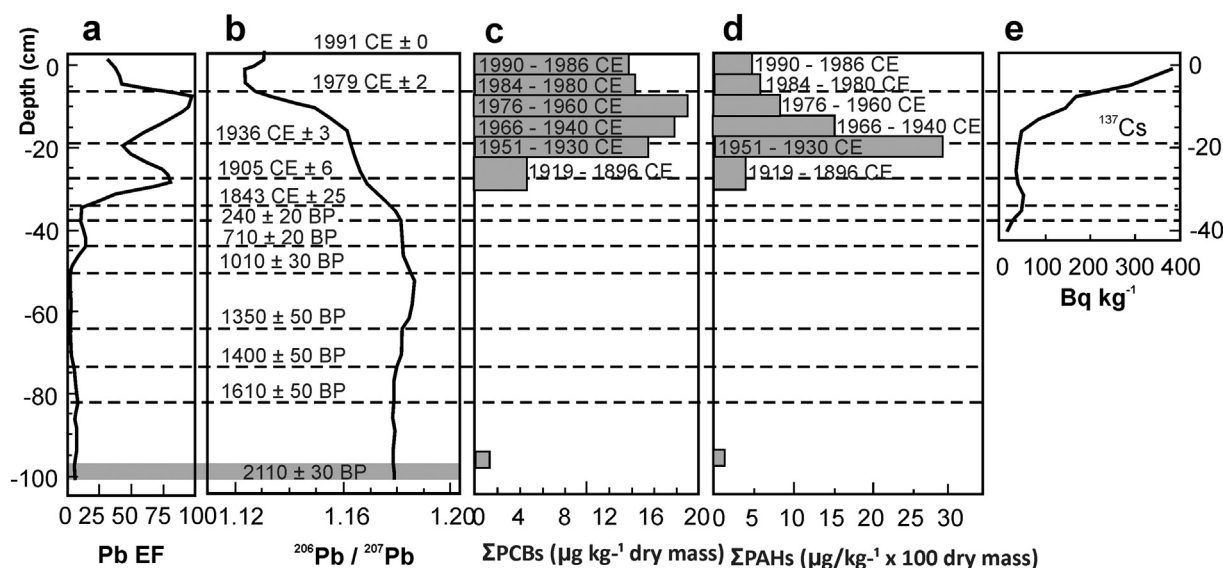


Fig. 21. Core from Etang de la Gruère, Switzerland, collected in 1991 CE: a) Pb Enrichment Factor (EF) calculated as the ratio of Pb/Sc in the peats, normalized to the background value (from Shotyk et al., 1998); b) Pb isotopic values (from Shotyk et al., 1998); c) ΣPCBs (from Berset et al., 2001); d) ΣPAHs (from Berset et al., 2001); e) ^{137}Cs (from Appleby et al., 1997). Succession dated using ^{210}Pb to 35 cm depth (Appleby et al., 1997) and ^{14}C yr BP from 35 cm to base (Shotyk et al., 1998).

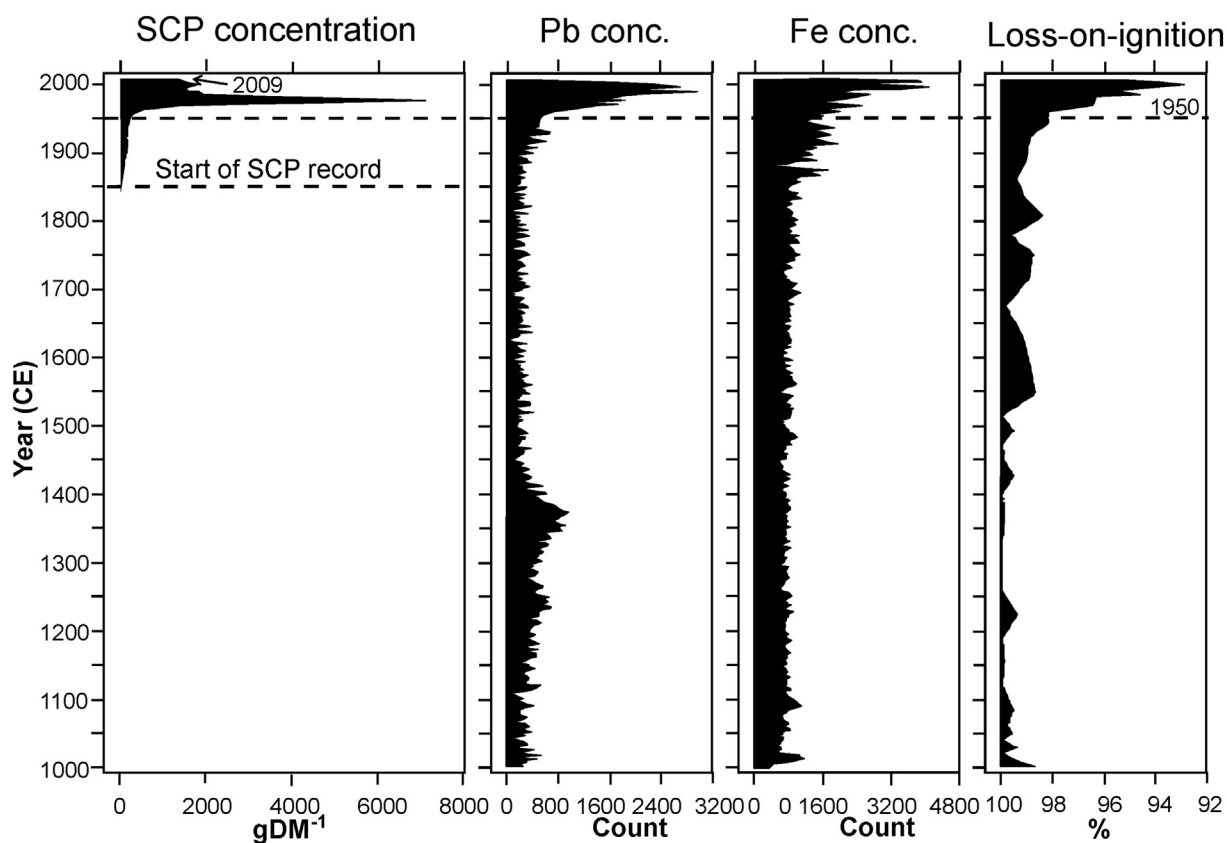


Fig. 22. Spheroidal carbonaceous particles (SCPs), lead and iron concentrations from Malham Tarn Moss, England (from Swindles et al., 2015).

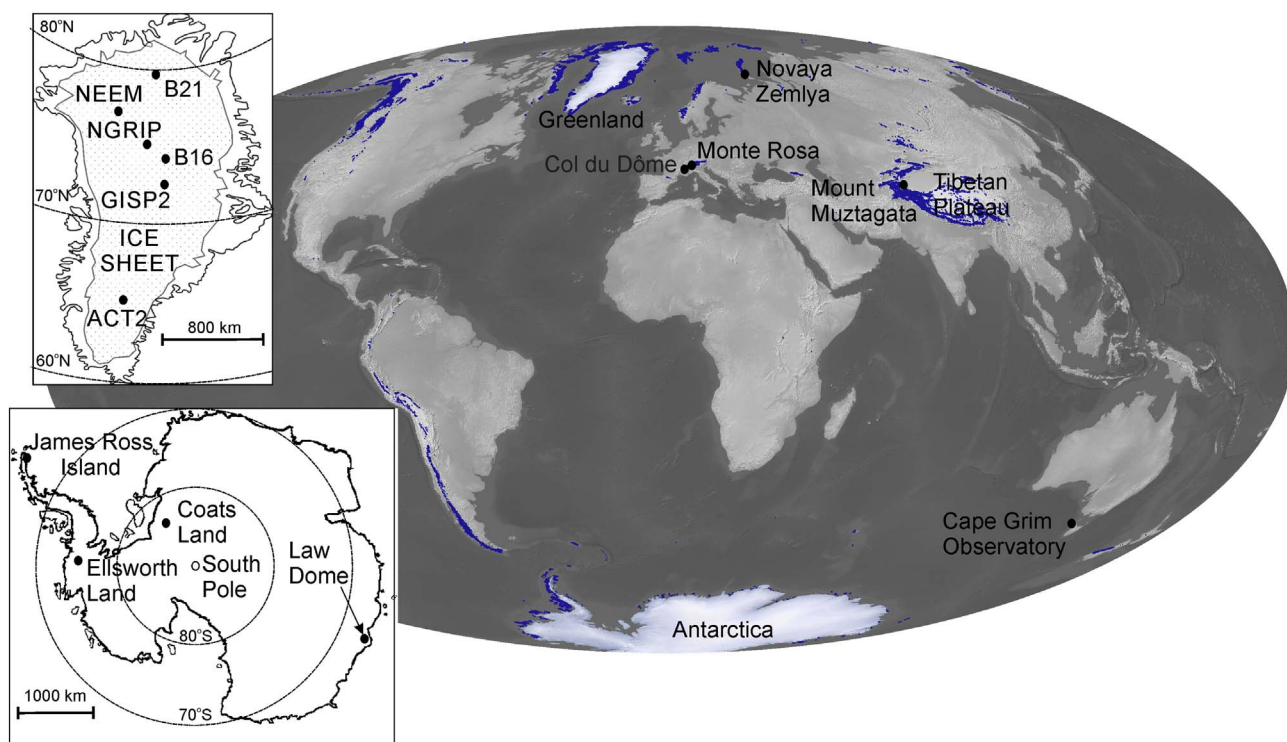


Fig. 23. Distribution of ice caps and glaciers (blue) and ice sheets (white) from NASA Earth Observatory Randall Glacier Inventory in 2014 CE <http://earthobservatory.nasa.gov/IOTD/view.php?id=83918> (acquired 7th May 2014; produced by Jesse Allen and Robert Simmon). Inset maps showing the main polar drilling sites in Antarctica and Greenland mentioned in the text.

rapidly being lost. Alpine glaciers from high altitudes may be suitable as potential GSSPs, but those from lower elevations are prone to significant seasonal melting and potential loss of laminae. Land glaciers in tropical regions are now melting from the top down and losing their younger layers (e.g. in Peru). Within alpine glaciers, unlike on polar ice sheets, meltwater processes can mobilize or remove solutes, attenuating or complicating the environmental signal (Schuster et al., 2002). In contrast, polar land ice is a more permanent record. The Greenland ice sheet shows the effects of greater local contamination from Northern Hemisphere industrial activities, while Antarctic land ice is more pristine.

Quaternary stratigraphy is closely linked to climatic variations during the ~2.6 million years of bipolar glaciation. Ice cores provide the best record of climate-related data for up to 800 ka, including isotopic ratios for $\delta^{18}\text{O}$ and δD , variations in CO_2 and CH_4 concentrations and dust, and S concentrations related to distinct volcanic events. The use of a Greenland ice core for the Holocene GSSP (Walker et al., 2009) provides a precedent that may be considered for the Anthropocene.

Atmospheric CO_2 and CH_4 concentrations and novel compounds such as sulphur hexafluoride SF_6 can be detected in ice-core air bubbles (Wolff, 2014). The ice record also captures anthropogenic signals derived from aerosols, including radionuclide signals from nuclear bomb testing (e.g. ^{239}Pu , ^{14}C), unprecedented increases in black carbon (BC) and metals such as Pb from industrial activity and automobile emissions, sulphate especially from coal combustion, and nitrate from fertilisers. These signals occur within annually resolvable laminae, the ages of which are typically determined by counting or from seasonal cycles in $\delta^{18}\text{O}$, and then verified using known volcanic eruption signals, which can be identified from electrical conductivity measurements in Greenland (e.g. Masson-Delmotte et al., 2015), and at Law Dome, Antarctica (Etheridge et al., 1996). Silicate dust concentrations in polar snow are exceedingly small, a mean of 2 ppb in Antarctica and 35 ppb in Greenland, with the latter showing no systematic trends over two centuries from 1750 CE (Murozumi et al., 1969).

