Quaternary Science Reviews 106 (2014) 47–66

Contents lists available at ScienceDirect

Quaternary Science Reviews

journal homepage: www.elsevier.com/locate/quascirev

The importance of independent chronology in integrating records of past climate change for the 60–8 ka INTIMATE time interval

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ARTICLE INFO

Article history:
Received 17 November 2013
Received in revised form
20 June 2014
Accepted 5 July 2014
Available online 28 August 2014

Keywords:
Quaternary
Chronology
Dating methods
Palaeoclimatology
INTIMATE

ABSTRACT

This paper provides a brief overview of the most common dating techniques applied in palaeoclimate and palaeoenvironmental studies including four radiometric and isotopic dating methods (radiocarbon, ^{230}Th disequilibrium, luminescence, cosmogenic nuclides) and two incremental methods based on layer counting (ice layer, varves). For each method, concise background information about the fundamental principles and methodological approaches is provided. We concentrate on the time interval of focus for the INTIMATE (Integrating Ice core, MArine and TErrestrial records) community (60–8 ka). This dating guide addresses palaeoclimatologists who aim at interpretation of their often regional and local proxy time series in a wider spatial context and, therefore, have to rely on correlation with proxy records obtained from different archives from various regions. For this reason, we especially emphasise scientific approaches for harmonising chronologies for sophisticated and robust proxy data integration. In this respect, up-to-date age modelling techniques are presented as well as tools for linking records by age equivalence including tephrochronology, cosmogenic ^{10}Be and palaeomagnetic variations. Finally, to avoid inadequate documentation of chronologies and assure reliable correlation of proxy time series, this paper provides recommendations for minimum standards of uncertainty and age datum reporting.

1. Introduction

It is commonly accepted that the wealth of information about climate change and environmental responses recorded in various sedimentary deposits can only be adequately utilised when it is placed into a robust chronological framework. In recent decades the demands for precise and accurate chronologies has rapidly increased since it has been realised that climate changed not only on time scales of tens or hundreds of thousands of years, but that climate changes even occurred over less than a human lifetime. Therefore, in modern Quaternary science, information on the timing of past changes is needed, i.e. how much time passed during a shift from one climatic state to another or between different climatic shifts. This will further allow investigation of various mechanisms of climate change in relation to the time scales on which they occur.

Another emerging challenge for palaeoclimate research, and in particular for the INTIMATE community, is tracing potential leads and lags in regional climate response to global change and their possible driving mechanisms. Since such regional disparities in climate change are likely to be in the range of sub- to multi-decadal rather than millennial time scales, an extremely precise correlation of different records from different regions is essential for detection of such leads and lags. However, currently records are often still synchronised through wiggle-matching of proxy data based on the assumption that climatic changes always occur contemporaneously,

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http://dx.doi.org/10.1016/j.quascirev.2014.07.006
0277-3791/

although it has been demonstrated that climate change may be spatially and temporally time-transgressive (Lane et al., 2013). Moreover, climate records derived from different proxies are often wiggle-matched neglecting different temporal responses by those proxies (e.g. pollen in terrestrial records and stable isotopes in Greenland ice cores) with some proxies responding to climate shifts more rapidly than others do. In practice, the concept of assuming general synchronicity of climate change and proxy response to climate changes prevents the tracing of potential leads and lags in regional climate and proxy response times. However, independent integration of palaeoclimate records from different regions is not a trivial task and requires major efforts in both reducing uncertainties in individual chronologies, and in developing robust synchronisation tools. In future, this will attract even more attention with increasing attempts to integrate climate archives over larger regions such as those shown in this issue for the Alps (Heiri et al., 2014), Western Europe (Moreno et al., 2014) and Eastern Europe (Feurdean et al., 2014).

The great diversity of sediment archives and related dating techniques, partly based on fundamentally different concepts (e.g. isotopic/radiometric *versus* layer counting), as well as the rapid technical progress, makes it increasingly difficult for non-specialists to keep track of the various dating approaches and their inherent uncertainties when using them for correlation purposes. This paper aims to provide a brief guide to the most commonly used dating methods for palaeoclimate records within the INTIMATE time frame of 60–8 ka, outlining the fundamental methodological principles, the inherent sources of uncertainty, and discussing the reporting of ages. This guide is intended to provide basic information and references relevant for applying chronologies but does not replace comprehensive reviews for individual dating methods (e.g. Walker, 2005).

Moreover, this paper provides an inventory of age reporting protocols applied by the different contributing dating communities with a particular focus on the reference year, or datum, used. Since this is a controversial issue and recommendations of defining a common datum for all dating methods (cf. Wolff, 2007) have not yet been commonly accepted, an overview on how each dating community deals with this problem is provided below.

2. Overview of dating methods

As mentioned above, we will not present a comprehensive list of all available dating methods, but focus on the most common geological dating methods applied for the INTIMATE time frame of 60–8 ka, which include traditional methods such as radiocarbon dating and layer counting (varves and ice), but also recently emerging and rapidly evolving methods such as luminescence and exposure age dating. The latter play an important role in complementing classical INTIMATE palaeoclimate archives like marine and lake sediments or ice cores with geomorphic features like, for example, moraines and dunes or sand sheets. We divide this paper into isotopic or radiometric dating on discrete samples, followed by incremental dating based on continuous layer counting and the latest age modelling approaches.

2.1. Radiocarbon based chronologies

Radiocarbon (^{14}C) dating, a method that was established by Willard Libby and co-workers (Libby et al., 1949; Arnold and Libby, 1951), has been applied to dating natural archives for more than six decades. It has been employed in climate research from the early days of the method (Olsson, 2009 and references therein). Chronologies of marine and terrestrial records of past climate were first established using counting of beta particles, i.e. the product of the decay of ^{14}C atoms. With the development of accelerator mass spectrometry (AMS) in the 1970s sample size requirements were reduced dramatically allowing for higher resolution studies, which led to an expansion of the field (see summary by Povinec et al., 2009). The first attempts to reconstruct the chronology of the deglaciation (Lateglacial) in the North Atlantic region relied heavily on ^{14}C dating. Prior to the INTIMATE programme, the IGCP-253 North Atlantic Region program utilised ^{14}C dating to correlate the timing of climatic changes that took place during the Lateglacial in the North Atlantic region (Lowe et al., 1994). Twenty years later, various improvements in the radiocarbon method have been realised. Among these is a reduction of sample size from 1 mg of carbon to only tens of micrograms of carbon (Ruff et al., 2010), improved precision of AMS analyses (Synal et al., 2007; Synal and Wacker, 2010), extension of the calibration curve (Reimer et al., 2013) and development of calibration software (Aitchison et al., 1989; Bronk Ramsey, 1995; Buck et al., 1999; Blaauw et al., 2003).

The great value of ^{14}C dating is that the method can be applied to any carbon bearing material – meaning that the samples of direct interest may be analysed. Commonly dated substances include: cellulose-containing materials (wood, seeds, plant remains); charcoal and charred material; carbonates (including speleothems, foraminifera, shells); collagen-containing materials (bone, tooth, antler, and ivory); hair; and, bulk sediment. Given its broad applicability and the extensive development of the technique over the last six decades, ^{14}C remains the most commonly applied scientific dating method available for sample materials up to 50 thousand years old.

2.1.1. ^{14}C dating method

Of the three naturally occurring carbon isotopes, ^{14}C is the only one to be radioactive and has a half-life ($t_{1/2}$) of 5730 ± 40 yrs (Godwin, 1962). Concentration of this cosmogenic isotope (produced in the atmosphere by cosmic rays) is very low (i.e., $^{14}C/^{12}C{\sim}10^{-12}$). Oxidised to CO_2, atoms of ^{14}C enter the global carbon cycle and become incorporated into carbon-bearing material that can later be used for dating (Libby et al., 1949). An accurate ^{14}C age requires that the carbon isolated from the sample is representative of the material at the time of deposition. Various methods of sample preparation and measurement have been developed over the years allowing for more accurate and precise chronologies of natural archives (for overview see Hajdas, 2008 and references therein).

2.1.2. ^{14}C age calculation, corrections, and reporting of results

Measured radiocarbon concentrations, determined either by counting methods (decay) or by AMS (counting ^{14}C atoms), result in conventional radiocarbon ages that have been calculated using the original Libby half-life (5568 years) (Table 1) (Stuiver and Polach, 1977; Mook and van der Plicht, 1999; Reimer et al., 2004). It is important to note that all conventional ^{14}C ages include a fractionation correction (i.e., a $\delta^{13}C$ based correction for mass fractionation of ^{14}C atoms that occurs through natural bio-geochemical processes as well as during sample preparation and measurement). By convention, all data are corrected to −25‰, a representative $\delta^{13}C$ value for wood (Stuiver and Polach, 1977).

Radiocarbon ages are reported with $\pm 1\sigma$ uncertainty (reflecting counting statistics, correction for blanks and standards) in ^{14}C yrs BP (Before Present = A.D. 1950). Laboratory sample IDs, which are given to the samples by the radiocarbon dating laboratory, enable the laboratory to be identified and should always be published alongside the ^{14}C measurements.

2.1.3. Calibration of radiocarbon ages

Natural variability in the concentration of atmospheric ^{14}C caused by changes in production rate and exchange between reservoirs of carbon (atmosphere-ocean), and an underestimated

Table 1
Method and data presentation (notation for activity and concentration follows the cited literature).