The upper part of an ice sheet or glacier, which includes the succession that covers the mid-20th century interval, comprises firn

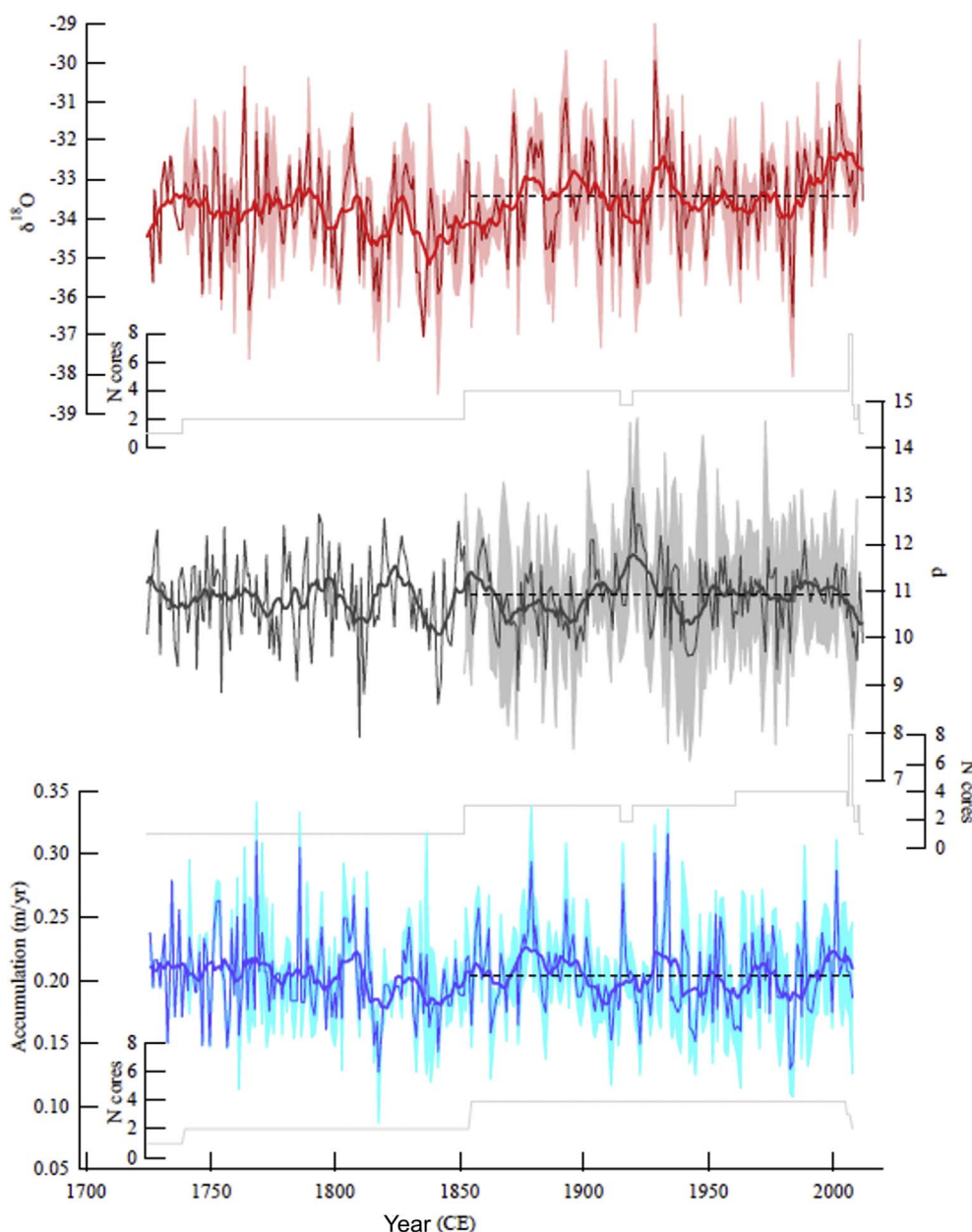


Fig. 24. $\delta^{18}\text{O}$, δD and accumulation rates for the North Greenland Eemian Ice Drilling site (NEEM) (Masson-Delmotte et al., 2015).

(partially compacted snow that is not yet fully ice), where bubbles are not yet closed from connection to the atmosphere. For most signals that are trapped directly in the solid ice this does not represent a problem, but CO_2 , CH_4 and N_2O are contained in trapped air bubbles. These air bubbles and their environmental signals are therefore always younger than the enclosing ice, and this effect is significant over the short timescales considered for the Anthropocene. However, recent work shows that the nitrogen isotopes in air bubbles can be used to estimate their ages correctly, enabling direct comparison with ice of the same age (Parrenin et al., 2013). The depth and hence timing of the firn-to-ice transition is dependent upon the accumulation rate, and so sites with rapid accumulation rates are preferable. At the South Pole, where the accumulation rate is only 8 cm yr^{-1} the firn-to-ice transition depth is at about 123 m (Rubino et al., 2013).

Over the past 350 years in East Antarctica, δD proxy values show a long-term warming trend of about $1 \pm 0.2^\circ\text{C}$, with a marked colder period from 1750 to 1860 CE corresponding to the end of the “Little Ice Age (LIA)” (Ekaykin et al., 2017). This is consistent with James Ross Island and other parts of the Antarctic Peninsula, these areas warming significantly from about 1950 CE onwards (Quayle et al., 2002; Abram et al., 2013). In north Greenland, $\delta^{18}\text{O}$ shows multi-decadal increasing trends in the late-19th century and since the 1980s, whereas the early-19th century had the lowest signals, associated with the Little Ice Age. Overall, Greenland δD and (laterally variable) $\delta^{18}\text{O}$ show no significant variation across the mid-20th century, with the most prominent climatically linked changes occurring about 1850 CE (Masson-Delmotte et al., 2015; Ekaykin et al., 2017, Fig. 24 herein).

One of the main climate-related signals evident in ice sheets is the increase in CO_2 concentration (Fig. 25), now 30% above the highest level of the last 800 ka (Wolff, 2014), and the $\delta^{13}\text{C}$ change to more negative values, both occurring at about 1960 CE in Law Dome ice core (Rubino et al., 2013) and discussed below. This may reflect the divergence of fossil fuel use (which has accelerated) and land use as independent sources of atmospheric CO_2 and hence of the $\delta^{13}\text{C}$ signal (Rubino et al., 2013). CH_4 is now at $\sim 1800 \text{ ppb}$ (Fig. 25), double the highest level of the last 800 ka (Wolff, 2014).

The most abrupt anthropogenic signal in ice is radioactive fallout from the atmospheric testing of nuclear devices. This is clearly evident in ice core as a marked increase in beta radioactivity initially in 1954 CE and subsequently in 1964 CE to form a bomb-spike peak in 1966 CE (Fig. 25), followed by slow decay towards natural background levels (Wolff et al., 1999; Wolff, 2014) after testing went underground following the Partial Test Ban Treaty in 1963 CE and essentially ceased following the Comprehensive Test Ban Treaty in 1996 CE (see Waters et al., 2015). The beta radiation is from several radioactive isotopes, some with short half-lives, e.g. ^{90}Sr and tritium. The long-lived radionuclide ^{239}Pu in polar ice has been transferred primarily via the stratosphere, with atmospheric residence times of up to 5 years and with seasonal transfer to the troposphere occurring in late winter and spring in the Northern Hemisphere (Arienzo et al., 2016). The accumulation rate of ^{239}Pu is typically greater in ice, and shows less post-depositional alteration or mixing, than in corals and lake sediments (Arienzo et al., 2016). Analysis for ^{239}Pu using traditional accelerator mass spectrometry (AMS) requires large sample size, typically reducing the resolution to about 3 years (Arienzo et al., 2016). A high-resolution (0.5–1.5 years) ^{239}Pu profile from Monte Rosa on the Italian-Swiss border, using ICP-SFMS, shows an initial peak from 1954 to 1955 to 1958 CE (Gabrieli et al., 2011). Following a temporary halt of testing in 1959–1960 CE, in which the Pu concentration decreased by half with respect to the 1958 CE peak, in 1961–1962 CE Pu concentrations increased rapidly to a peak in 1963 CE (Gabrieli et al., 2011). Following the Partial Test Ban Treaty, Pu deposition decreased sharply to a minimum in 1967 CE (Gabrieli et al., 2011). The interval 1967–1975 CE is characterised by small irregular Pu smaller peaks likely due to the deposition of Saharan dust contaminated by French nuclear tests of the 1960s, this effect diminishing after 1975 CE

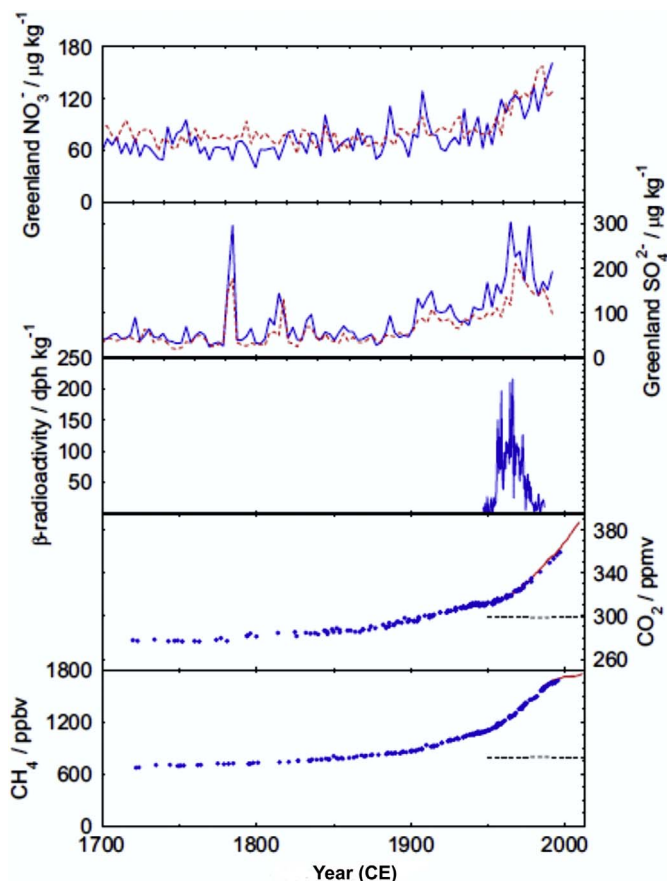


Fig. 25. Ice core signals since 1700 CE (from Wolff, 2014). CH_4 and CO_2 ice-core data (blue dots) are from Law Dome, Antarctica (MacFarling Meure et al., 2006), and recent atmospheric data (red lines) from Mauna Loa (CO_2) and Cape Grim (CH_4) observatories. The horizontal dashed lines are the highest values observed in ice cores of the last 800,000 years prior to the period shown. Beta radioactivity is from Coats Land, Antarctica (Wolff et al., 1999). Sulphate and nitrate are shown for two cores from Greenland: B16 (dashed red) and B21 (solid blue) (Fischer et al., 1998). The 18th- and 19th-century spikes in sulphate are signals of volcanic eruptions.

(Gabrieli et al., 2011). Arctic ice typically has a higher ^{239}Pu activity (differing by about a factor of three) than Antarctic ice (Arienzo et al., 2016) and, despite the first significant atmospheric thermonuclear detonations occurring in 1952 CE, the first detection of ^{239}Pu in polar ice was in 1953 CE (Arienzo et al., 2016). The post-moratorium ^{239}Pu bomb-spike (peak signal) occurs in the 1962 CE ice laminae in the Arctic, but is not strongly resolved in the Antarctic, mainly because the principal testing occurred in Novaya Zemlya (northern Russia), and little of this fallout crossed into the Southern Hemisphere (Arienzo et al., 2016).

Ice cores provide the most straightforward archives of sulphur, both as a record of global volcanic events (Schuster et al., 2002), but also in documenting the long-range effects of pollution unaffected by the local pollution seen in tree rings and speleothems (Fairchild et al., 2009). In Greenland ice, sulphate (SO_4^{2-}) concentrations increased markedly from 1900 to 1920 CE and 1940 to 1980 CE (Figs. 25 and 26), and by 1980 CE concentrations were a factor of 2–5 above pre-industrial peak values (Fischer et al., 1998), but did not exceed concentrations linked to large volcanic eruptions or during the Last Glacial Maximum (Wolff, 2014). In central Asia, in ice on Mount Muztagata (China), the steep rise in sulphate concentrations that started after the mid-1970s and decreased in the 1990s is associated with changes in industrial activity (Zhao et al., 2011). No equivalent signature is recorded in Antarctic ice, which is more remote from the main sources of sulphate pollution (Wolff, 2014).

N_2O , a greenhouse gas, has an atmospheric lifetime of about 120 years, so any changes recorded in ice cores reflect global changes, thought to mainly reflect increased application of agricultural fertilisers (Wolff, 2013). N_2O concentrations recorded in air bubbles within the ice have risen by about 20% in the last 200 years (see Law Dome example below) and are at levels higher than recorded for the last 800 ka (Wolff, 2013). In contrast, ammonium (NH_4^+), mainly deposited as aerosol, and nitrate (NO_3^-), deposited as an aerosol or as nitric acid, have much shorter atmospheric residence times, and changes in their concentrations reflect local changes in source areas (Wolff, 2013). Mid-latitude glaciers, such as Mount Muztagata in China, show an approximately three-fold increase in ammonium concentrations between 1960 and 1990 CE, probably as a consequence of agricultural emissions (Zhao et al., 2011). However, there is no significant recent trend at either pole (Wolff, 2013). Whereas ammonium is permanently fixed in the snow, nitrate can be re-emitted back to the atmosphere (Fischer et al., 1998; Wolff, 2013). Increased nitrate concentrations evident in the Northern Hemisphere (Fig. 25) are mainly due to NO_x emissions from fossil fuel combustion (Wolff, 2013, 2014), including automobiles and coal-fired power stations. In Greenland ice the main phase of nitrate increase, which has risen by a factor of 2–3, occurred from 1950 to 1980 CE (Fischer et al., 1998; Wolff, 2013); by 1980 CE levels were higher than over the past 100 kyr (Wolff, 2014). In the French Alps, at the Col du Dôme, high snow accumulation rates allow distinction between summer and winter snow, with higher summer nitrate values showing a fivefold increase during the 20th century, most markedly between 1960 and 1980 CE and sourced by emissions within 1000 km of the glacier (Preunkert et al., 2003). However, nitrate does not provide a consistent global signal. In central Asia, at Mount Muztagata, the steep rise in nitrate concentrations, which increased by a factor of two, started after the late-1970s and had already peaked by 2000 CE (Zhao et al., 2011), whereas no significant increase occurs then in Antarctic ice (Wolff, 2013).

A shallow (100 m) GISP2 ice core from Summit (Greenland), ranging in age from 1718 to 2006 CE, shows a clear decreasing $\delta^{15}\text{N}$ trend in atmospheric nitrate, from pre-industrial values around 11 to -1%

in the early 21st century (Hastings et al., 2009). The $\delta^{15}\text{N}$ trend began about 1850 CE (Fig. 14), whereas the increase in nitrate from the same core rises in about 1890 CE (Hastings et al., 2009). The greatest rate of change in $\delta^{15}\text{N}$ occurred from 1950 to 1980 CE, coincident with a rapid rise in fossil fuel emissions (Hastings et al., 2009).

During pre-industrial times, about half the lead in the troposphere was associated with soil dust and half with volcanic gases (Patterson and Settle, 1987). Pb, As, Sb, Bi, Cu, and Zn enrichments in Arctic ice cores started at around 3300–3000 yr BP, with an additional significant Roman phase (Hong et al., 1994; Krachler et al., 2008, 2009). Further increases of Pb, Tl, and Cd in response to the Industrial Revolution are more significant, with elevated concentrations from the 1860s, accelerated concentrations from 1887 to 1889 CE and peaks in the early 20th century (Tl in 1911–1915 CE, Cd in 1906 CE, Pb in 1915 CE) (McConnell and Edwards, 2008, Fig. 26). This early peak is considered to reflect increased coal burning in North America and Europe (McConnell and Edwards, 2008). However, the greatest rises in Pb concentrations are from the 1950s, peaking in the 1960s, associated with emissions of Pb used in alkyl-leaded gasoline (Murozumi et al., 1969; Boutron et al., 1991; McConnell and Edwards, 2008). Tl shows no equivalent mid-20th century peak, whereas Cd has a 1946 CE peak concentration (Fig. 26). At its peak, Pb levels in Greenland ice were above Holocene background levels by a factor of 200 and exceeded concentrations seen during the last glacial maximum (Murozumi et al., 1969); in Antarctic ice they were a factor of five above pre-Industrial levels (Patterson and Settle, 1987; Wolff and Suttie, 1994). Use of lead additives in gasoline became limited in the USA and other Northern Hemisphere countries from about 1970 CE, and concentrations decreased in Greenland by a factor of 7.5 between 1970 and 1990 CE, with Cd (Fig. 26) and Zn sourced from industrial processes showing a decrease by a factor of 2.5 over the same interval (Boutron et al., 1991, McConnell and Edwards, 2008). In Antarctica, by 1920 CE anthropogenic Pb (mainly from metal processing) dominated concentrations that were 2 to 4 times higher than pre-Industrial levels (Wolff and Suttie, 1994). Lower concentrations in the 1930s to mid-1940s are followed by a marked upturn in concentrations by a factor of > 2 from

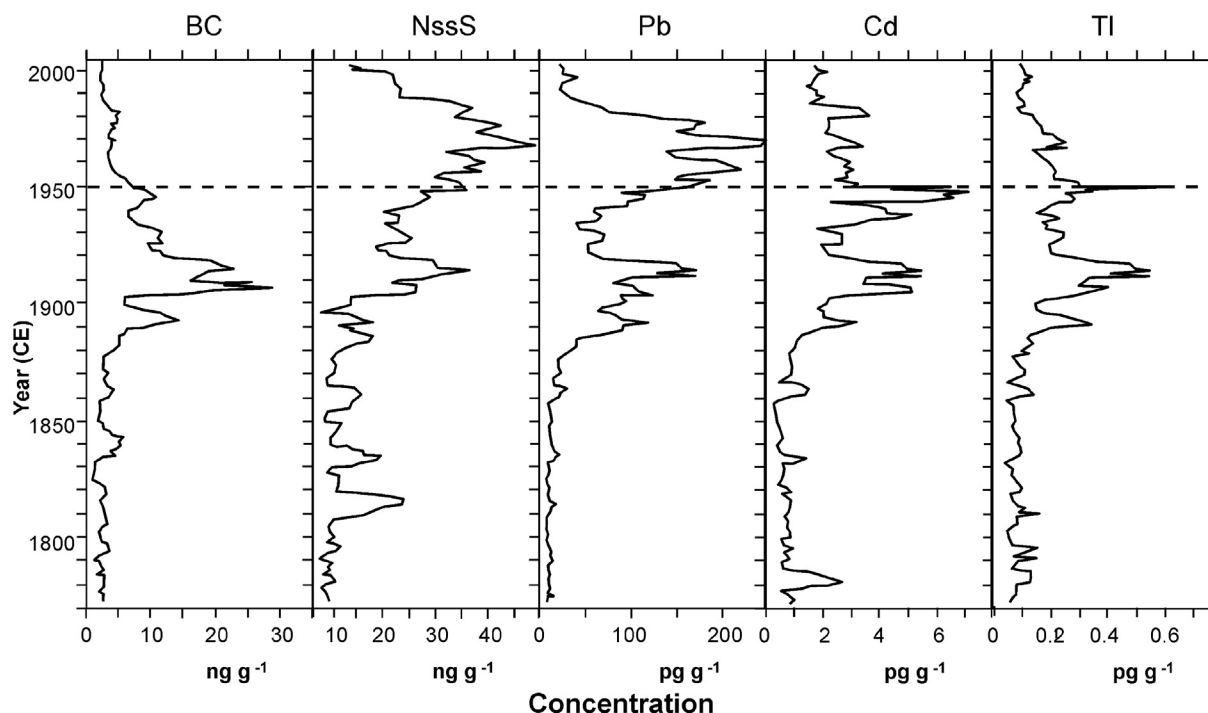


Fig. 26. Concentrations (5-year running means) in Greenland ACT2 ice core from 1772 to 2003 CE of sulphur (NssS), thallium (Tl), cadmium (Cd) and lead (Pb), compared to black carbon (BC), from McConnell and Edwards (2008). Sulphur peaks in 1817 and 1883 CE relate to the Tambora and Krakatoa volcanic eruptions, respectively. © (2008) National Academy of Sciences, U.S.A.

1950 to 1980 CE, with subsequent decline (Wolff and Suttie, 1994). The first use of unleaded gasoline in many Southern Hemisphere countries started as late as the 1980s and 1990s (Wolff and Suttie, 1994). Because most combustion of gasoline occurred in the Northern Hemisphere, it is in Greenland that the rise and subsequent fall of Pb signals is greatest. Lead aerosols have a 10-day residence time in the atmosphere resulting in little inter-hemispheric interchange (Patterson and Settle, 1987).

Atmospheric Hg contamination in a glacier in Wyoming (USA) from ~1720 to 1993 CE shows a twenty-fold increase from pre-1840 CE levels to a ~1984 CE peak (Schuster et al., 2002). Early spikes include both volcanic sources from Tambora (1815 CE) and Krakatoa (1883 CE), showing cross-hemispheric transfer of Hg, and an anthropogenic source with a double peak and elevated values from ~1850 to 1884 CE associated with the California Gold Rush in western USA (Schuster et al., 2002).

In Greenland, black carbon is attributed by McConnell and Edwards (2008) as: Pre-industrial (1772–1860 CE), coal-dominated industrial (1860–1940 CE), and oil-dominated industrial (1940–2003 CE), with peak concentrations in the first decade of the 20th century (Fig. 26). However, peak combustion of coal is later in Asia. Black carbon and organic carbon concentrations in Tibetan ice cores show a marked high during the 1950s and 1960s, attributed to the inflows of industrial emissions, mainly from Europe, with a decline during the 1970s and 1980s (Xu et al., 2009). Glaciers in the southern part of the plateau show black and organic carbon increases after the 1990s, from a southern source (Xu et al., 2009). Increased black carbon on glacial ice reduces surface reflectivity and increases melting (Xu et al., 2009), potentially causing significant hiatuses in modern laminae.

Microplastics have been reported in Arctic sea-ice (Obbard et al., 2014) from contaminated surface sea-water that has been frozen. However, there is no record of microplastics in polar continental ice sheets or glaciers. The potential for airborne transfer of microfibrils (Dris et al., 2016) suggests that microplastic fallout in glacial ice might occur, especially where glaciers are close to urban areas.

Table 10 summarises the key advantages and disadvantages of having a potential GSSP within glacial ice, as is the case for the base of the Holocene. The Law Dome ice core from Antarctica has a high-resolution multi-proxy record for the last 1000–2000 years and shows many characteristics that make it suitable for hosting a potential GSSP. Other similarly suitable ice cores are available elsewhere in both Antarctica and Greenland (Fig. 23), although many of the sites are unsuitable because the rate of accumulation is too low (e.g. Dome C, Antarctica).

3.7.1. Law Dome ice core, East Antarctica

A high-precision 1000–2000 year history of CO₂, CH₄, N₂O and $\delta^{13}\text{C}$ has been acquired from the Law Dome site in East Antarctica (Etheridge et al., 1996; Francey et al., 1999; Ferretti et al., 2005; MacFarling Meure et al., 2006; Rubino et al., 2013). As regards hosting a potential GSSP site, the high snow accumulation rates, low temperatures, and

small quantities of impurities at Law Dome make this record particularly suitable, especially for recognising a high-resolution $\delta^{13}\text{C}$ signal (Rubino et al., 2013). The Law Dome site comprises three cores drilled between 1987 and 1993 CE, (Etheridge et al., 1996; Rubino et al., 2013), dated by counting annual layers in oxygen isotope ratios ($\delta^{18}\text{O}$), ice electroconductivity measurements and hydrogen peroxide (H₂O₂) concentrations (Etheridge et al., 1996), with a gas-age/ice-age difference of between ~31 and ~55 years for CO₂ (Rubino et al., 2013).

Between 1000 and 1600 CE, the CO₂ concentration varies between 278 and 284 ppm (Fig. 27a), and between 1000 and 1500 CE, the $\delta^{13}\text{C}$ values remained constant at about –6.55‰ (Rubino et al., 2013). The brief decrease in concentration of CO₂ in the Law Dome ice core around 1600 CE (Etheridge et al., 1996; MacFarling Meure et al., 2006) was used by Lewis and Maslin (2015) as their “Orbis Event” which they regarded as sufficiently important to be a candidate for defining the beginning of the Anthropocene. Analysis of changes in atmospheric carbonyl sulphide concentration, which is linked to changes in gross primary production of terrestrial ecosystems, shows that temperature change, rather than vegetation re-growth, was the main cause of the increased terrestrial storage and hence drop in atmospheric CO₂ around 1600 CE (Rubino et al., 2016). After a recovery, CO₂ maintained comparatively low concentrations and elevated $\delta^{13}\text{C}$ through to 1750 CE (Rubino et al., 2013). Over the past 200 years, atmospheric CO₂ concentrations at Law Dome have increased by 29% (MacFarling Meure et al., 2006). Following the Industrial Revolution, Rubino et al. (2013) show two inflection points in the 20th century $\delta^{13}\text{C}$ signal: from 1915 CE, and from 1955 CE. There is stabilization of atmospheric CO₂ concentrations at 310–312 ppm from ~1940 to 1955 CE (Fig. 25) (MacFarling Meure et al., 2006); the subsequent rise appears to coincide with increase in fossil fuel use at about 1950 CE (Rubino et al., 2013). The Francey et al. (1999) high-precision record of atmospheric $\delta^{13}\text{C}$ based on Antarctic ice cores shows two straight segments: one between 1850 and 1961 CE, and a steeper segment between 1962 and 1980 CE.