		Comments
Time range	0–50,000 a	Post 1950 AD, 'bomb peak' Reimer et al., 2004
Technique	Counting β-decay: ca 1 g AMS 100 μg – 1 mg C AMS: Gas ion source 10–50 μg	 Ruff et al., 2010
Measured values ^{14}C concentration ^{14}C age (conventional, Libby)	$^{14}a = {}^{14}A$ (sample)/^{14}A (reference) $^{14}a_N = F^{14}C$ (normalised for $\delta^{13}C$) $T = -8033 \ln {}^{14}a_N$	Stuiver and Polach, 1977; Mook and van der Plicht, 1999; Reimer et al., 2004
Reporting data	Example ^{14}C age: 1000 ± 15 BP	Lab no. required
Conventional ^{14}C age	A.D. 1950 = 0 BP (Before Present)	
Calibrated ages Calibration Curves	Cal AD/BC, Cal BP IntCal13, Marine13, and SHCal13	 Hogg et al., 2013; Reimer et al., 2013 http://www.radiocarbon.org/IntCal13.htm
Calibration software	CALIB, OxCal, BCal	http://www.radiocarbon.org/Info/index.html#programs
Modern samples	$F^{14}C > 1$ (negative ^{14}C age)	Hua et al., 2013

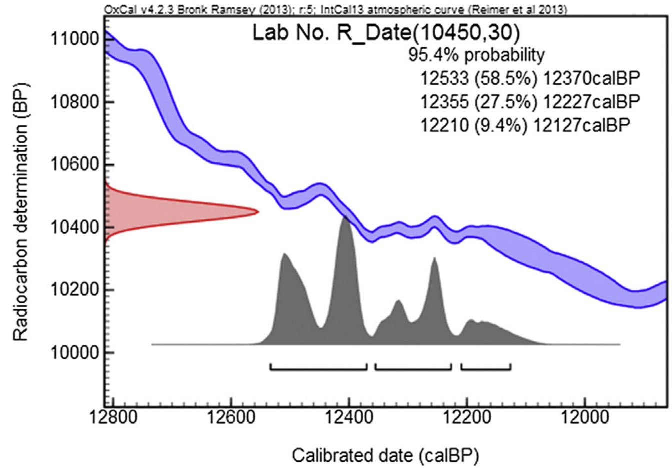

Fig. 1. Example calibration of radiocarbon age 10,450 ± 30 BP (left, in red) obtained using OxCal v4.2.3 and IntCal13 atmospheric curve in blue (Reimer et al., 2013) for sample "Lab No." at 95,4% confidence level. The reported calendar intervals are: 12,533–12,370 cal BP, 12,355–12,227 cal BP and 12,210–12,127 cal BP. Note that this radiocarbon age sits on a ^{14}C age "plateau" or "slow" ^{14}C clock. The time period between 12,600 and 12,100 cal BP corresponds to the decline in the atmospheric ^{14}C concentration (due to changes in the ^{14}C production rate change or changes in the carbon cycle). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

half-life ($t_{1/2}$ = 5568 years) that has been used, by convention, to remain consistent with all data produced in the early days of the method, require calibration of radiocarbon ages. This is achieved using reconstruction of atmospheric ^{14}C concentration based on measurement of ^{14}C concentration in independently dated archives such as: tree-rings (dendrochronology), macrofossils (varve chronology), corals and speleothems (U–Th). The tree-ring based calibration curve, which is a product of many decades of work by various research groups, covers the last 13,900 cal years BP (Reimer et al., 2013). The extension beyond the tree rings is an ongoing process and includes data sets in various stages of development. To maintain clarity and continuity, the radiocarbon community recommends that calibration of radiocarbon ages be done using the most up-to-date international consensus (IntCal) data sets, which are compiled by the IntCal Working Group and require approval by participants of the International Radiocarbon conference. The most recent data set, IntCal13 (Reimer et al., 2013), is now included in all calibration software. A variety of calibration data sets and programs is available online (Table 1). The output of calibration is an interval of possible calendar ages that correspond to the ^{14}C age calculated from the measured ^{14}C concentration. Often multiple intervals correspond to the measured concentration. Fig. 1 illustrates such a case along with the recommended interpretation of the calibrated ages. Bayesian models allow construction of generally more precise calendar chronologies (e.g., Buck et al., 1996 and Section 2.7). For example, age–depth models for sediment sequences can be built that significantly improve the precision of calibrated ages (Hajdas and Michczynski, 2010). Calibration of records from the southern Hemisphere requires use of the SHCal13 curve (Hogg et al., 2013), and marine records require knowledge of the local marine reservoir age (see below, Table 2).

2.1.4. *Reservoir ages*

Radiocarbon ages are calculated with the assumption that the ^{14}C concentration corresponds to the atmospheric level at the time the carbon-bearing matter was formed. Therefore, building an accurate chronology requires careful selection of material along with an estimation of potential depletion in ^{14}C as compared to the contemporary atmosphere at the time of formation. The offset, the so-called reservoir age, depends on the archive. In many cases the reservoir offset can be calculated and utilised to correct the ^{14}C age (Reimer and Reimer, 2001) (see Table 2). The best known example is the marine reservoir age that affects marine samples (molluscs and foraminifera, for example). The marine reservoir effect can be corrected for by subtraction of the ^{14}C reservoir age for the region or by calibration using the MARINE data set and ΔR (i.e., the regional offset from the global reservoir age correction of 405 ^{14}C years (Hughen et al., 2004)). Similar to marine samples, ^{14}C ages of total organic carbon in lake sediments, or ages of speleothems, can be affected by reservoir ages and must be corrected prior to calibration by subtraction of the age offset estimated using the measured ^{14}C concentration of known age samples. However, one must keep in mind that variability of the reservoir ages cannot be excluded and this might result in inaccurate chronologies.

2.1.5. *^{14}C dating and INTIMATE studies*

For many natural archives INTIMATE-relevant chronologies of the last 50 kyrs can be successfully constructed using ^{14}C dating. The latest developments in measurement techniques as well as in calibration issues (IntCal13) and age depth model developments (Section 2.7) have significantly improved the accuracy and precision of ^{14}C based time scales. All ^{14}C relevant information on the

Table 2
Palaeoclimate Archives for ^{14}C dating.

Archive	Material	Reservoir age
Lake sediments	TOC	Possible
	Terrestrial macrofossils	No
	Pollen grains	No
Peat	Fine fraction (>150 um) macrofossils	Possible rootle[t]s
Fossil trees	Wood and charcoal	Possible 'old wood' effect
Marine records	Sediments (foraminifera)	Marine calibration,
	Corals	ΔR http://calib.qub.ac.uk/marine/
Speleothem	Carbonate	Dead Carbon Fraction, varies with location of cave

latest developments (calibration issues, data sets, calibration programs and lab inter-comparison) can be found on the web page of the journal Radiocarbon (www.radiocarbon.org). As with all dating techniques, inter-comparison with other dating methods is necessary. Tables 1 and 2 provide basic information on ^{14}C dating of records.

2.2. ^{230}Th disequilibrium dating: principles and uses

^{230}Th disequilibrium dating is a well-established technique used in palaeoclimate research to obtain independent, accurate and precise chronologies on a wide range of materials including speleothems, corals, travertines, tufas, and less commonly molluscs, marine and lacustrine sediments and peat. The method is based on the radioactive decay series beginning with ^{238}U, encompassing ^{234}U and ^{230}Th, and ending with ^{206}Pb. Where secondary deposits are formed, isotopic fractionation caused by chemical, physical or nuclear processes creates a state of disequilibria (Fig. 2-1). The date of the fractionation event can be determined from the subsequent ingrowth and decay of nuclides (Fig. 2-2) provided that: 1) U and Th remain in a closed system; 2) initial ^{230}Th is either negligible or can be corrected for; 3) there are measurable quantities of U and Th; and 4) the decay coefficients are known exactly. For the ^{238}U–^{234}U–^{230}Th decay series, secular equilibrium is approached after c. 600–700 ka, thus providing the upper limit of the dating method. For further details on ^{230}Th disequilibrium dating, refer to Bourdon et al. (2003).

Over the 60–8 ka focus-period of INTIMATE, the palaeoclimate archives most commonly dated using the ^{230}Th method are speleothems. Speleothems record proxy information (δ^{18}O, δ^{13}C, trace elements, fluid inclusions) (Fig. 2-3) at high resolution, may be deposited over long continuous time scales, have low diagenetic potential (as opposed to e.g. corals), and have a wide (nearly global) distribution. In addition, independent ^{230}Th chronologies derived from speleothems provide robust and reliable calibration points for other dating techniques (e.g ^{14}C; Reimer et al., 2013). Given the abundance and importance of speleothem research to the INTIMATE period, we focus here on the associated sampling, chemical and analytical protocols for the speleothems. However, the chemical and analytical procedures are largely similar and adaptable to other ^{230}Th datable archives.

2.2.1. Sampling protocols

Several approaches may be taken to prevent unnecessary over-collection of speleothems (Frappier, 2008) including low-invasive drilling techniques (Spötl and Mattey, 2012). Prior to sub-sampling for ^{230}Th analysis, petrographic studies are useful in identifying important fabric changes and diagenetic effects. Sub-sampling for ^{230}Th dating depends on a combination of sample type, structure, isotopic concentration, analytical method and required spatial resolution. For stalagmites and flowstones, powders (Fig. 2-3) or slabs are sampled from a specific growth layer using either a precision drill or wire saw respectively. Stalactites are commonly avoided because they have a complicated internal structure, are prone to diagenesis, and are unlikely to survive beyond the Holocene. For stalagmites, the “cleanest” material is often found along the central axis because the impacting water drip flushes particulates towards the flanks (Richards and Dorale, 2003). For U-rich samples, or studies requiring a high spatial resolution, micro-milling or in-situ laser ablation (LA) may be employed (Hoffmann et al., 2009). Using modern multi-collector inductively coupled plasma mass spectrometric (MC-ICPMS) techniques, permil levels of chronological precision are achievable with 20–200 mg of sample from speleothems 5–100 ka in age with ng g^{-1} to low μg g^{-1} concentrations of U (Shen et al., 2012). For