The Law Dome ice core records also show atmospheric methane (CH₄) concentrations ranging from 590 to 760 parts per billion (ppb) through much of the Holocene, up to 1700 CE, with as little as ~55 ppb difference from 1000 to 1700 CE (Ferretti et al., 2005). This is followed by an unprecedented increase to 1700 ppb by 2004 CE (Ferretti et al., 2005, Fig. 27b), an increase of 150% over 200 years (MacFarling Meure et al., 2006). Growth rates in CH₄ decreased during ~1940–1955 CE (MacFarling Meure et al., 2006). The $\delta^{13}\text{C}$ curve for CH₄ shows a marked decrease of ~2‰ from ~1000 to 1700 CE, perhaps in response to reduced biomass burning, and a subsequent abrupt rise from ~1875 CE to the present value of ~–47‰, reflecting increasing pyrogenic emissions (Ferretti et al., 2005, Fig. 27b).

Over the past 200 years, atmospheric N₂O concentrations at Law Dome increased by 21% (MacFarling Meure et al., 2006). N₂O concentrations ranged from 260 to 270 ppb between 1 and 1850 CE, followed by a steady rise to 290 ppb by 1950 CE, from when there was a

Table 10
Reasons for and against using an ice core as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> • Annual layers, high resolution • Extensive distribution, but only Greenland and Antarctic ice sheets likely to persist • Independent dating using ²¹⁰Pb, but not ¹⁴C • Major volcanic eruption events provide clear S spikes • SO₄²⁻ clear spike starts mid-20th century, but diachronous • NO₃⁻ increase and $\delta^{15}\text{N}$ decrease from ~1950 CE • Increased lead accumulation from 1950s, especially from gasoline source • Increased Zn, Cd, and Cu concentrations • Radionuclides, clearly resolved bomb spike • Holocene GSSP precedent in Greenland 	<ul style="list-style-type: none"> • Surficial laminae (recent decades) are more prone to melting by global warming than deeper laminae (Holocene), especially in alpine glaciers and coastal ice sheets • Alpine glaciers melting at rapid rate • Lag between age of ice and younger age of air bubbles (100 years in Greenland, possibly up to 1000 years in Antarctica) • δD and $\delta^{18}\text{O}$ no significant variation across mid-20th century • Increased dust flux not consistently evident • Mid-20th century increased CO₂ and CH₄ concentrations and $\delta^{13}\text{C}$ reduced signal not fixed in air bubbles • Core difficult/expensive to procure and to store safely

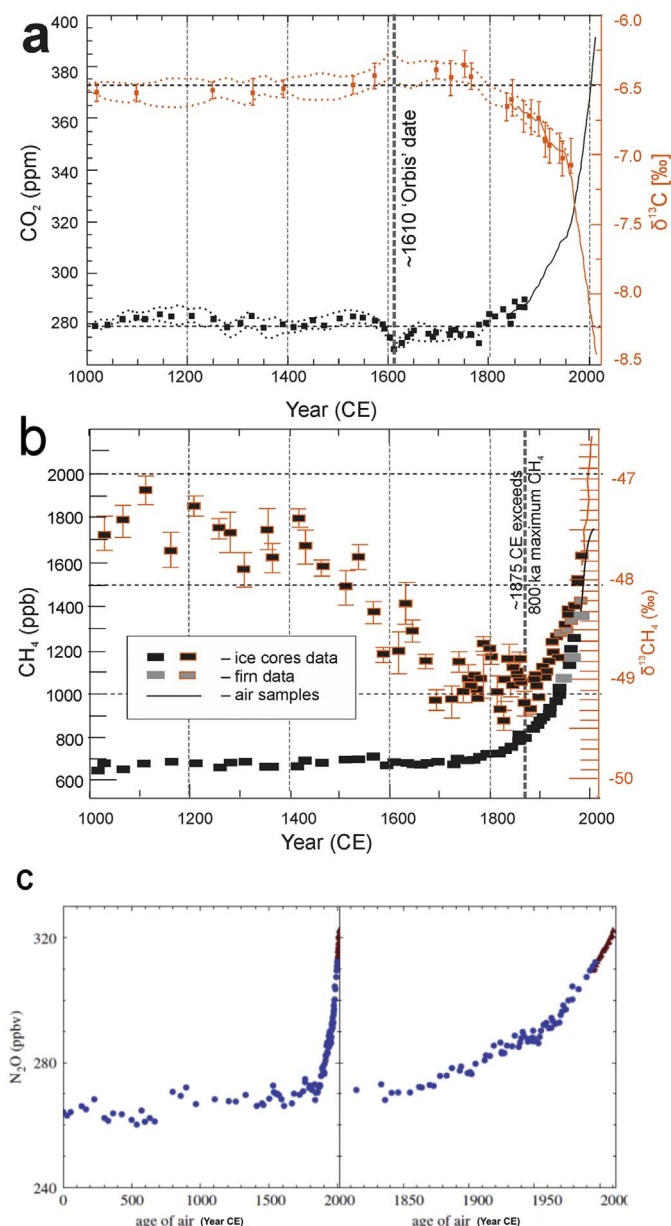


Fig. 27. a) Law Dome ice core and firn records for CO₂ concentration and δ¹³C for atmospheric CO₂ for the past 1000 years (from Rubino et al., 2013); b) Law Dome ice core and firn records for CH₄ concentration and δ¹³C for atmospheric CH₄ for the past 1000 years (from Ferretti et al., 2005); c) N₂O concentrations in the Law Dome ice core over the past 2000 years and 200 years (Wolff, 2013).

rapid rise to 320 ppb by 2000 CE (Wolff, 2013, Fig. 27c). The δ¹⁵N signal in firn ice at Law Dome is more scattered than would be expected, most probably due to random analytical noise (Rubino et al., 2013).

3.8. Speleothems

Annually laminated calcareous speleothems (typically stalagmites) occur within natural cave systems, typical of karst environments, or within artificial tunnels where either the adjacent bedrock or the degradation of mortar in concrete linings of the tunnel contributes the calcium carbonate. Karst landscapes occur across extensive parts of Eurasia, Australasia and the Americas, in total about 10% of the Earth's land mass (Fig. 28). There is precedent in using speleothems for the location of a GSSP, as Walker et al. (2012) used a stalagmite from Mawmluh Cave, India, as the proposed GSSP for the Upper Holocene

Subseries (= proposed Meghalayan Stage) at 4.2 ka.

In addition to lamina counting, the age of older speleothems is typically determined by ²³⁴U–²³⁰Th dating, although this is unsuitable for carbonates precipitated over the past century. Independent dating using ²¹⁰Pb–²²⁶Ra has been undertaken at a few locations (e.g. by Condomines and Rihs, 2006 at St. Nectaire, France, and Bonotto et al., 2012 at Santana Cave in Brazil). Radiocarbon has quite commonly been used to confirm an age model from annual laminae, but does not provide annually resolved dates. If Pu is chosen as a primary tool for correlation, the lack of Pu isotope determination would rule out a speleothem as GSSP for the Anthropocene (Fairchild, 2017), but one could be chosen as an auxiliary site.

Natural speleothem growth is typically slow, on the order of tens to hundreds of micrometres per year. In contrast, speleothems related to hyperalkaline groundwaters, commonly from anthropogenic sources, can grow at a rate of 10 mm per year (Baker et al., 1998), hence potentially permitting fine-scale resolution of environmental changes, although the chemistry will be strongly kinetically modified (Hartland et al., 2014; Newton et al., 2015). ²¹⁰Pb dating of stalactites in Santana Cave (Brazil) shows a longitudinal growth rate of 1.3 mm yr⁻¹ and lateral rate of 0.01 mm yr⁻¹ (Bonotto et al., 2012). A study of hydrothermal stalagmites at St Nectaire (France) using ²¹⁰Pb/²²⁶Ra ratios showed growth rates varying from 5.3 ± 0.5 mm to 2.6 ± 0.2 mm yr⁻¹ (Condomines and Rihs, 2006).

Speleothems record changing environmental conditions in the atmosphere, soil and ecosystem through geochemical signals transmitted from the ground surface, through rock, to the subterranean void. The processes of speleothem formation and their chemistry are detailed by Fairchild and Baker (2012), and their relevance to recognition of Anthropocene signals is given in Fairchild and Frisia (2014) and Fairchild (2017). Principal signals that may be used to recognise a potential Holocene–Anthropocene boundary (Table 2) include shifts in atmospheric ¹⁴C and sulphur (sulphate concentration and δ³⁴S), which relate to global-scale changes to atmospheric chemistry (Fairchild and Frisia, 2014, Fig. 29). Local signals include: 1) variations in growth rates of laminae and δ¹⁸O which relate to air temperature and humidity, 2) δ¹³C as an indicator of deforestation and/or introduction of C₄ plants, 3) development of biomarkers such as a reduction in the ratio of C₂₇/C₃₁ n-alkanes and increase in n-alkanols as observed at Mechara (Ethiopia) reflecting the local introduction of agriculture about 1935 CE (Blyth et al., 2016), and 4) shifts in trace elements and isotope ratios (Fairchild, 2017). Although of local source and probably diachronous in nature, such signals may link to global patterns of environmental change and so contribute to characterising a potential GSSP site. For example, Pb anomalies have been recorded in speleothems in Somerset, UK near mines active in pre-Roman, Roman and mediaeval times (McFarlane et al., 2013). Sulphur pollution from coal combustion increased during the Industrial Revolution and culminated in peak S concentrations and lowered δ³⁴S values in the mid- to late-20th century in developed countries (Wynn et al., 2010, Fig. 29d). However, speleothems are insensitive to the significant increases in atmospheric CO₂ concentrations and reduction in δ¹³C over the past century (Fairchild and Frisia, 2014). Changes in the nitrogen cycle via N abundance or isotopic fractionation related to fertiliser use have not been investigated in speleothems (Fairchild, 2017). Phosphorus is strongly modulated by seasonal vegetational die-back and changed infiltration rates (Fairchild and Frisia, 2014) rather than by fertiliser application.

Transport of solutes via plants, soil and rock commonly results in signals being variably attenuated and delayed (Fairchild and Frisia, 2014; Fairchild, 2017). For example, speleothems in SW France and Gibraltar show ¹⁴C concentrations with differing degrees of attenuation and delay of the peak signal, but the initial rise was consistent at 1958 CE and coincides with changes in atmospheric concentrations (Fairchild and Frisia, 2014). The timing of onset of the ¹⁴C rise in speleothems is typically within 1–2 years of the atmospheric change (Genty and Massault, 1999).

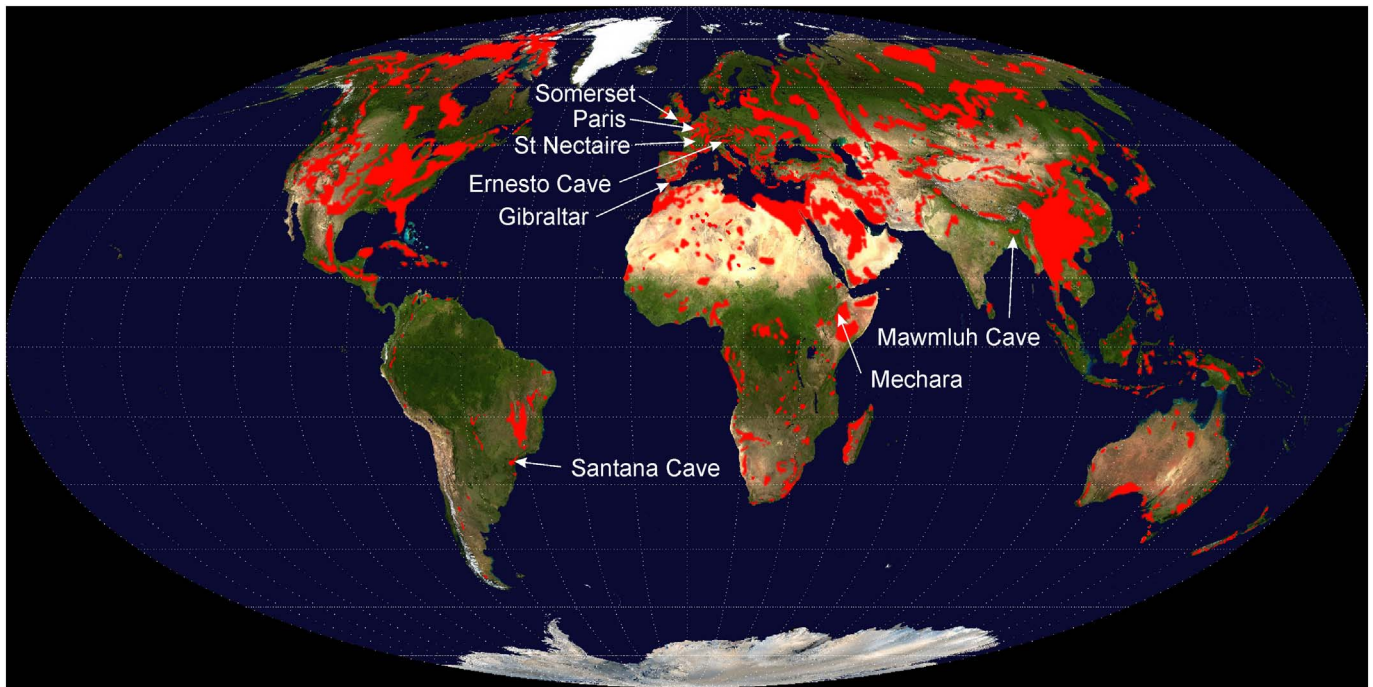


Fig. 28. Extent of carbonate outcrops present across global land mass by Ulrichstill <https://commons.wikimedia.org/w/index.php?curid=9412430> Created 6th February 2010. Karst landscapes and speleothems occur within these regions, and locations mentioned in the text are indicated.

Table 11 summarises the key advantages and disadvantages of such a potential GSSP for the Anthropocene. As examples, the Ernesto Cave in Italy has multi-proxy studies of speleothem laminae within a natural cave system, but urban speleothems may show greater tuning to modified anthropogenic signals.

3.8.1. Ernesto Cave, Italy

Ernesto Cave, northern Italy, closely studied for > 20 years, has supplied various proxy data series from annually-laminated speleothems independently dated using U-Th series (Frisia et al., 2003). These show a strong annual signal of fluorescent laminae associated with a range of colloid-transported elements (Fairchild and Frisia, 2014). A pronounced change is recognised since about 1840 CE, from open and impurity-rich crystals within relatively thin laminae to thicker laminae of clean calcite (Fig. 29a), representing higher growth rates, closely correlated with temperature increases (Frisia et al., 2003; Fairchild and Frisia, 2014) associated with the local ending of the Little Ice Age. This pattern broadly coincides with the initial increase in sulphate concentrations within speleothems in this cave at about 1880 CE (Frisia et al., 2005), associated with the initiation of atmospheric pollution in response to the Industrial Revolution.

No appreciable change in stable isotopes has occurred across the mid-20th century (Fig. 29b), the most significant change in $\delta^{13}\text{C}$ being a marked positive excursion at ~1840 CE (Scholz et al., 2012).

The ^{14}C bomb-spike has been recorded from the cave (Fig. 29c), but with a decade lag compared with atmospheric signals (Fohlmeister et al., 2011). Records of sulphur loading and lowering of $\delta^{34}\text{S}$ in both speleothem and tree rings show a similar lag in the speleothems of about 15–20 years (Fig. 29d), such that in 2000 CE the speleothem was still recording peak atmospheric loading (Wynn et al., 2010, 2014) that was locally experienced in the 1980s (Fohlmeister et al., 2011), but which is now declining (Borsato et al., 2015). Trace element concentrations increased greatly during the early-20th century, a time of deforestation in the area (Borsato et al., 2007).

3.8.2. Urban speleothem, Paris

Rises in several trace metals and rare earth elements are found in urban speleothems in an underground aqueduct beneath Paris that have grown over 300 years with a record of two laminae per year, independently dated using U-Th series (Pons-Branchu et al., 2015). Two distinct phases of heavy metal pollution are recognised: a) 18th century metal contamination resulting from use of urban waste (night soil) as fertiliser; and b) since 1900 CE elevated Pb, Mn, V, Cu, Cd and Al concentrations, attributed to urbanization, with a more marked increase since 1960–1970 CE (Pons-Branchu et al., 2015). $^{206}\text{Pb}/^{207}\text{Pb}$ isotope ratios show no consistent pattern in these speleothems. The dominant value of 1.181 ± 0.003 , is typical of values associated with lead from northern Europe (Pons-Branchu et al., 2015), though one

Table 11

Reasons for and against using a speleothem as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none"> • Annual lamination • Undisturbed locations • Independent ^{14}C markers • S, $\delta^{34}\text{S}$, ^{14}C reflect atmospheric composition, but modified by soil-ecosystem • Solid samples easier to handle than unconsolidated sediment • Precedent of Mawmluh Cave, India, for the proposed Upper Holocene GSSP 	<ul style="list-style-type: none"> • The atmospheric Suess effect on $\delta^{13}\text{C}$ is smaller than fractionation effects in the soil-ecosystem so an inflection corresponding to atmospheric change is not normally present • Although S, $\delta^{34}\text{S}$, ^{14}C may show initial rise as atmosphere changes, the peak is delayed because of storage in soil/ecosystem • Changes in trace pollutant metals (e.g. Pb, Zn, REE) are strongly linked to organic matter from soil and increases tend to reflect soil disturbance/deforestation effects • Not able to reliably detect signals usable elsewhere, e.g. Pu, anthropogenic detritus

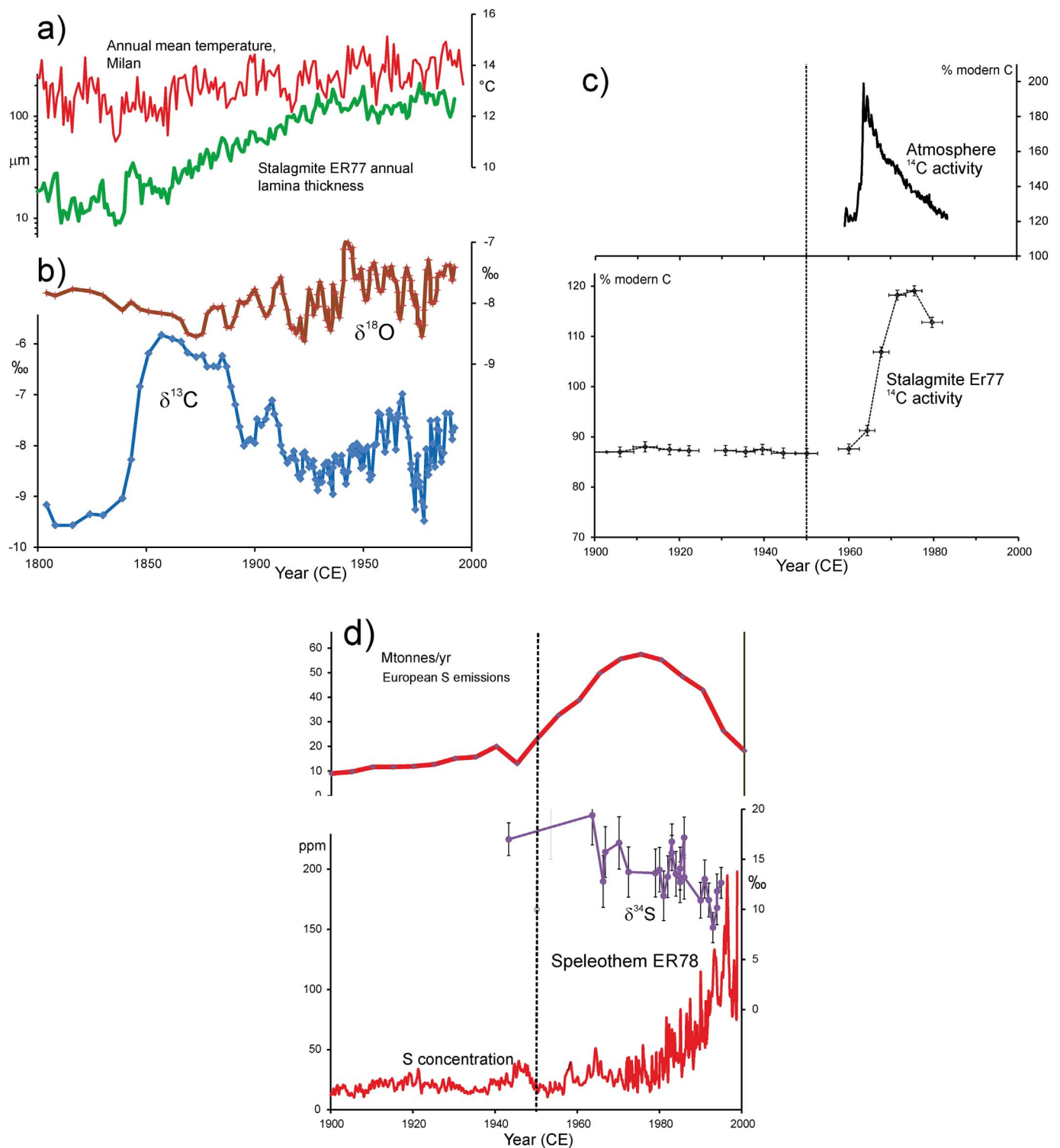


Fig. 29. Ernesto Cave, Italy: a) Age model based on lamina counting related to local mean air temperature (Frisia et al., 2003); b) $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles (Scholz et al., 2012); c) Radiocarbon profile and comparable European atmospheric emissions (Fohlmeister et al., 2011); d) S concentration and $\delta^{34}\text{S}$ (Frisia et al., 2005; Wynn et al., 2010).

speleothem showed a change after the mid-19th century, suggesting a new contribution from coal burning or from Pb from Spain. Pb ratios typical of leaded gasoline are not discriminated (Pons-Branchu et al., 2015).

3.9. Trees

Forest coverage in 2005 CE was about 39 million km², or ~30% of

the world's land area (FAO, 2006) and there are an estimated 3 trillion trees (Crowther et al., 2015). Forests are extensive on all continents except Antarctica, and include the boreal forests of North America and Eurasia and tropical forests of South America, central Africa and south-east Asia (Fig. 30). Tropical trees from aseasonal tropical regions lack reliable annual rings, so trees from temperate or boreal environments would be most suitable for a GSSP as the tree rings themselves can provide a precise annually-resolved chronology.

Tree rings provide a perfect annually-resolved chronology, and within this an annually resolved archive of environmental change, most notably palaeoclimate, at local to global scales. This signal can be evident in the physical properties of the tree ring (width, density, reflectance) or chemical indicators such as the stable isotopic ratios of carbon (sourced from atmospheric CO₂), and hydrogen and oxygen (sourced from precipitation, soil or ground water) (Fritts, 1976; Speer, 2010; McCarroll and Loader, 2004). In many species, the tree rings may be further divided into early wood (formed during the spring) and late wood (formed during the summer, primarily from photosynthates fixed during the growth year). Rings from living trees can provide environmental information spanning both the mid-20th century putative boundary for the Anthropocene and the late-18th century transition from pre-industrial times through the development of the Industrial Revolution. Similar records can also be found in preserved dead trees that accumulated in bogs or lakes, and also in building timbers. Where trees co-existed in time they will co-record a shared environmental signal that may therefore be cross-dated to assign chronology or to provide independent replication between trees or across study regions.

Table 12 summarises the key advantages and disadvantages of having a potential GSSP within tree rings. Proxy signals associated with palaeoclimate, stable carbon isotopes, sulphur concentrations and isotopic ratios, radiocarbon fallout and heavy metal concentrations are discussed below.

3.9.1. Tree rings and palaeoclimate signals

Dendrochronology relies upon the counting and statistical synchronisation of annual changes in tree ring widths, and their assignment to a calendrical or relative timescale. This is possible because the variability in tree growth driving this signal is related to regional climatic changes. Trees growing at the same time in history,

and experiencing a similar climate history, will in the absence of external disturbances yield similar patterns of wide and narrow growth rings that may be cross-dated by dendrochronology to develop a precise timescale. Due to the physical increase in the diameter of a tree as it grows, tree-rings show a characteristic decline in width with increasing tree age (size). Where a climatic signal is to be extracted from the ring width data, statistical techniques are required to remove this growth or age-related trend. This statistical processing may cause climate signals of frequency longer or equal to the age-span of the tree to be lost, reducing the value of tree-ring widths in recording long-term climate change (Cook et al., 1995; Esper et al., 2002). However, this effect can be significantly reduced/removed by using more complex detrending protocols and numerous multi-centennial records of temperature changes from 800 to 1990 CE have been determined from high-elevation and high-latitude locations across the Northern Hemisphere using Regional Curve-based Standardization techniques (Esper et al., 2002; Melvin and Briffa, 2008). These data contain evidence for a Northern Hemisphere Medieval Climatic Anomaly (MCA) between about 900 and 1200 CE, and for below average temperatures typical of the Little Ice Age (LIA) from 1200 to 1850 CE, and for warming since 1850 CE comparable in scale to that of the MCA (Wilson et al., 2016; Esper et al., 2002, Fig. 31a).