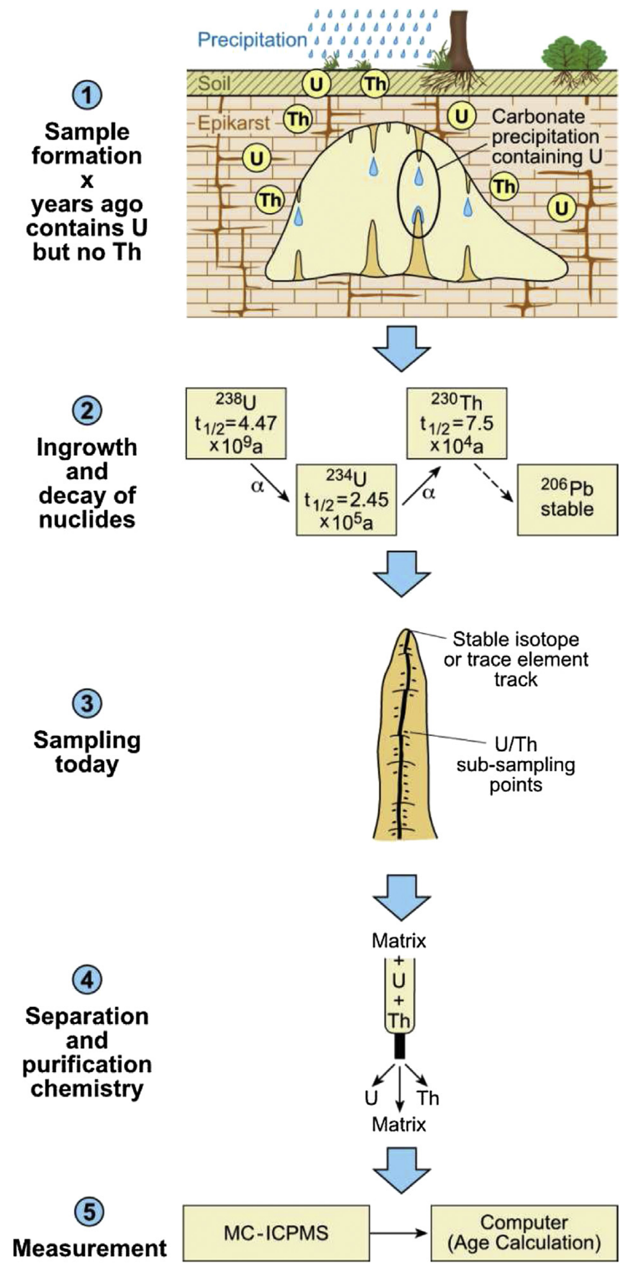

Fig. 2. Schematic diagram showing processes involved in obtaining a ^{230}Th age from a speleothem.

corals, sample sizes of 10–50 mg are sufficient. However, for speleothems with exceptionally high U concentrations, as little as 1 mg can produce precisions better than 1% (Hoffmann et al., 2009).

2.2.2. Chemical and analytical procedures

U and Th aliquots are separated and purified from the carbonate matrix in a clean laboratory prior to analysis in order to reduce the influence of matrix effects during measurement (Fig. 2-4). Separation and purification procedures typically include either: 1) Fe co-precipitation followed by a 1-step anion-exchange column (e.g. Edwards et al., 1987; Shen et al., 2012); 2) a simple 1-step anion-exchange column (e.g. Luo et al., 1997), or rapid 1-step U/TEVA chromatographic resin column (Douville et al., 2010); or 3) a 2-step anion-exchange column (e.g. Hoffmann, 2008).

Technical improvements over the last few decades have seen a shift from thermal ionisation mass spectrometry (TIMS) (e.g. Edwards et al., 1987) to multicollector (MC)-ICPMS for ^{230}Th dating (e.g. Luo et al., 1997; Hellstrom, 2003; Goldstein and Stirling, 2003; chronicled by Cheng et al., 2013 and references therein) (Fig. 2-5). The primary advantages of MC-ICPMS in comparison to TIMS are: 1) smaller sample sizes for a given precision due to the highly efficient ionisation of most elements; 2) solution concentration controls the signal intensity, rather than the mass of loaded U and Th; 3) instrumental mass discrimination can be corrected for, and; 4) routine analysis times are reduced to minutes.

2.2.3. Age calculation, corrections and reporting of results

Calculation of ^{230}Th dates from the measured isotopic ratios is done by iterative solution (Hellstrom, 2003) of the ^{230}Th age equation (Kaufman and Broecker, 1965), and yields a radio-isotopic date relative to the datum of measurement, by convention given $\pm 2\sigma$ uncertainty. The ability to reliably make comparisons with other geochronological data is thus dependent on knowing this datum. Historically, the reporting of ^{230}Th dates has not been standardised. Dates may have been reported as “Before Present (BP)” meaning A.D. 1950 or, alternatively and rather confusingly “before some other” datum (Wolff, 2007), or “before A.D. 2000 (b2k)” as introduced by Rasmussen et al. (2006) for Greenland ice cores. The worst-case scenario is that no datum is defined, thus creating a present maximum but growing uncertainty up to 64 years (A.D. 1950–2013) in the accuracy of the age. Given the high levels of precision (1–2‰) now attainable for ^{230}Th dating under optimum conditions (Cheng et al., 2013), this is significantly more than or equal to the 2σ uncertainty achievable on dates up to 60 ka ago.

When reporting data it is important to provide enough information so that age estimates may be refined as necessary. Over the decades, several revisions of decay constants have occurred (Cheng et al., 2000, 2013; and references therein), each resulting in a shift up to several permil.

Whilst the most desirable samples for ^{230}Th dating have high U concentrations and negligible detrital Th content, it is often the case that these conditions are not met. If the detrital ^{230}Th component is not corrected for, then the calculated ^{230}Th date will be too old. However, as time progresses, the detrital ^{230}Th component decays, thus the uncertainty in the dating accuracy caused by detrital ^{230}Th decreases with increasing age. Identification of detrital ^{230}Th at the time of formation is achieved by monitoring the long-lived and chemically equivalent ^{232}Th. The sensitive nature of today's mass spectrometric methods necessitates a $(^{230}\text{Th}/^{232}\text{Th})_{\text{activity}}$ to be greater than 100–300 (Richards and Dorale, 2003), otherwise the data must be corrected or rejected, though Hellstrom (2006) suggested that this value may be too low. Where significant detrital Th components exist it is important to state the initial $(^{230}\text{Th}/^{232}\text{Th})_{\text{activity}}$ used in the correction and whether it is based on an *a priori* estimate, stratigraphical constraint (e.g. Hellstrom, 2006) or a direct determination using isochron methods (e.g. Richards and Dorale, 2003) so that an assessment of the reliability of the quoted accuracy and precision may be made.

2.3. Luminescence dating

The term ‘luminescence dating’ encompasses a range of chronologic methods that are typically applied to quartz and feldspar minerals. Each of the methods exploits a signal which builds up in mineral grains through exposure to naturally-occurring ionising radiation (principally from uranium, thorium, and potassium), using this signal to assess the time elapsed since the mineral grains were last exposed to sunlight or to heating. In the case of sediment archives, the event being dated is typically the last exposure of the sediments to sunlight, and hence the numerical age directly dates the deposition of the sediment. The time-range covered by the family of luminescence dating methods is very broad, typically extending from single years towards ~200,000 years (e.g. see review by Rhodes, 2011), comfortably spanning the INTIMATE time range. In certain circumstances, particularly where the environmental dose rate is low, the upper age range can extend beyond this to many hundreds of thousands of years (e.g. Huntley et al., 1993). The two minerals primarily used for dating (quartz and feldspar) are very common, so it is rarely problematic to find sufficient material for dating. Due to recent improvements in accuracy and precision, luminescence dating techniques are becoming increasingly valuable geochronologic tools in the study of past environmental and climatic change. Traditionally, when luminescence methods were applied to sediments they were used to date aeolian deposits such as loess and dune sands, because these were thought to have the greatest potential for bleaching (i.e. re-setting) any pre-existing luminescence signal prior to deposition. As the techniques have evolved over the years, sediments with potentially more complex, mixed bleaching histories have been increasingly examined, such as glacial and fluvial sediments, and more recently sediments from marine and lacustrine environments.

2.3.1. Sampling and laboratory procedures

Luminescence dating methods are typically applied to grains of quartz or feldspar, usually either fine-silt (4–11 μm diameter, ‘fine-grains’) or a narrow (~30 μm) range of sand-sized grains between 63 and 300 μm diameter (e.g. 180–210 μm, ‘coarse-grains’). Different luminescence measurement protocols have evolved over the last 35–45 years. Some exploit signals that are stimulated by heat (i.e. ‘Thermoluminescence’ or ‘TL’ signals), and some utilise a signal stimulated by light (‘Optical’ or ‘Optically stimulated luminescence’ or ‘OSL’ signals). Optical dating methods are more commonly used for dating sediments than TL methods because OSL signals are more easily bleached in nature by exposure to sunlight during sediment transport, thereby removing any pre-existing OSL signal more fully prior to deposition and burial than a TL signal. Measurements can be made using either ‘multiple aliquot’ or ‘single aliquot’ methods. In practical terms, many individual aliquots are used to determine an equivalent dose (D_e) regardless of which method is used, but using ‘single aliquot’ methods enables a D_e value to be determined for each individual aliquot measured (rather than requiring data from several aliquots to determine one D_e value), and hence a distribution of D_e values can be assessed prior to calculation of a luminescence age (Equation (1), see below), thereby improving precision and potentially also the accuracy of ages when statistical models are applied to the D_e values. The approach taken for dating will be influenced by several factors, including the anticipated age-range of the samples to be dated, the grain size and mineralogy of the sediments, and the depositional context. For these reasons, it is mutually beneficial for a close dialogue to be maintained between field experts and luminescence specialists from the earliest stages of planning a project and devising a sampling strategy, through the data analysis and interpretation phases, and continuing to publication.

At times the pace of change within luminescence dating has been extremely rapid; for this reason, it can be difficult for the non-specialist to compare and assess the various dating approaches taken in luminescence dating. Since introduction of the sensitivity-corrected ‘Single Aliquot Regenerative dose’ (SAR) OSL protocol (Murray and Wintle, 2000; Wintle and Murray, 2006), the mineral of choice for dating the past ~100–150 ky is quartz (c.f. Roberts, 2008); this method revolutionised luminescence dating in terms of both the accuracy and the precision of the ages generated. Prior to the introduction of the SAR OSL protocol for quartz, feldspars were the minerals most commonly used for dating. Feldspars offer potential advantages over quartz for luminescence dating, because they give a bright luminescence signal (and hence offer improved precision), and feldspars can extend back further in time than quartz before the signal saturates. However, a major source of uncertainty in feldspar dating has been the phenomenon of 'anomalous fading' (Wintle, 1973), which some workers claim occurs in all feldspars (Huntley and Lamothe, 2001). If undetected or uncorrected, anomalous fading of the optical signal from feldspars results in age underestimations. Recent work by Thomsen et al. (2008) identified a much more stable signal than the infra-red stimulated luminescence signal (IRSL) hitherto used for dating, potentially sparking a revolution in feldspar dating similar to that which has occurred for quartz since 2000. This 'post-IR IRSL' signal (Thomsen et al., 2008) demonstrates minimal to negligible rates of anomalous fading, and as such holds great promise for generating reliable ages (e.g. see review by Buylaert et al., 2012) and is therefore arguably the current signal of choice when dating with feldspars.