The suitability of using tree rings to establish the temperature profile post-1950 CE has been questioned due to a conspicuous increase in the statistical uncertainty in reconstructions after 1975 CE, which may reflect an anomalous reduction in growth performance at some high northern latitude conifer sites, but is also often coincident with a decrease in the replication of the composite record (Esper et al., 2002). This “divergence” between growth-based reconstructions and observed climate has been the subject of significant attention by

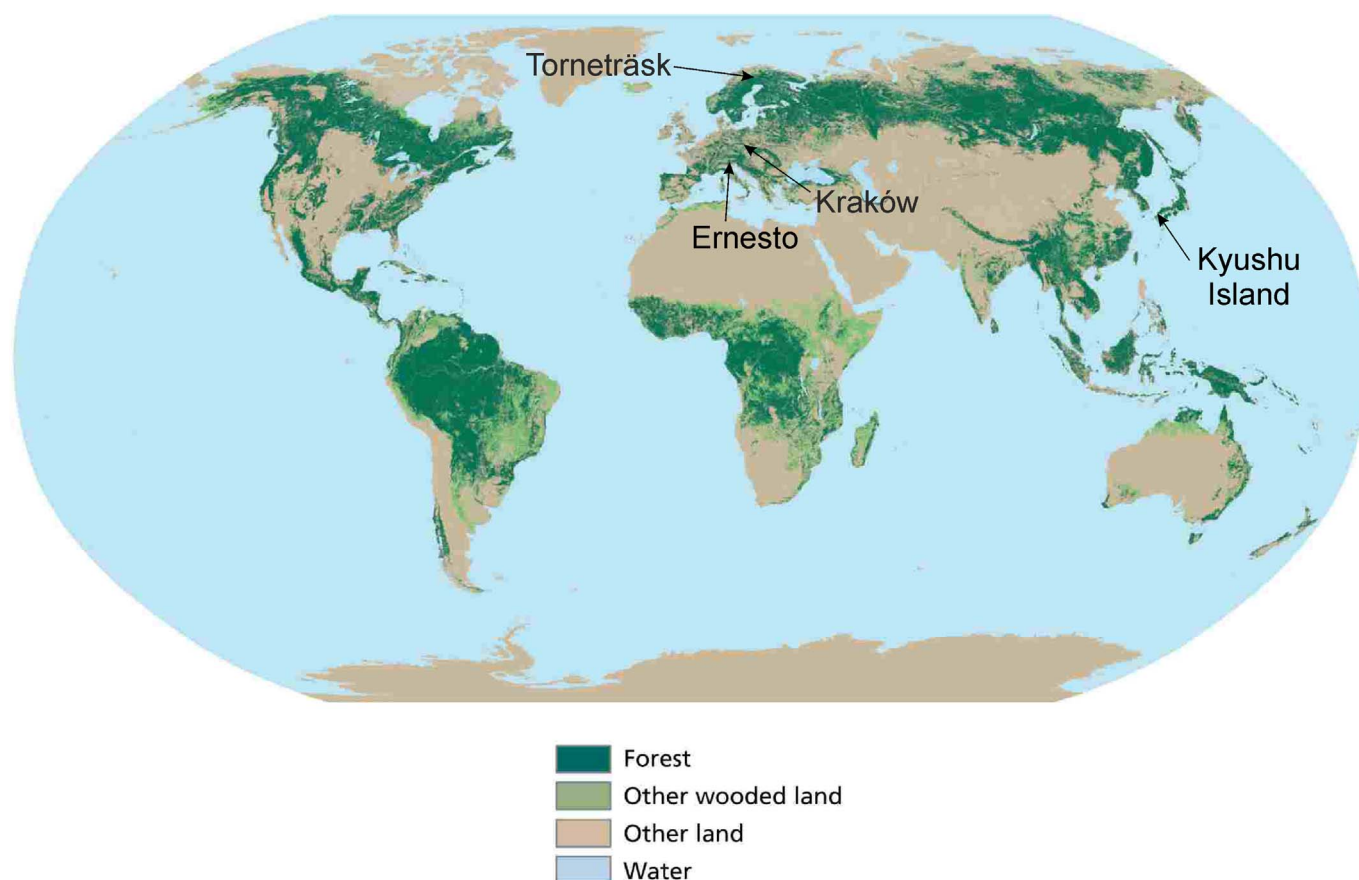


Fig. 30. Forest map of the world with key locations mentioned in the text. © (2006) FAO.

Table 12
Reasons for and against using a tree as a potential host for a GSSP.

For:	Against:
<ul style="list-style-type: none">• Global expression and distribution; widespread terrestrial extent• Bi-annual layers• Precise dating through dendrochronology• Independent ¹⁴C dating possible• Environmental signals preserved in deadwood• δ¹⁸O and δD: record source water, which contains a climate signal• δ¹³C is highly sensitive to the Suess effect on atmospheric CO₂• Sulphur and δ³⁴S signal commencing mid-20th century• ¹⁴C bomb spike clearly defined• High potential for independent replication• Tangible, stable and archivable	<ul style="list-style-type: none">• Growth ring width may not be suitable for picking mid-20th century climate signal (due to local effects or ‘divergence’)• Complex record of local signals in physical and chemical proxies• δ¹³C can be affected by local modifications• Heavy metal concentrations may be unsuitable proxy for pollution events due to radial mobility• No precedent of using a living/deceased organism for GSSP

dendroclimatologists and many potential causes, from unbalanced replication and signal processing effects to changes in seasonality and CO₂ fertilisation, have been considered (D'Arrigo et al., 2007). While this may reduce current confidence in the use of ringwidth-based reconstructions to define climatic changes attributed to the start of the Anthropocene, maximum late wood density appears to be less impacted upon by such divergence issues and density-based reconstructions of past climate remain some of the best records of past climatic variability currently available to climate scientists. There are few palaeoclimate proxies other than tree-rings capable of providing the same precisely dated, annually-resolved palaeoclimate reconstructions with the same

capacity for independent verification through replication, capable of being objectively calibrated against instrumental observations and independently verified.

3.9.2. Stable carbon isotopic signal

Stable isotope variability in tree rings appears to be less affected by non-climatic age-related tendencies (Young et al., 2011). Long-term trends of stable carbon isotopic data in tree rings may therefore mark the start of the Anthropocene in the tree ring archive. The stable carbon isotopes in tree rings record primarily the stable isotopic composition of the source CO₂ sampled by the tree during photosynthesis. This signal is

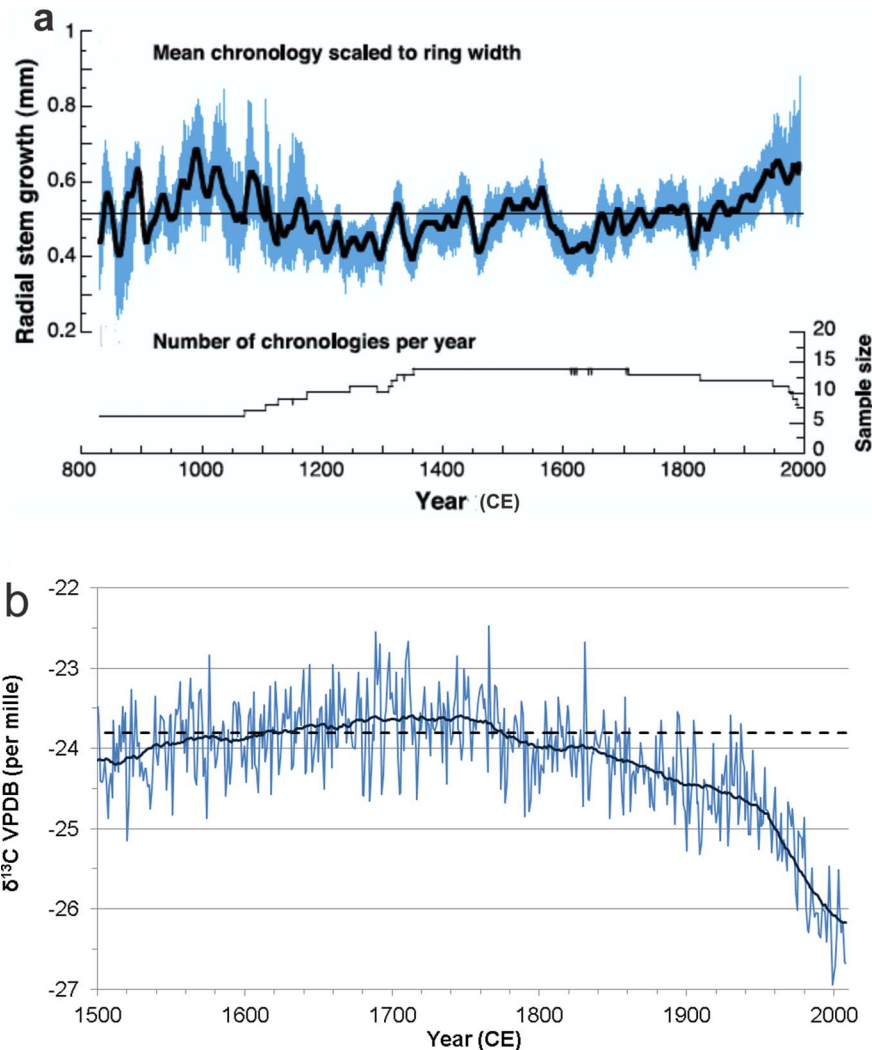


Fig. 31. a) The 20-year smoothed Northern Hemisphere extra-tropics reconstruction of radial stem productivity in high elevation and high latitude forest environments since 800 CE (black) and two-tailed 95% bootstrap confidence intervals (blue) (from Esper et al., 2002); b) δ¹³C variability from Loader et al. (2013a) for the period 1500–2008 CE measured in tree-ring cellulose for a composite tree ring stable isotope chronology developed using *Pinus sylvestris* trees from northern Fennoscandia. Fine line represents annually-resolved δ¹³C variability, thick solid line presents the annual data smoothed with a centrally-weighted 51-year moving average. Dashed line represents the mean δ¹³C value for the “pre-industrial” period 1500–1799 CE. Mean annual replication for the record is > 13 trees. Analytical precision σ_{n-1} = 0.12 per mille n = 951 which compares favourably with the analytical precision of the method typically reported (σ_{n-1} = 0.10 n = 10) (Boettger et al., 2007; McCarroll and Loader, 2004; Loader et al., 2013a).

further imprinted upon by the tree's physiological response to climatic change (the balance between stomatal conductance and photosynthetic rate), photosynthetic and respirative processes (Farquhar et al., 1982; McCarroll and Loader, 2004, and references therein). Inter-annually, $\delta^{13}\text{C}$ values will vary in response to changes in irradiance, soil moisture, precipitation and relative humidity and this has allowed the tree rings to be used as a record of past climate variability. As a biological system, tree ring stable carbon isotopic ratios will also vary depending upon the species of tree (angiosperm versus gymnosperm) and their individual environmental history. In addition, factors such as leaf morphology, ecological amplitude, pollution, forest management history, disturbance and nutrient availability, can all influence the resulting tree ring isotopic record, but at all times, the basic physiological response of the tree will remain the same. This natural inter-tree variability has been characterised in resampling experiments (Loader et al., 2013b), and although extreme cases have been reported (Li et al., 2005), natural intra-site variability falls within reasonable limits defined by current analytical uncertainties.

When viewed over a long (centennial-millennial) timescale, trees have the capacity to record and contextualise the unprecedented anthropogenic changes in atmospheric $\delta^{13}\text{C}$ associated with global industrialisation and land-use change. By combining records from different geographical regions, location-specific year-to-year climatic variability and the effect of random disturbances may be reduced to yield a regionally relevant record of carbon isotopic variability in atmospheric CO_2 . A reduction of atmospheric $\delta^{13}\text{C}$ values by $\sim 2\text{‰}$ has occurred since the start of the global Industrial Revolution ~ 1820 CE (the $\delta^{13}\text{C}$ Suess effect), with marked inflection at ~ 1950 CE (Fig. 31b). This effect is observed in tree rings worldwide and is also clearly observable in ringless trees from the aseasonal tropics (Loader et al., 2011).

The $\delta^{13}\text{C}$ Suess effect is commonly removed in tree ring carbon isotope series through mathematical detrending without loss of low-frequency climatic information (McCarroll and Loader, 2004). This makes the data suitable for numerical calibration against instrumental climate data, but prevents its use as an arbiter of atmospheric $\delta^{13}\text{C}$ changes. To assess the potential base of the Anthropocene using $\delta^{13}\text{C}$ changes, raw carbon isotope data are required.

A further consideration when assessing the candidature of tree ring $\delta^{13}\text{C}$ as a marker for the Anthropocene is the influence of increased atmospheric CO_2 concentration on photosynthesis. This “fertilisation” or “greening” effect has been widely reported and in the context of stable carbon isotopic variability is seen as a change in the intrinsic water-use efficiency of the tree (Seibt et al., 2008; Saurer et al., 2014; Frank et al., 2015). Methods have been developed to quantify and to correct for this effect, which will vary from tree to tree (Treydte et al., 2009; McCarroll et al., 2009). Several studies have also identified a limit to the degree to which trees can adapt to higher atmospheric CO_2 concentrations. This “fertilisation” effect does not negate the use of $\delta^{13}\text{C}$ as a potential marker, but it should be taken into consideration.

A large-scale/global composite of global carbon isotope time-series from tree rings could help locate the boundary of the Anthropocene, as individual records are affected by local climate, forest disturbance, or the local impact of atmospheric pollution on stomatal conductance and disease (McCarroll and Loader, 2004 and references therein, Bukata and Kyser, 2007, Treydte et al., 2007, 2009, Loader et al., 2013b, Boettger et al., 2014, Saurer et al., 2014, Frank et al., 2015), all of which could potentially obscure the true expression of the global $\delta^{13}\text{C}$ Suess effect in tree rings. Nevertheless, the change in tree ring carbon isotopic values resulting from global industrialisation is far greater than anything observed in the tree ring isotope record during the last 1100 years. The combination of a temporally stable proxy preserved within a precisely constrained chronology for the tree rings represents an opportunity to constrain a boundary to the Anthropocene.

3.9.3. Sulphur concentrations and isotopic ratios

Bulk analyses of the conifer *Abies alba* from near the Ernesto cave site, NE Italy, show that S concentrations within annual rings (Fig. 32c) reflect atmospheric SO_2 pollution (Fig. 32b), with significantly higher values in the second half of the 20th century (Fairchild et al., 2009). Both the trees and the nearby stalagmite record the overall increasing trend in the 20th century, from the 1960s onwards, with the trees showing a more immediate (earlier) response to the high S emissions of the 1960s and 1970s (Fairchild et al., 2009; Wynn et al., 2014). The *Abies* samples show considerable noise, including variation between individual trees (Fairchild et al., 2009). Furthermore, the S is mainly sourced from soil-waters rather than directly from the atmosphere, and the S is stored both in soil and biomass, which means that the original atmospheric signal is necessarily modified (Fairchild et al., 2009). These trees also show a trend in $\delta^{34}\text{S}$ towards lighter isotopes in recent rings, with pre-industrial values of $+7.5\text{‰}$ and modern values of $+0.7\text{‰}$, with no significant signature corresponding to the mid-20th century (Wynn et al., 2014, Fig. 32d).

$\delta^{34}\text{S}$ is potentially a useful chronological proxy for past atmospheric S pollution, particularly as there is limited fractionation during incorporation into the tree (Kawamura et al., 2006). Three conifers studied from Kyushu Island, Japan provide a significant record from 1945 CE, where a dominant organically-bound sulphur (OBS) fraction and a minor water-soluble sulphur (WSS) fraction were discriminated (Kawamura et al., 2006). These showed a general increase, then decrease, in the second part of the 20th century, with higher values and a greater range seen in urban as compared with rural areas. Increased atmospheric S emissions from petroleum combustion with negative $\delta^{34}\text{S}$ values were mainly sourced from the Middle East since the 1950s, until legislation was introduced in 1968 CE. There is a ~ 5 year delay in $\delta^{34}\text{S}_{\text{OBS}}$ tree ring values compared with maximum atmospheric S concentrations, probably reflecting the time taken for sulphur deposited in soil to mobilize to the roots, and for the metabolism of sulphur after absorption (Kawamura et al., 2006).

3.9.4. The radiocarbon bomb spike and heavy metal concentrations

Atmospheric $\Delta^{14}\text{C}$ decreased by about 20.0‰ between 1890 and 1950 CE in response to the Suess effect of adding low-activity fossil carbon, punctuated by a marked bomb spike in response to nuclear weapons testing, peaking in ~ 1963 CE that led to a level two times higher than that of pre-1850 CE times (Rakowski et al., 2013). Annual rings of a pine tree (*Pinus sylvestris*) taken in the Niepołomice area, near Kraków, Poland from 1960 to 2003 CE show a clear $\Delta^{14}\text{C}$ bomb-spike in 1964 CE (Rakowski et al., 2013, Fig. 32a). A suitable GSSP candidate should have a data set of both $\Delta^{14}\text{C}$ and $\delta^{13}\text{C}$ extending back to pre-industrial times; the data in the tree analysed by Rakowski et al. (2013) began close to the peak bomb-spike interval, so lack an adequate record of the transition across a potential Holocene–Anthropocene boundary. $^{239} + ^{240}\text{Pu}$ records in tree rings are less studied, but the distribution of Pu in tree rings and in lake sediment cores seems to show similar trends (Mahara and Kudo, 1995) suggesting Pu as a potentially suitable signal.

Historical changes in trace metal levels can be recognised in some tree ring records, although there may well be a time lag between metal deposition and the passage through soils and root system. Element concentrations in tree rings for a given year may differ markedly between trees from a single location (Watmough, 1999). Studies on the sugar maple (*Acer saccharum*) suggest that there is both a rapid uptake of trace metals (e.g. Cu, Ni, Cr, Zn, Cd, Co and As) and a minimal lateral movement between rings (Watmough, 1999), although radial movement of trace elements has been reported in other studies (Hietz et al., 2015).

4. Summary

Many of the signals described above are highly resolved, widespread and correlatable. Notably, heavy metals, $\delta^{13}\text{C}$ and radiogenic fallout

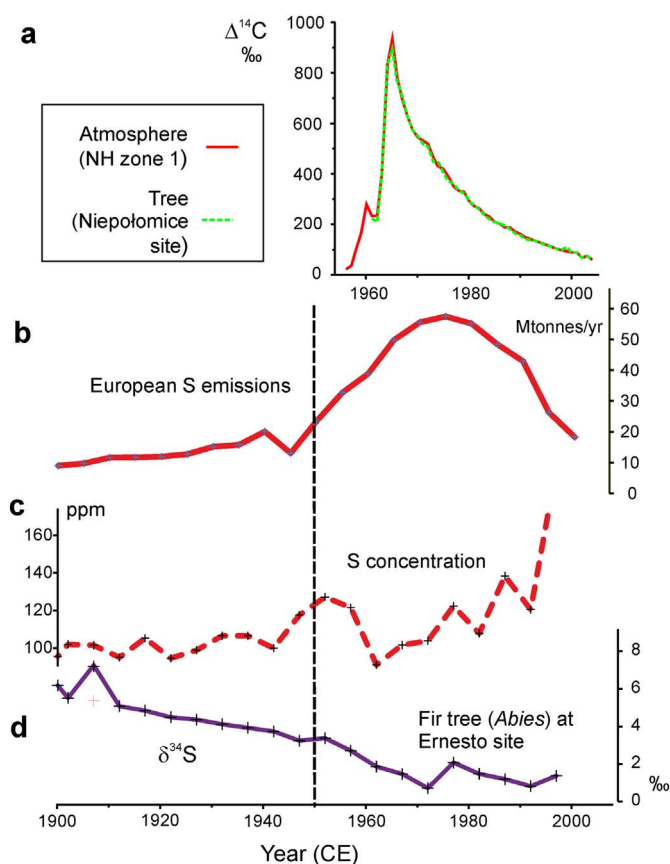


Fig. 32. a) Tree rings (*Pinus sylvestris*) samples from 1960 to 2003 CE at Niepołomice (Poland) showing changes of radiocarbon concentration (from Rakowski et al., 2013) compared with Northern Hemisphere (Zone 1) atmospheric values (from Hua and Barbetti, 2004); b) European S emissions; c) S concentrations; and d) $\delta^{34}\text{S}$ in *Abies alba* from NE Italy compared with European S emissions (Wynn et al., 2014).

signals tend to be recorded in most environments that might conceivably be suitable for hosting a GSSP (Table 2). Other airborne signals, such as fly ash, nitrates and $\delta^{15}\text{N}$, and to a lesser extent sulphur and sulphates, CO_2 and CH_4 concentrations and $\delta^{18}\text{O}$ occur in many environments and provide additional means of correlation. Biotic assemblages have been demonstrated to respond rapidly to local environmental changes and can provide useful secondary and local markers, but seem on current evidence unlikely to provide a globally correlatable and synchronous marker that is as temporally well-resolved as $\delta^{13}\text{C}$ and radiogenic fallout or other airborne signals.

These more nearly synchronous and widespread signals, combined with the possibility of directly dating successions by such means as lamina counting, show that a well-chosen GSSP level might be precisely dated to the nearest year (or season) and might be correlated to other well-studied high-resolution sections with a resolution of a year or a few years. In less well-studied sections, or those lacking annual lamination, use of the very wide range of available proxy signals should still be able to provide correlative resolution of a few to several years, sufficient to widely locate a practically effective Holocene–Anthropocene boundary, even in apparently unpromising geological settings. For instance, Smith et al. (2016) identified what would be regarded as an Anthropocene succession of sub-ice shelf deposits at Pine Island in West Antarctica, by identifying bomb-derived plutonium within the deposits.

The location of such a Holocene–Anthropocene boundary within relatively recent successions – of the last few centuries – we regard as having use beyond the formal value of the establishment of chronostratigraphic divisions. An extensive, variably interconnected suite of physical, chemical and biological changes to the Earth System has been

taking place, some local, some global, many of which leave stratigraphically detectable traces. These include phenomena such as the spread of various kinds of industrially related pollutants and their biotic effects, changes to the atmosphere and hence to the Earth's radiative properties, a complex pattern of species extinctions, extirpations and invasions, and the emergence and evolution of technological systems on a planetary scale. For practical analysis of these changes – many of which are ongoing, or accelerating – placing them in an effective framework of space and time is a prerequisite. We suggest that a chronostratigraphic Holocene–Anthropocene boundary could make a significant contribution to this analytical process, not least in that this time interval is emerging as a critical transition in the geological history of this planet (e.g. Steffen et al., 2016). The survey we provide here suggests that such a boundary would be workable in practice, across the range of the environments we survey here and likely more widely. Table 13 surveys the range of chemostratigraphic signals recorded here and provides their onset and peak date for key environments. This shows the wide range of potential proxies that could be used to mark a mid-20th century boundary.

With regard to the environments most suitable for seeking a candidate GSSP, we offer the following guidance:

- Anthropogenic deposits are common in terrestrial environments. They may be relatively thick (up to tens of metres), contain signals that closely reflect human influence, and are readily datable, particularly using technofossils. They tend to lack lateral continuity, may contain omission surfaces, are not annually laminated, and may lack 'preservability', all of which are likely to discount many such deposits as hosts for primary or auxiliary GSSP candidates. However, much the same criticisms also apply to many Holocene deposits. Suitable potential GSSP candidates may well exist within anthropogenic deposits, or perhaps at the interface between anthropogenic and non-anthropogenic strata, although none have yet been identified or described. More work is required in this area of investigation.
- Marine anoxic basins tend to display undisturbed annually resolved laminae that can be independently dated. The radionuclide fallout signal may show some smearing due to processes affecting the settling of radioisotopes through the water column, which can inhibit development of a sharp bomb spike. Even so, the early 1950s start of the signal (Table 13) is typically robust. In these environments, heavy metals including Pb, organic compounds, such as PCBs and pesticides (Table 13), microplastics and biotic signals, including planktonic foraminifera and diatoms (and their stable isotope patterns), tend to be controlled by local environmental variations. Although they broadly display changes in the mid- to late-20th century, the onset of their signal can vary by decades between regions. Low sedimentation rates may reduce the thickness of Anthropocene strata to a few centimetres. Even so, the signals of change may still be clearly resolved and laterally extensive. This environment has potential for hosting an Anthropocene GSSP, provided that the fallout signal, especially ^{239}Pu , is not significantly smeared by slow settling rates or by bioturbation.
- Coral growth bands can exhibit very-high resolution records, which, along with rapid growth rates commonly exceeding sediment accumulation rates, provide a potentially suitable medium for hosting a GSSP. The onset of the ^{239}Pu signal is particularly robust as an early 1950s marker (Table 13), with no concerns about the influence of settling rates through the water column; peak signals are more variable between localities. However, reefs that are distal to nuclear detonation test sites might be more suitable so as to more clearly show the global pattern, rather than including some residues of the early, pre-1952 CE tests that only left a local signal. The Caribbean, Red Sea, western and northern Indian Ocean are distant from the main testing grounds (Waters et al., 2015) and are more likely to most clearly show the global atmospheric signal. Radiocarbon tends

Table 13

Summary of key mid-20th century proxy signals and potential palaeoenvironments for a GSSP. The initial date represents the marked onset of the signal; the peak signal is shown in brackets. Key signals and environments are shown in boxes. Reference numbers relate to main geographical locations mentioned in the text: (1) Santa Barbara, USA; (2) Caribbean; (3) Nova Scotia, Canada; (4) Clyde Estuary, Scotland; (5) Urola Estuary, Spain; (6) Lake Victoria, Australia; (7) North America; (8) China; (9) Lochnagar, Scotland; (10) Lilla Öresjön, Sweden; (11) Northern England; (12) Switzerland; (13) Antarctica; (14) Greenland; (15) Ernesto Cave, Italy; (16) Fenno-Scandinavia; (17) Poland; (18) Ontario, Canada.