The upper limit of any given luminescence technique is not defined in years, but is instead a function of the maximum equivalent dose (D_e, measured in grays [Gy]) that the mineral grains can reliably record, and the rate of exposure to ionising radiation in the natural environment ('dose-rate', measured in Gy/ka) [Equation (1)]; e.g. for quartz, the upper dating limit of loess is typically much lower (e.g. 10^4 a) than that of a dune-sand (e.g. 10^5 a) due to differences in the dose-rate. Work by Chapot et al. (2012) is helping to constrain the reliable upper limit of luminescence dating (in Gy), by comparing the response of natural signals in the field with those generated in the laboratory.

$$\text{Luminescence age (ka)} = \text{equivalent dose ('}D_e\text{', in Gy)/dose-rate (Gy/ka)} \quad (1)$$

Details of how to design a luminescence sampling strategy and the practicalities of taking sediment samples in the field for dating are given by Duller (2008).

2.3.2. Uncertainties associated with luminescence ages

Luminescence ages are usually reported using 1σ uncertainties. The uncertainty quoted on any given age typically combines the random and systematic errors, without making a distinction between them; it is much less common to find random and systematic errors expressed separately within luminescence age tables and hence it is often not clear in any given case what the key sources of uncertainty are in any given luminescence age determination. Using single aliquot procedures involving multiple determinations of equivalent dose (D_e), uncertainties are typically ~5–10%, but the size of the uncertainty can vary according to the depositional context of the sample. For example, ages determined for aeolian sediments may have smaller uncertainties than those for fluvial or glacial sediments, due to the smaller range of D_e values for aeolian sediments caused by the enhanced opportunities for bleaching. Water content is another key contributor to the final uncertainty given for an age determination, affecting the dose-rate to the sample. Interaction between field experts and luminescence specialists is required to determine an appropriate value for the water content, reflecting the likely mean water content over the entire depositional history of the sediments being dated (plus uncertainties in that value).

In a review considering the precision and accuracy of quartz OSL, Murray and Olley (2002) suggest that the minimum realistic value for the total (i.e. random and systematic) uncertainties associated with OSL ages is ~5%. Smaller uncertainties are occasionally reported, but using current measurement techniques these are not necessarily realistic and hence could conceivably pose problems when these OSL ages are combined with data from other dating methods. If uncertainties on individual luminescence age determinations are to be reduced further, then one approach would be to reduce the number of parameters measured, e.g. using isochron methods, or auto-regeneration (e.g. Wheeler, 1988). An alternative approach, as for any chronologic method where multiple determinations of age are available, is to incorporate knowledge of the stratigraphic relationships between samples within a Bayesian age–depth model (e.g. Rhodes et al., 2003; Muhs et al., 2013).

2.3.3. Reporting of a datum for luminescence ages

Luminescence dating methods provide a means of assessing the time that has elapsed since the occurrence of the event that is being dated, and the time of sample collection or dating measurements being made; as such, luminescence ages refer to 'years ago' or, simply 'a', as recommended by IUPAC-IUGS Task Group (2006) (cited by Rose, 2007). Unlike radiocarbon dating (see chapter 2.1), there is no standardised method of reporting for luminescence ages, and no agreed datum. Instead, the way in which luminescence ages are reported has been dependent upon the context in which the ages are to be used, and each study conducted will typically use its own (often unspecified) datum. For example, in archaeological studies, luminescence ages are often converted into calendar dates or date ranges (expressed as AD/BC or CE/BCE), whereas in geomorphologic or Quaternary studies luminescence ages are usually presented as 'years ago' or 'a' because referring to a datum 2000 years ago holds no meaning, particularly when considering events which straddle the BC/AD boundary. Of course, the ages relate to the time of sample collection or measurement, and so the datum is always shifting. In the past, publications presenting luminescence ages tended not to formally state a datum; in such cases the datum that should be used is the date of publication of the paper as the nearest approximation to the likely datum. Only more recently have publications tended to report a datum for luminescence ages because of the increasing ability to generate modern-to-recent ages, where the years elapsed since a study was conducted becomes significant when considering the 'age' of the sediments dated. A recent article (Duller, 2011) considered the pros and cons of having an agreed datum for luminescence ages, including the problem inherent with any fixed datum, namely that at some point the ages generated will become negative. The diverse application of the technique makes the selection of one universally appropriate datum difficult. Recent discussions at the International Luminescence and Electron Spin Resonance meeting in Poland (2011) of the ideas presented in Duller (2011), concluded that no single datum should be adopted by the luminescence community at large, but the delegates unanimously agreed that: 1) the choice of datum used in each study should be clearly stated in any table and paper presenting luminescence ages, and 2) the use of 'BP' should be reserved exclusively for radiocarbon ages and should not be used with luminescence ages.

2.3.4. Publication of luminescence ages

To demonstrate the quality of luminescence ages generated and enable assessment at the time of publication or in the future, additional information should be provided to support luminescence ages when they are first published, and a reference to the source of this supporting information should be given when ages are subsequently cited. It is not sufficient to simply publish an age with an uncertainty, without making reference to this supporting information. The additional information required includes details of the material and grain-size used for dating, the measurements made and the methods of analysis, the results of quality control checks, and also summary tables of the key parameters used to calculate an age including D_e values, water content, and dose-rates for each sample (see e.g. Duller, 2008 for further details). To demonstrate that the signal is neither saturated, nor near the detection limit, ideally a representative dose–response curve and signal decay curve or glow curve would also be shown; a D_e distribution plot is also important, particularly when working in complex depositional environments. It is recommended that any work presenting luminescence ages for the first time should be written in consultation with a luminescence expert, who will advise on the interpretation of ages and the appropriate level of information to include in the publication.

2.4. Cosmogenic nuclide dating

It was recognised early on that surface exposure dating with cosmogenic nuclides had the potential to provide an unrivalled tool for the study of past ice margins (Davis and Schaeffer, 1955). Glaciers vary in volume and length in tune with changes in temperature and precipitation. Moraines build up on the margin of a glacier and after it has melted back, record its former extent. If one can directly date moraines one can construct a chronological structure for past glacier volume variations and therefore past climatic fluctuations. Before the advent of surface exposure dating, former ice margins were difficult to date because of the dearth of organic material for radiocarbon dating during the associated cold period. Surface exposure dating allows direct dating of the record left by the glacier itself: the moraine, the glaciofluvial outwash terrace, or the erratic boulder (Ivy-Ochs and Schaller, 2010). Finally and uniquely, with surface exposure dating glacially eroded bedrock surfaces can be dated. Determining the exposure age of the top surface of a boulder on a moraine allows estimation of how long that boulder has been exposed to cosmic rays and consequently when the moraine stabilised. The time point of moraine stabilisation closely reflects the moment when the glacier was no longer actively delivering material to the moraine. Depending on the position of the moraine studied and the moraine complex architecture, the period of occupation and the end of the advance can be determined (Gosse, 2005).

Cosmogenic nuclides are produced in rocks and sediment exposed to cosmic rays at or near the surface of the Earth (Gosse and Phillips, 2001). The radionuclides ^{10}Be, ^{14}C, ^{26}Al, and ^{36}Cl and noble gases ^{3}He, ^{21}Ne, are today the most commonly used cosmogenic nuclides for problems of Quaternary landscape history and geomorphology (Ivy-Ochs and Kober, 2008). Cosmogenic nuclides are produced within minerals due to nuclear reactions induced by impingement of cosmic ray particles. For example, in a quartz crystal, a cosmic ray particle hits the nucleus of a ^{16}O atom, spallation comprising ejection of several particles occurs, and a ^{10}Be atom is left in the lattice site. Because of the different half-lives (or stable in the case of the noble gases) and varied production mechanisms, nearly all rock types can be investigated with one of the available isotopes (Table 3).

2.4.1. Sampling considerations and laboratory procedures

Cosmogenic nuclides are used to date rocks and sediment in two ways; i) surface exposure dating and depth-profile dating, which are based on the build-up of nuclides, and ii) burial dating which is based on the difference in decay (half-life) of two nuclides measured in the same sample (commonly ^{10}Be and ^{26}Al). The latter approach is useful for sediments that were deposited more than a few hundred thousand years ago and will not be discussed here further (see Dunai, 2010). The fact that the rate of build-up in surface materials decreases rapidly and regularly with depth, makes depth-profile dating useful for dating sedimentary units that were deposited rapidly followed by top surface abandonment (Ivy-Ochs et al., 2013). Both surface exposure dating, where the upper few cm of a rock surface are analysed, and depth-profile dating, where several samples down to a few metres depth are taken, allow dating of landforms ranging in age from a few hundred years (Akçar et al., 2012) up to hundreds of thousands of years in temperate climates. In regions where rates of landscape change are slow (Antarctica, Australia) ages up to many millions of years have been reported (Ivy-Ochs and Kober, 2008, and references therein).

For the dating of palaeo-ice-margins surface exposure dating, most commonly with ^{10}Be measured in the upper surface of boulders on moraines or erratic boulders, is applied. The largest (>1.5 m high), broadest, and flattest boulders located on the crest of the moraine are sampled. Large boulders have a greater probability of having remained stable and of not having been covered by sediment or snow during exposure. The sampled rock surface must have undergone single-stage (no pre-exposure), continuous (not covered) exposure in the same position (not shifted), and not have spalled. Several hundred grams of the upper few centimetres of the

Table 3
An overview of cosmogenic nuclides applied in surface exposure dating.

^{14}Nuclide	Half-life	Other isotopes	Meas. method	Target elements	Production rate atoms/g yr[a]	Advantages/minerals used	Disadvantages
^{10}Be	1.4 My	^{9}Be	AMS	O (Si)	4.5	quartz resistant and ubiquitous	Low production rate generally restricted to quartz (no meteoric ^{10}Be)
^{26}Al	720 ky	^{27}Al	AMS	Si	30	High production rate quartz resistant and ubiquitous	Restricted to quartz (low Al) accurate determination of ^{27}Al required
^{36}Cl	300 ky	^{35}Cl, ^{37}Cl	AMS	Ca	Composition dep.	Any rock type, silicates & carbonates	Complicated production
				K			^{36}S interference in AMS
				^{35}Cl	e.g. 10 granite e.g. 20 limestone		Accurate determination of total Cl required Determination of rock composition required
^{14}C	5.73 ky	^{12}C, ^{13}C	AMS	O	16	Useful for short time scales quartz resistant and ubiquitous	Short half-life atmospheric ^{14}C contamination
^{3}He	Stable	^{4}He	Static mass spec.	Many	120	High production rate useful for long time scales pyroxene, olivine	Diffuses out of quartz or volcanic groundmass radiogenic/nucleogenic/magmatic correction beware pre-exposure
^{21}Ne	Stable	^{20}Ne, ^{22}Ne	Static mass spec.	Mg Si	20	Useful for long time scales, >50 ka quartz, olivine, pyroxene	Nucleogenic/magmatic correction high air background possible beware pre-exposure

[a] Production rates (Gosse and Phillips, 2001; Balco et al., 2008).

rock surface are chipped off with a hammer and chisel. Sample size depends on location, exposure age, and mineralogy of the sampled surface. Pure quartz is separated from the rock sample, Be is extracted from the rock and $^{10}Be/^{9}Be$ ratios are measured with AMS. Data are given in ^{10}Be atoms per gram of quartz. Data required for exposure age calculation include, in addition to the ^{10}Be concentration, the latitude, longitude and elevation of the site, the thickness of the sample, and the factor for correcting for surrounding topographic shielding (Dunai and Stuart, 2009). Exposure ages are calculated using the CRONUS-Earth online calculator (http://hess.ess.washington.edu/) set-up by Balco et al. (2008). With this calculation portal all workers (can) now all use the same method for calculating exposure ages. This has led to increased transparency and straight-forward comparison of data sets from different groups. With updates to production rates, data from earlier papers can be recalculated with the most recent values.

2.4.2. Uncertainties

Uncertainties from the AMS measurement can be as low as 3%; total systematic uncertainties of an exposure age are on the order of 7–10% (Gosse and Phillips, 2001). Recent production rate determinations from independently dated sites across the globe indicate remarkable convergence around a sea level high latitude value of about 4 ^{10}Be atoms per gram of quartz per year, with an uncertainty of 5% or less (Young et al., 2013; Heyman, 2014). This leads to an overall reduction of uncertainties and in turn increases comparability between widely separated sites. Nevertheless, Southern Hemisphere sites and/or those at lower latitudes, where past magnetic field effects would have notable influence, may warrant use of a locally determined production rate (Putnam et al., 2010). Several uncertainties must be recognised related to rock surface history and obtained age results. Although inheritance (a non-zero initial concentration of the determined cosmogenic nuclide in the rock surface, also termed pre-exposure) that would result in age overestimates is theoretically possible, very few cases have been identified in moraine boulders (Heyman et al., 2011). In contrast, post-depositional processes that lead to ‘erroneously young’ exposure ages are more common (Porter & Swanson, 2008). Ages younger than the true age of deposition result as boulders spall (whereby the uppermost few cm with the highest nuclide concentrations are lost), topple or are exhumed. Because of these effects not only must sampling be undertaken cautiously, but suites of obtained exposure ages must be scrutinised judiciously and in light of morphostratigraphic relationships and independent age information. It is unwise to simply take the mean of all determined boulder exposure ages in all cases (Ivy-Ochs et al., 2007). When exposure ages from one landform cluster tightly then a mean age is determined. Arithmetic means are preferred rather than error-weighted means as the analytical uncertainty (dominated by AMS) is a poor measure of reliability of the date (boulder). In any case, all determined ages from a site are always reported in a publication.

2.4.3. Age reporting

Ages are reported in years, representing the period of time between boulder deposition and the year of sampling. As uncertainties are in the range of several percent or more, inaccuracies from omitting age datum reporting are negligible. The ‘years’ of surface exposure ages are roughly equivalent to calibrated radiocarbon years or calendar years.

In the time range of interest for INTIMATE 60–8 ka ago, surface exposure dating with cosmogenic nuclides can be widely applied, with large data sets being produced on the dating of Last Glacial Maximum, Lateglacial and Holocene ice margins (Ivy-Ochs et al., 2008). As the aim of INTIMATE is to link ice core, marine and terrestrial records of past climate changes, including the palaeoclimatic information reflected by former ice-marginal positions is crucial.

2.5. Ice layer dating/stratigraphic ice core chronologies

Ice cores are most often dated by annual layer counting, ice flow modelling, stratigraphic linking, or by a combination thereof. Seasonal variations in isotopic composition and impurity content of the snowfall provide the basis for a distinct annual signal in the snowpack, which may be preserved in ice under favourable conditions. The capability of an ice core to provide (sub-)annual information depends on the accumulation rate, which ranges from a few centimetres of ice per year in high elevation areas of Antarctica to several metres at coastal sites of Greenland and in lower latitudes. In the upper part of an ice sheet, snow slowly compacts to incompressible ice. Due to the continuous accumulation of snow, annual layers become buried in the ice sheet over time and gravity causes the ice to flow toward ice margins (Fig. 3D) which, in turn, results in thinning of annual layers with depth (Fig. 3A–C). Therefore, even in cores from high-accumulation areas, annual layers may be identifiable only in the upper parts of the ice core. Only ice cores from Greenland and Antarctica have sufficiently high temporal resolution to allow for annual layer counting within the 60–8 ka INTIMATE time frame.

Glacier ice contains air bubbles representing past atmospheric composition. Air moves freely in the upper layers of an ice sheet (known as the firn), and the transformation of snow to glacial ice (known as firn densification or firnification) isolates this air and forms bubbles at a depth of 50–100 m (the ‘lock-in’ depth). Hence, the age of bubbles is younger than that of surrounding ice by an amount known as Δage. The magnitude of Δage is estimated by densification models constrained by data at abrupt warmings where abrupt changes can be identified in both gas and ice measurements. It varies from a few decades at high-accumulation sites to about 5 millennia on the East Antarctic Plateau during the LGM. Because of this, ice core time scales generally include parallel time scales for the ice and gas records.

2.5.1. Dating by annual layer counting

The ratios of stable oxygen and hydrogen isotopes in glacier ice (expressed as $\delta^{18}O$ and δD values) reflect seasonal variations in local temperature at the time of precipitation (Dansgaard, 1964). In ice cores from high-accumulation sites, this annual signal may be traced thousands of years back in time. However, flow-induced layer thinning and diffusion of water molecules ultimately obliterates the seasonal $\delta^{18}O$ cycle (Johnsen et al., 2000). Therefore, in deeper ice, chemical impurities and dust particles that are less prone to diffusion are employed for annual layer detection. Many of those ice core impurities display a distinct seasonal variation. For example, Greenland dust concentrations generally reach maximum during springtime, whereas the amount of sea-salt aerosols peaks during winter (Rasmussen et al., 2006). In general, impurity records are more difficult to interpret than isotope records as they also contain non-annual signatures, such as input from volcanic eruptions, biomass burning, and other episodic sources. For this reason, the parallel analysis of several impurity records of different origin is recommended when establishing a robust chronology by so-called “multi-proxy dating”. Even then, however, the records contain “uncertain annual layers”, i.e. features that are difficult to interpret and may or may not represent an annual layer. As with other annually laminated records, uncertainties in the layer count accumulate along the core, and the absolute uncertainty in deeper ice can be considerable.

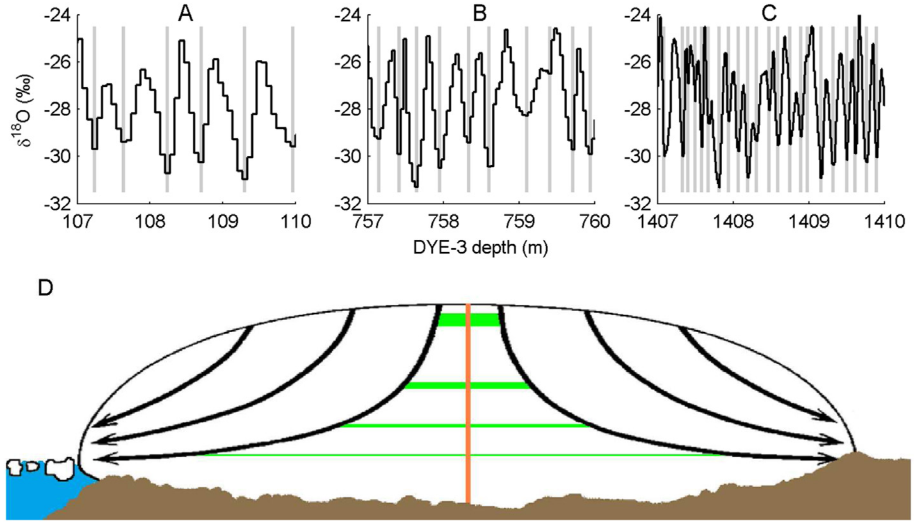

Fig. 3. Annual layers in ice cores. A, B, and C: gradual thinning of annual layers (grey vertical lines) as expressed in water isotopes ($\delta^{18}O$) at three different depths of the DYE-3 ice core from southern Greenland. D: schematic vertical cross section of the Greenland ice sheet with indication of the main ice flow pattern (thick black arrows), thinning of annual layers with depth (horizontal green lines), and an ice core borehole (vertical orange line). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Nevertheless, recent periods of abrupt climate change can be dated with an outstanding precision, and periods of past abrupt climate change can be mapped out with a relative dating precision of only a few years (Steffensen et al., 2008).

2.5.2. Key chronologies

Over the last decades a number of different ice core chronologies have become available from Greenland and Antarctica. Traditionally, each new ice core has been accompanied by a new time scale and, while some time scales are not coherent (Southon, 2004), many now involve the use of marker horizons that enable consistent chronologies to be established across several cores (Parrenin et al., 2012; Seierstad et al., 2014).

The Greenland Ice Core Chronology 2005 (GICC05) is the most recent Greenland ice core chronology that is applied to the major Greenland ice cores (GRIP, GISP2, NGRIP, and NEEM) and forms the basis of the current INTIMATE event stratigraphy (Rasmussen et al., 2014). It is entirely based on layer counting back to 60 ka (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2008), the uncertainty of which is discussed below. In the Last Glacial period the time scale has an estimated uncertainty of around 5% based on the Maximum Counting Error (MCE) approach (see below). Within the 60–8 ka time interval, GICC05 deviates significantly from several of the earlier Greenland ice core chronologies (Svensson et al., 2008).