Marker Environment	Black Carbon	Fly ash	Lead	PCBs	Pesticides (DDT)	NO ₃ ⁻	δ ¹⁵ N	Sulphur	SO ₄ ²⁻	CO ₂	CH ₄	δ ¹³ C	δ ¹⁸ O	D &/or dust	²³⁹ Pu	¹⁴ C
Marine anoxic basins (1)	?	?	1965 (1970s)	1945 (1967)	1952 (1967)	?	x	x	x	x	x	?	?	x	1950-54 (1970s)	?
Coral bioherms & marine bivalves (2) (3)	?	?	Late 1940s (1970s) (2)	?	x	~1970 (3)	~1950 (3)	x	x	x	x	1955 (2)	~1950 (2)	x	1952 (1964) (2)	1958 (1972) (3)
Estuaries/Deltas (4) (5)	?	1900s (5)	1915 (4)	1950s (1965-1977) (4)	?	?	x	x	x	x	x	?	?	x	?	?
Lakes (6) (7) (8) (9) (10)	1940s-1950s (2004-6) (8)	~1950 (1970-1990)	1960s (Post-2000) (8)	~1960 (9)	1950s (9)	?	~1950 (7)	~1950 (10)	x	x	x	?	?	x	~1955 (1964) (6)	?
Peat & peat-lands (11) (12)	x	1950s (1970s) (11)	1810 (1979) (12)	(1960-1976) (12)	x	x	x	x	x	x	x	?	x	x	x	x
Ice (13) (14)	~1880 (~1910) (14)	?	~1940 (~1970) (14)	?	?	~1950 (14)	~1950 (1980) (14)	~1940 (~1970) (14)	~1940 (14)	~1950 (13)	~1950 (13)	1955 (13)	1980s (2010) (14)	~1850 (13)	1953 (1962) (13)	1954 (1966) (13)
Speleothems (15)	x	x	?	?	?	?	?	1980 (2000)	1880 (15)	x	x	~1840 (1860)	1875 (1940)	x	x	~1965 (~1975)
Trees (16) (17) (18)	x	x	?	?	?	?	1945 (18)	1960s (15)	?	x	x	1940 (1750) (16)	?	1960	?	1956 (1964) (17)

to have latitudinal diachroneity and shows a delayed signal in shallow-water corals of about a decade compared with atmospheric ¹⁴C and the ²³⁹Pu signal (Table 13), with increased lags also associated with greater water depth of coral growth. The decline of δ¹³C in corals is most marked in Atlantic corals from 1957 CE (Table 13), with the accumulation rate being more variable and of much smaller magnitude in the Indian and Pacific oceans. δ¹⁵N values are complex, related to source areas of ocean currents and types of anthropogenic sources. Heavy metals can be concentrated within certain species of coral, but they tend to record only local contamination events. Temperature records are a mix of local and global factors and a single core would probably not be suitable as a GSSP. If corals are to be considered as a potential host for a GSSP, examples from the Caribbean, using ²³⁹Pu as the primary signal, may be most suitable.

- Estuaries and deltas are typically responsive to anthropogenic influence, whether biotic change in response to environmental modifications and widespread introduction of neobiota, changes to sediment flux through alteration of catchment erosion rates or impounding sediments through dam construction, or through contamination by fly ash, heavy metals such as Pb, organic chemicals such as PCBs (Table 13), or microplastics. These environments, while offering good potential for hosting a candidate GSSP, may suffer from strong modulation by local influences, a lack of lateral continuity, and the common presence of omission surfaces.
- Lakes, despite being laterally disconnected, are found across large parts of the planet and, although commonly strongly modulated by local processes, also display numerous globally coherent signals. As with estuaries and deltas, lakes tend to be responsive to anthropogenic influence, resulting in a spectrum of highly resolved signals,

including black carbon, fly ash, organic compounds, δ¹⁵N, S and ²³⁹Pu which provide potential 1950s markers (Table 13). Despite the Anthropocene successions being typically thin, shallow perennial lakes are advantageous for hosting a potential GSSP given that settling rates of radiogenic fallout and heavy metals will be minimal and at the same time there should be few omission surfaces. The latter issue limits the suitability of many arid saline lakes, but numerous Northern Hemisphere meromictic lakes could host suitable candidate GSSP or auxiliary sections. Lake sediments have been accepted as hosts for global auxiliary stratotypes for the base of the Holocene.

- Ombrotrophic peat bogs provide widespread local archives that faithfully record aerosol distribution from local, regional and global sources. The range of potential suitable signals are diverse (Table 8), with fly ash providing the most robust 1950s marker (Table 13). The behaviour of trace elements may be influenced by changing redox conditions and has to be carefully documented. As there is no varved sedimentation, and as some signals such as the Pu bomb spike show redistribution effects, hosting a GSSP in peat bogs seems not appropriate, but this environment may provide a useful auxiliary stratotype.
- Glacial ice provides a broad spectrum of annually resolvable atmospheric signals including radionuclides, δ¹⁸O, sulphates, nitrates, δ¹⁵N, Pb and other metals, and so may provide a suitable mid-20th century marker (Table 13). Given the precedent of locating the Holocene GSSP in glacial ice, there is a strong argument for locating in an ice sheet at least an auxiliary stratotype section for the base of the Anthropocene, not least to show how its signals contrast with those at the base of the Holocene. A significant issue for the temporal resolution required for the Anthropocene is the significant

time lag between the age of the ice and the included air bubbles. This lag affects CO_2 , CH_4 , $\delta^{13}\text{C}$ and N_2O values, which are consistently younger than the enclosing ice. This effect is most pronounced where snow accumulation is slowest, so is typically more of an issue in Antarctica than Greenland. Even so, it has been recently demonstrated that the age of air bubbles can be matched to that of the associated ice through the use of nitrogen isotopes (Parrenin et al., 2013), which may circumvent what until recently appeared to be a significant problem. Slowly accumulating Antarctic ice tends to show only the more global signals, whereas Greenland ice can reveal many regional Northern Hemisphere influences. Nevertheless, Antarctic coastal ice cores do accumulate at high rates (e.g. Law Dome) and may provide an appropriate GSSP candidate section. Care would have to be taken to locate such a prospective GSSP in a region where the danger of calving of significant areas of ice from ice shelves and ice loss due to global climate change is less imminent, given the shallow depths at which the signal would be evident. Ellsworth Land, at the base of the Antarctic Peninsula might provide a suitable location, since the observed increase in snow accumulation there during the late 20th century is unprecedented in the context of the past 300 years (Thomas et al., 2015).

- Speleothems suffer from modulation of environmental signals by the soil and ecosystem, making the interpretation of signal variation complex. Sulphur loading, the lower $\delta^{34}\text{S}$ values and the ^{14}C (but importantly not ^{239}Pu) bomb-spike can be clearly resolved in speleothems, with an onset close to the atmospheric signal, although the peak tends to show a decadal-scale lag (Table 13). These environments also tend to lack mid-20th century $\delta^{13}\text{C}$, $\delta^{18}\text{O}$ and heavy metal shifts as potential environmental markers.
- Tree rings, compared with speleothems, are more responsive to and provide a clearer archive of changing atmospheric chemistries. They benefit from a widespread distribution and are capable of precise calendrical dating through dendrochronology without reliance upon external dating methods. Stable carbon isotopes provide a strong “candidate” measure from tree rings as they have been demonstrated to record worldwide the change in atmospheric $\delta^{13}\text{C}$ since the onset of global industrialisation. Such a uniformly expressed, well-characterised ~1940 CE departure (Table 13) has not been observed in the older tree-ring record, and the co-recorded chronology enables this trend to be clearly recognised. Sulphur loading and reduction of $\delta^{34}\text{S}$ values are evident in developed countries, with a lagged signal starting in the 1960s. The enrichment of excess ^{14}C from the early 1950s to a 1964 CE peak (Table 13) provides a clear marker. Although certainly widespread, tree-ring signals can be influenced by complex local environmental processes, yet these may be resolved through a capacity for large-scale independent replication. There is no precedent for using a living or deceased organism (such as a tree) as the host for a GSSP, but as an archive containing a signal of atmospheric carbon isotope variability and precise timescale there are close parallels to the use of the North Greenland ice core (NGRIP) for establishing the base of the Holocene.

To summarize, the information provided here is a first overall appraisal of suitable palaeoenvironments as regards their overall potentials for hosting an Anthropocene GSSP and a range of auxiliary sections. Clearly, many candidates that are geologically suitable exist, and are found in a range of different depositional settings. The review presented here is a critical prelude to help guide the necessary next steps, which involve selecting a small number of maximally suitable sites from the most appropriate geological facies, and subjecting them to multi-proxy analysis in order to assess changes in stratigraphically-relevant proxies from pre-industrial times through to the present.

5. Conclusions

- Many widespread, correlatable, and highly resolved signals could be used as the primary marker and as secondary markers for the base of an Anthropocene Series. $\delta^{13}\text{C}$ and radionuclide fallout signals are applicable across most environments. The latter provides the most abrupt signal, with a marked upturn in abundance of radioisotopes of ^{239}Pu or ^{14}C , in 1952 and 1954 CE respectively, appearing to provide a consistent horizon for correlation, whereas the peak ‘bomb-signals’ appear more diachronous.
- Varved deposits, such as those accumulating in marine anoxic basins, meromictic lakes, and estuaries and deltas, along with layers in glacial ice and growth rings, such as developed in corals, marine bivalves, speleothems and trees, represent those palaeoenvironments yielding highest resolution ‘stratigraphy’ at annual or even seasonal scale, and are hence preferred targets for searching for a likely GSSP section. The use of lamina or layer counting to determine ages can be verified, precisely in the case of tree rings, or radiometrically via ^{210}Pb (marine anoxic basins, corals and bivalves, lakes, ice) and ^{14}C (marine anoxic basins, corals and bivalves, lakes, speleothems and trees) dating.
- Ombrotrophic peat bogs and some anthropogenic deposits provide clear expression of numerous key stratigraphical markers and can show high stratal accumulation rates, but lack the annual lamination seen in other environments.
- Preservation of continuous laminated successions without missing ‘strata’ is essential. Estuaries and deltas are prone to missing laminae through erosive events, ice laminae from high-altitude glaciers through melting events, and speleothems through periods of aridity.
- Time lags between signal generation and recording within varves or laminae on decadal scales represents a significant correlatory problem in speleothems.
- A lag between the age of ice and the younger age of air bubbles within the ice limits the potential use of atmospheric CO_2 and CH_4 signals from ice cores, but sites of rapid accumulation on the Antarctic Peninsula may be suitable, as well as sites in other coastal regions (Law Dome in Antarctica), or in Greenland.
- Deep marine environments can display a significant settling delay that affects radioisotopes and heavy metal signals, though this effect mainly delays the peak signal rather than onset. Except for some marine anoxic basins and perhaps cold-water corals, the rates of sedimentation are too low and the rates of bioturbation are too high for the development of suitable sections for a GSSP.
- Proximity to the source of signals can result in local discrepancies in the timing of their initial onset within successions. In the search for a GSSP, there is advantage in looking at locations that are distant from such perturbations and provide a global signal, e.g. Antarctica for glacial ice, or the Caribbean for corals.
- Overall, the range of high-resolution stratigraphic proxy signals, and the varied and widespread nature of the palaeoenvironments where they can be systematically preserved in strata, suggests that there is excellent potential for locating an effective candidate Anthropocene GSSP together with auxiliary stratotypes.

Acknowledgements

Lisa Barber is thanked for the drafting of Fig. 5. This paper is a contribution of the Anthropocene Working Group (AWG), a constituent working group of the Subcommission on Quaternary Stratigraphy of the International Commission on Stratigraphy. Lucy Edwards and Jim Rose are thanked for their helpful reviews of the original manuscript.

Declaration of interest

The AWG receives no direct funding to carry out its research, and the authors declare no competing financial interests.

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