The Antarctic Ice Core Chronology 2012 (AICC2012) provides the most recent attempt at integrating ice core chronologies from several sites using all available dating information (Bazin et al., 2013; Veres et al., 2013). AICC2012 is applied to several Antarctic ice cores (EDC, EDML, Vostok, and TALDICE) where it is mainly based on ice-flow modelling. Within the 60–8 ka time frame AICC2012 is constrained by and consistent with the Greenland GICC05 time scale. The Greenland and Antarctic ice cores are synchronised using global marker horizons such as bipolar volcanic markers (Veres et al., 2013), common variations in methane concentrations, and geomagnetic events. The recently retrieved ice core from the relatively high accumulation site on the West Antarctic Ice Sheet Divide (WAIS Divide) is dated by annual layer counting and provides an opportunity for the construction of multi-proxy chronologies back into the glacial period (WAIS Divide Project Members, 2013).

2.5.3. Uncertainties

It is not currently clear how the gradually increasing uncertainty of layer counted ice core chronologies can be properly quantified and different ways to estimate the uncertainty have been reported. For example, counting differences between different investigators and the use of different data series have been applied to provide an age uncertainty. Rasmussen et al. (2006) discussed how to combine counting uncertainties under different assumptions and concluded that the dominating error sources are neither uncorrelated nor fully correlated. Based on this observation, the adoption of the Maximum Counting Error (MCE) concept was suggested, in which the uncertain annual layers are counted as ½ ± ½ year and summed up linearly as if they were correlated, while in return ignoring other and smaller sources of uncertainty.

The MCE is intended to represent all uncertainty contributions except possible bias in the annual layer identification process that, by the nature of the problem, cannot be estimated without the existence of independent data. Comparison to independently dated reference markers suggests, however, that the MCE is a highly conservative uncertainty estimate and that any possible bias is much smaller than the MCE (Svensson et al., 2008). The MCE is not a Gaussian uncertainty measure and MCE values cannot, strictly speaking, be compared to Gaussian-style dating uncertainties of other records. However, in the absence of a more appropriate uncertainty estimate, when comparing with other records where the error is based on a Gaussian probability distribution, we recommend that the MCE is regarded as the equivalent of a 2σ uncertainty (Andersen et al., 2006). As such, the MCE is the standard uncertainty measure of the GICC05 time scale on which the recent INTIMATE event stratigraphy is based.

2.5.4. Reporting datum

Concerning the use of a datum, ice core ages have traditionally not been reported in a consistent way. The BP notation was for a period routinely used for reporting age estimates with “present” being A.D. 1950, but it has also in several instances been used with “present” referring to the upper layer in the ice core, i.e. “present” being the start year of the drilling of that particular ice core (Wolff, 2007). To avoid further confusion, and to underline the conceptual difference from radiocarbon-based dating, the notation 'b2k' was introduced as shorthand for "calendar years before the year A.D. 2000" (Rasmussen et al., 2006), being both short and unambiguous. The b2k notation is used when reporting ice core ages as part of the GICC05 dating framework, but has not gained much momentum outside of this context. In general, it is recommended that ice core ages are reported relative to a clearly specified datum, i.e. as "b2k" (years before A.D. 2000) or "years before 1950", whereas the use of BP (regardless of the choice of "present") is not encouraged for ice core ages.

2.6. *Varve chronology*

A varve chronology provided the first 'absolute dating' for the timing of a climatic change in the geological past. The Swedish geologist Gerard De Geer had recognised that the laminated sediments he studied in proglacial palaeolake deposits reflect the annual cycle of spring snow melt. He used the term 'varve' (shortened from the Swedish 'varvig lera' meaning layered clay) for these annual laminations and counted them by naked eye in a transect of palaeolake deposits. From these varve counts he developed a 12,000 year-long master chronology of the ice retreat in southern Sweden at the end of the last glaciation (De Geer, 1912). This was the birth of varve chronology.

Today, varves are reported from a large variety of mainly lacustrine environments (Ojala et al., 2012) and rarely also from marine sediments. The formation of varves is driven by prevailing seasonal climates causing cyclic variations in sediment deposition. Depending on the climatic zone and local catchment geology, various varve types form, of which the most common are minerogenic, biogenic and evaporitic facies (Brauer, 2004; Zolitschka, 2003). This large variety of facies is a specific characteristic of varves. However, in most lakes varves are not preserved due to bioturbation, wind-induced mixing and degassing of the sediments. Only in few lakes with particular morphometric and limnological characteristics are varves preserved, but the number of known varved lake sediments is increasing, especially after systematic prospecting for such sediments (Ojala et al., 2000).

Since fine laminations are not necessarily of annual origin (e.g. Lambert and Hsü, 1979) the proof of true varves is a main prerequisite for establishing varve chronologies. Such a proof can be achieved by different approaches including observation of recent deposition in sediment traps or by taking sediment surface cores in consecutive years to depict the annual sediment increments. An annual origin of fine laminations can further be proven by detailed micro-stratigraphic varve models. Based on microscopic identification of typical seasonal components, such as specific diatom species or mineral precipitates, the annual cycle of sedimentation can be confirmed (Kelts and Hsü, 1978; Brauer et al., 1999a). Microfacies analyses further allow the distinction between simple light-dark couplets (e.g. Lotter and Lemcke, 1999) and more complex systems with up to six seasonal sub-layers (e.g. Neugebauer et al., 2012).

The vast majority of varve chronologies cover Holocene and Lateglacial time intervals because they have been established from lakes formed after the last glaciation. Only in a few cases independent varve chronologies reach back into the Last Glacial Maximum (e.g. Zolitschka et al., 2000) or even cover the full INTIMATE time scale, such as the Suigetsu (Schlolaut et al., 2012) and Monticchio (Brauer et al., 2007) chronologies. In the case of the Suigetsu chronology, varve dating has even been used for calibrating the radiocarbon time scale (Bronk Ramsey et al., 2012). Moreover, there are a few examples of varve chronologies beyond the INTIMATE time scale covering either entire or parts of previous interglacials (Müller, 1974; Brauer et al., 2007; Mangili et al., 2007).

Varve chronologies commonly cover time intervals from a few hundred years to several millennia depending on the stability of lake conditions favourable for varve preservation. An important aspect for varve chronologies is if seasonal layer formation continues up to the present time. Only in such cases are independent 'absolute' chronologies established, starting from the time of core retrieval. All other varve chronologies are considered as 'floating' and require anchoring to absolute chronologies by other dating methods, for example, by ^{137}Cs and ^{210}Pb dating, tephrochronology (Brauer et al., 1999a), or linking high-resolution ^{14}C ages to the radiocarbon calibration curve (Staff et al., 2013).

2.6.1. *Methods*

An obvious pre-requisite for reliable varve chronologies is high-quality and undisturbed sediment cores obtained with high-precision coring devices (Mingram et al., 2007). Varve counting is traditionally based on optical methods utilising the colour contrasts between seasonal layers to recognise individual varves. A broad spectrum of methodological approaches has been applied with different technical and time demands on core and sample preparation and analyses. A standardised protocol for varve analyses is not meaningful because of the broad variety of sediment and varve types, each requiring specifically adapted methods. In general, it can be stated that the more complex a varve facies is, and the lower the colour contrasts are, the greater the efforts for varve analyses and counting must be. Three main methods of varve counting are applied: (1) manual counting, either directly on cores or using various types of images, (2) manual counting on petrographic thin sections, (3) automated image analysis based counting either directly on cores or on thin section images. (1) The fastest way is manual varve counting of the cleaned surface of freshly split sediment cores either by naked eye or using low-magnification microscopy. (2) Higher operating expenses are required for varve counting on overlapping petrographic thin-sections of epoxy-impregnated sediment blocks using a combination of various high and low-magnification techniques and light conditions (cross-polarised, plain parallel, incident, dark/bright field). The main advantage of this method is the ability to precisely define varve boundaries based on micro-facies data. In particular, for complex varve facies with more than two sub-layers and low colour contrasts, thin section analysis is essential. Details of different techniques of epoxy impregnation of wet sediments are given by Lotter and Lemcke (1999) and Lamoureux (2001). (3) More recently, automated image analysis techniques have been applied on digital core photographs, radiographs or thin section images at various resolutions in order to reduce the time for analyses and provide a more objective measure of counting (Francus et al., 2002). As for manual varve counting, image analysis usually makes use of the colour differences between the seasonal sub-layers. Comparison of the different counting methods has demonstrated that counting of fresh cores and image analyses tend to underestimate the number of varves compared to thin section counts (Lotter and Lemcke, 1999; Hajdas and Michczynski, 2010), because the varve boundaries are not as precisely identified as with a microscope. Recently, high-resolution μ-XRF scanning has been applied to use geochemical signatures of assumed seasonal nature for varve counting (Marshall et al., 2012), but there are still too few case studies to reliably assess the potential of this technique for a wider use.

2.6.2. *Uncertainties*

Uncertainties in varve chronologies primarily depend on the clarity of varves, which itself is determined by: (i) the distinctiveness of seasonal sub-layer boundaries (micro-facies, colour); and (ii) sediment disturbance either due to post-depositional sedimentary processes (bioturbation, wind-induced mixing, sediment-degassing) or due to coring or preparation artifacts. Since varve clarity usually varies through sediment records, the uncertainties may also vary significantly between different intervals of a chronology (Brauer et al., 1999b).

In addition, hardly detectable uncertainties exist, for example, hiatuses involving either single varves (Fig. 4) or tens to even hundreds of varves. With single core analyses such gaps often remain undetected. Locally (at one specific core location), missing single varves are not unusual because sub-millimetre scale laminations are not always equally deposited over the entire lake bottom. Larger hiatuses are less common and can be related to erosion events which may even occur in the deepest part of a lake basin. Therefore, varve chronologies established from single cores likely underestimate the true varve number. Overestimation of varve numbers is less common but may also occur for some varve types, such as clastic varves, by misinterpreting intercalated event layers as seasonal sub-layers. Even if detailed micro-facies analyses help to reduce the misinterpretation of event layers, this source of error cannot be fully excluded. Therefore, most varve chronologies have been cross-checked with other independent dating methods (Ojala et al., 2012).

The uncertainties discussed above can be significantly reduced by multiple counting, preferably by different analysts and applying micro-stratigraphic techniques (Swierczynski et al., 2013), especially for thin varves or those with complex sub-annual structures or low colour contrasts. For reducing errors due to missing varves, multiple core analyses based on detailed core comparison with well-defined micro and macro marker layers (Fig. 4; Neugebauer et al., 2012) is essential. Obviously, this approach is cost-intensive and time-consuming and might not be applicable for each study. However, it should be applied for key palaeoclimate records.

2.6.3. Reporting varve chronologies

Accepted protocols for reporting the methodological approach and uncertainties currently are still lacking for varve chronologies. As for all incremental chronologies that are not fully automated, a well-defined analytical (machine) error cannot be given. Error ranges in varve chronologies are thus usually given as mean values from repeated counts or the difference between maximum and minimum counts (e.g. Swierczynski et al., 2013). If a quality index for each counted varve is determined, variable uncertainties within a chronology can be pointed out (Brauer et al., 1999b).

Currently no common agreement on age reporting has been achieved. The use of 'varve years' is recommended rather than 'calendar years' which can cause confusion with calibrated radiocarbon dates, especially when the misleading abbreviation 'cal years' is used. The most common datum notation is BP (Before Present = 1950), to enable correlation with chronologies from non-varved lake records that usually are based on radiocarbon dating with BP 1950 as datum. Varve chronologies comprising only historical times scales or which are used in archaeological contexts are given in AD/BC notation.

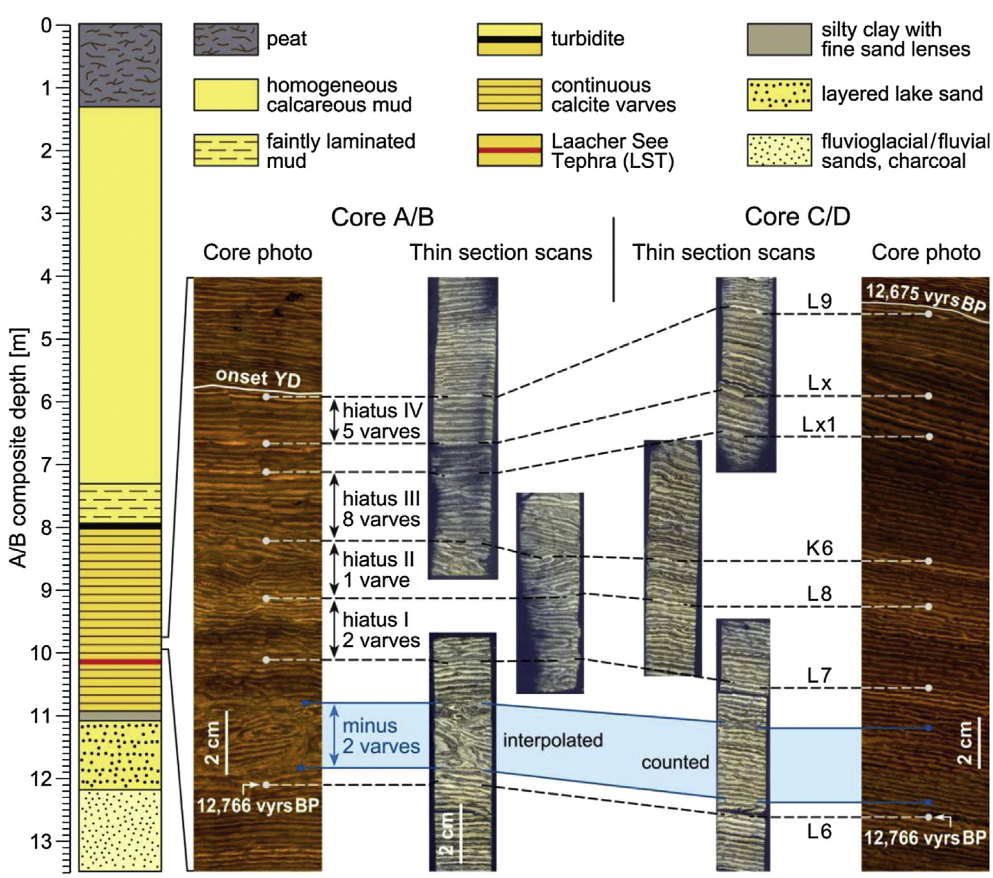

Fig. 4. Example from the Lateglacial varve chronology from Lake Rehwiese, Germany (after Neugebauer et al., 2012), demonstrating how even small hiatuses of 1–8 varves can be identified by detailed comparison of two parallel sediment cores based on well-defined micro-marker layers (labelled L9 – L6) using microscopic analyses on petrographic thin sections. In addition, the quality of interpolation of small disturbed intervals that can locally occur in single cores (see lower part of core on the left) can be tested if the same section is well-varved in a parallel core (to the right).

2.7. Age–depth modelling

Age–depth modelling is central to the aims of the INTIMATE initiative, since palaeoenvironmental data from ice-core, marine, or terrestrial records are generally provided as a time series of proxy records of past environmental change that require reliable placement on to an established time scale. Data may be from ice cores, (marine, lacustrine or terrestrial) sediment cores, palustrine cores, tree-ring sequences, coral sequences, speleothems etc., and in each of these cases, the reconstruction of past time is a function of depth.

Statistical approaches to chronological modelling have expanded dramatically over the last two decades, with the implementation of Bayesian data analysis facilitated by advances in computer processing power that enable the multiplicity of calculations required in such methods, which would not have been possible before. The development of such freely available Bayesian statistical computing packages as 'OxCal' (Bronk Ramsey, 1995, 2008, 2009a), 'BCal' (Buck et al., 1999), 'Datelab' (Jones and Nicholls, 2002), 'BPeat' (Blaauw et al., 2003; Blaauw and Christen, 2005) and 'Bacon' (Blauuw and Christen, 2011) has greatly advanced the science. These programs implement such random sampling procedures as Markov Chain Monte Carlo (MCMC) analysis, which generate accurate approximations of the required probability densities through numerical iterative means where analytical solutions are intractable.

On-going development of such statistical tools is improving both the accuracy and precision of chronological questions that can be answered. However, it is important to note that such modelling approaches cannot be used as a replacement for robust sampling procedures, with careful scrutiny and quality control of samples performed prior to any statistical analysis (Bronk Ramsey, 1998; Bayliss, 2009).

Age–depth modelling is an ideal application for Bayesian methodologies, since the raw determinations produced are (relatively) imprecise, are often numerically fewer than would be desirable (due both to paucity of suitable material for dating and sometimes costs), and may present complex statistical distributions (as is the case with calibrated ^{14}C dates). Yet these individual measurements can be combined with a mathematical model of the processes of sediment deposition, speleothem growth, etc. to produce (generally) more precise, more statistically robust chronological information through the palaeoenvironmental sequence.

2.7.1. Model output

It is of critical importance to be aware that the output of any data modelling (be it Bayesian, or otherwise) is contingent upon both the quality of the raw data entered into the model, as well as the construction of the model itself. A valid choice of model prior is crucial. Furthermore, the subsequent interpretation of "what the results of the Bayesian analysis actually mean" must be understood in light of this model construction (Bronk Ramsey, 2000). A simple example of this (in relation to Bayesian analyses of ^{14}C data) is that the calibrated age of an individual radiocarbon determination might vary according to the choice of calibration data set applied. It is therefore essential to state which calibration data set has been used in determining all calibrated radiocarbon dates (Reimer et al., 2009), and, similarly, the model coding applied to determine more complex model outcomes should also be stated explicitly (published alongside the conclusions drawn from the exercise; Millard, 2014).

Although, as noted above, the choice of model prior is fundamental to the interpretation of the modelling output (the 'posterior' probability distributions), there is "no one correct prior for a given situation" (Bronk Ramsey, 2000), which might necessitate some subjective choices on the part of the modeller. In practice, the alteration of certain prior parameters may make little difference to the modelling outcomes. It is therefore useful to run multiple models, with differing prior information, to determine the sensitivity of the modelling outcomes to particular prior parameters (Bronk Ramsey, 2000).

The output of Bayesian modelling is generally quoted as a range of values at a given probability (e.g. at 95.4%), obtained by numerical integration (highest probability density, HPD) of the probability density function (PDF) histograms. This range describes the values that include the 95.4% most likely results, based upon the model prior applied. This does not imply that any result falling outside of this stated range has a 95.4% probability of being false (in contrast to classical statistical methods). Usually, data are presented at the 68.2%, 95.4%, or 99.7% ranges, providing comparability with the one-, two-, or three standard deviation (σ) ranges provided for Normally-distributed data. However, since the PDFs produced may well not be Normally-distributed, as is the case with calibrated ^{14}C data, such ranges should not be quoted as '1σ', '2σ' or '3σ'.

2.7.2. Outlier analysis

Despite the precautions taken throughout sampling and laboratory processes, a proportion of chronological determinations might be inconsistent within a model (i.e. with respect to the other measurements, within the framework of the applied model prior). These may be the result of: inaccurate laboratory measurement (beyond the laboratory's quoted uncertainty, perhaps the result of an inability to fully remove environmental contamination from samples); environmental effects (e.g. ^{14}C 'reservoir effects'); residuality (samples bearing an 'inbuilt age'); or intrusivity (e.g. samples being moved up or down a sediment sequence).

Although it might be readily obvious to the user which samples are outliers, the application of automated outlier analysis within the available statistical packages renders this process more objective (e.g., Bronk Ramsey, 2009b; Bronk Ramsey et al., 2010; Blauuw and Christen, 2011). If at all possible, all data points should be included within an age–depth model and the software used to identify such outliers, so as to reduce the risk of the user unintentionally biasing the model output produced.

It should be re-iterated, however, that statistical methods cannot identify the reasons why samples might be outlying. Outlier analysis is no replacement for a sound sampling strategy, and cannot account for data sets with a high proportion of erroneous data.

3. Integrating records

As mentioned before, it is of utmost importance to compare the proxy signals between different records from different sites on a robust chronological basis. Besides improvements in precision and accuracy of individual chronologies, additional techniques for synchronising individual archives *by age equivalence* are essential. One example which, however, is limited to the correlation of ice cores is the use of gases like methane in air bubbles in the ice (Blunier and Brook, 2001). Of wider application is tephrochronology, which enables linkages to be established between terrestrial, marine and ice-core records. Recent advances in linking records using tephra, cosmogenic isotopes and palaeomagnetic data are described in the following sections.

3.1. Tephrochronology

Tephrochronology has become a key tool for the INTIMATE community over the last decade. The technique relies on the isolation and identification of volcanic ash layers in various stratigraphic contexts, including ice cores, marine and lacustrine archives and in some cases loess profiles (e.g. Davies et al., 2002; Wulf et al., 2004; Mortensen et al., 2005; Wastegård et al., 2006; Blockley et al., 2007; Bourne et al., 2010; Lane et al., 2011). Tephra layers are preserved in sediment sequences either as visible ash horizons or as microscopic layers not visible to the naked eye (cryptotephra). The identification and extraction of cryptotephra from sediment archives has now become a well-established approach and has significantly extended both the number of tephra layers available as chronostratigraphic markers and the range at which distal ash layers can be detected from their volcanic source. For example many important records contain a number of ash layers that are only located as cryptotephra deposits (e.g., Lane et al., 2013; Wulf et al., 2013). Tephrochronology is particularly useful for the INTIMATE community as many tephras can provide precise ages through direct and indirect dating (below) of ash horizons. In addition, where discrete tephra can be located in multiple archives they can act as a precise correlation tool and can help significantly in understanding leads and lags reflected in climatic and environmental proxies between sequences. Understanding these leads and lags is a core goal of INTIMATE and thus tephrochronology has become a key tool. Particular advances have been made by the INTIMATE community in the use of cryptotephra deposits to directly correlate between ice, marine and terrestrial deposits (e.g. Davies et al., 2010; Lane et al., 2013), and tephra layers are now a key element in the INTIMATE event stratigraphy (Blockley et al., 2012). However the use of tephra as a dating and correlation tool is not without difficulties, which has implications for both the interpretation and reporting of tephra horizons. There are a number of detailed reviews and protocol papers that relate to tephrochronology, and which form a useful guide to some of the issues associated with the technique (e.g. Lowe, 2011).

3.1.1. Dating tephra layers

For the INTIMATE initiative, volcanic centres in Iceland, Germany, France, Italy and the Eastern Mediterranean are the key sources of distal tephra. These ashes are suitable for dating by a range of techniques. For potassium-rich magmas, such as the Italian volcanic centres, proximal units contain K-rich feldspars (sanidines) and these have proved very useful for direct dating by $^{40}Ar/^{39}Ar$. In particular key widespread tephra from the Italian Campi Flegrei volcanoes have been dated by this technique with high precision. Most notably the very widespread Campanian Ignimbrite has a precise $^{40}Ar/^{39}Ar$ age reported of 39,282 + 110 ka (de Vivo et al., 2001), based on the dating and integration of a number of outcrops. However, there are a number of difficulties with the $^{40}Ar/^{39}Ar$ technique, not least the problem of xenocryst inclusion which can lead to ages being too old. For example the highest precision $^{40}Ar/^{39}Ar$ age for the Italian Neapolitan Yellow Tuff tephra is significantly older than the age of this ash derived from a number of radiocarbon ages for organic material associated with the eruption deposits (Blockley et al., 2008). A further issue for the technique is that for young tephra with ages below ~100 ka only K-rich melts contain minerals suitable for precise $^{40}Ar/^{39}Ar$ dating, precluding many of the key tephra used within the INTIMATE time frame from magmas with lower K contents. This includes key tephra from Iceland, which is a significant tephra source for the INTIMATE initiative.

Because of these and other problems, most tephras, especially those within the INTIMATE time frame, tend to be dated by association rather than directly.

For example, the widespread Vedde Ash is precisely dated in the Greenland ice core records and the important Laacher See tephra is equally well dated in the Meerfelder Maar varve record (Brauer et al., 1999a). In many cases, however, tephra in the INTIMATE time frame are dated by radiocarbon ages of associated organic material. This is either charcoal found in proximal deposits (Blockley et al., 2008) or radiocarbon dates within lacustrine or marine records where tephra are located. This means that many tephra ages are subject to the same issues of age modelling and age uncertainties associated with other radiocarbon-dated units. One mitigating factor, however, is the potential for the integration of multiple chronological records using Bayesian techniques. This allows the prior information, that a tephra unit co-located in several records should have an equivalent age, to be built into the Bayesian model (Buck et al., 2003). The deriving of ages for tephra does, however, mean that different age reporting protocols can cause confusion for tephra ages. For example the age usually reported for the Vedde Ash (12,171 b2k, MCE 114 yrs; Rasmussen et al., 2006) is reported in ice core years before A.D. 2000 but it is often incorporated into radiocarbon dated records that use the A.D. 1950 datum. Thus 50 years has to be deducted from the age of the Vedde prior to incorporating it into a radiocarbon-based age model. We recommend that when reporting the age of a tephra it is clear which system is being used to report and integrate the ages of tephra within a site chronology. This is particularly the case for newly reported tephra.

3.1.2. Correlating tephra records

A robust correlation between tephra records requires good stratigraphical control of the tephra deposit, a reasonable understanding of the age of the sequence, and secure chemical correlation of the tephra deposits between known eruptions. All of these criteria must be met before a correlation can be made. The stratigraphical issue is particularly significant in the case of distal cryptotephra, where there may be only a hundred or fewer shards of volcanic glass within a peak of a tephra layer. In some cases cryptotephra are deposited within a discrete unit of one or two centimetres and there is little evidence from the host sediment of reworking. This is usually the case with tephra in ice core records and in some lacustrine archives. However, in many cases (i.e. in marine sediments and peat sequences), shards of a particular ash may occur over many centimetres of the core and care must be taken when reporting the position of the tephra and when considering re-working. While the final decision on where to position a tephra isochron is site dependent, we would recommend as a minimum that shard counts for the whole profile are reported in a primary publication. In some cases it is also useful to consider detailed stratigraphic information such as thin section analyses of host sediment (Lane et al., 2013; Wulf et al., 2013).

The second key consideration is the age of the deposits that host the tephra horizons. While in some cases tephra shards have a unique chemical fingerprint (below) it is often the case that the reported tephra chemistry is very similar to other eruptions that are older or younger in age. Indeed Lane et al. (2012) report identical chemistry for the Vedde and Dimna ashes across major, minor and trace elements. Both are thought to be products of the Icelandic Katla Volcano and are separated by several thousand years. Nevertheless they are chemically indistinguishable. In this case discrimination is usually straightforward and the Vedde Ash is located in the mid Younger Dryas chronozone and in most records it is possible to distinguish this stratigraphic unit. A more problematic situation can arise, however, where there are multiple tephra layers with very similar chemical compositions that have been deposited over a short time interval. Here, great care must be taken to ensure precise correlation to other distal tephra and, where available, to proximal units. Key examples of the problem include work to correlate multiple tephra horizons in the long Lago di Monticchio record to proximal volcanic units in Italy and also to marine records in the Adriatic Sea (Wulf et al., 2004, 2008; Bourne et al., 2010; Smith et al., 2011).

The final key component to tephra correlation is the geochemical match between tephra products and, in some cases, the mineralogy of the tephra deposit. In the latter case, this is usually only useful for proximal ash deposits and for visible ash located relatively close to the volcanic source. As the goal of INTIMATE is the widespread correlation of records to test leads and lags it is critical that the far travelled component of tephra fallout is analysed in detail. In practice this usually means the vitreous component in the size fraction between 15 and 80 μm and occasionally up to 125 μm. These glasses can travel many thousands of kilometres from the source volcano and represent the most powerful, yet challenging, component of tephra fallout for correlation purposes. Vitreous tephra are normally analysed for a suite of major and minor elements by Electron Probe Micro Analyses (EPMA), and recently it has become common to add trace element profiles using LA-ICP-MS and Secondary Ion Mass Spectrometry (SIMS). The small size of these glass shards makes them often difficult to analyse. An additional problem is that in aqueous environments glasses can be prone to some degree of chemical alteration and secondary uptake of water (see e.g. Hunt and Hill, 2001; Tomlinson et al., 2010). Some researchers have argued that in ideal circumstances tephra correlations should be based on analyses on the same instrument to minimise analytical differences. While this view has significant merits it may limit the ability of the tephra community to undertake rapid correlation of tephra layers. It is more efficient to be able to determine tephra chemistry and correlate to data in the literature or appropriate databases. Moreover the chemical composition of major and now trace elements for many key tephra are well understood. We suggest, however, that there are a number of minimum requirements for the publication of tephra chemical data. These should always include instrument operating conditions and the use and the publication of analytical secondary standard data. These are samples of natural often fused glasses that have reported chemical ranges. A number of suitable standard glasses are available including standards available from USGS and MPI-DING.

3.1.3. Reporting issues

Due to the issues outlined above, it is critical that tephra are reported in the literature in an unambiguous manner. With many tephra layers having similar chemical compositions, correlations should be made carefully and new tephra layers only reported after substantial consideration of possible correlative tephra in the literature. As well as the stratigraphic, chronological and chemical considerations outlined above, we would also recommend that tephra are reported in a systematic manner. Ideally each tephra should be given a unique identifier, preferably an abbreviated site code and the depth of the reported isochron, which should also be reported as a range if the isochron is not clearly defined to one centimetre or less. In the primary paper that discusses the tephra at a site any proposed correlation should also be rigorously demonstrated making the case for chemical and chronological equivalence.

3.2. Cosmogenic ^{10}Be as a global time marker

Cosmogenic nuclides are constantly produced in the Earth's atmosphere through interactions of galactic cosmic rays with target nuclides present in air (mainly N_2, O_2, and the noble gases) (Lal and Peters, 1967). The flux of these highly energetic and charged galactic particles (mainly protons and alpha particles) is modulated by magnetic fields; the solar-induced interplanetary magnetic field and the Earth's magnetic field shield the incoming particles thereby inversely modulating the production of cosmogenic nuclides (Elsaesser et al., 1956). It is generally assumed that on long time scales (millennial and longer) this modulation is mainly caused by variations of the geomagnetic field. Because of the non-linear relationship between production and magnetic field (O'Brien, 1979), atmospheric cosmogenic nuclides are particularly sensitive to minima in the Earth's magnetic field strength. If, for example, the strength of the Earth's dipole field decreases to almost zero, the production of ^{10}Be more than doubles. On the other hand, a doubling of the recent dipole field strength would cause global ^{10}Be production to drop by only about 25% (Masarik and Beer, 1999).

The general principle of using cosmogenic nuclides as global time markers is relatively simple: geomagnetic events (see chapter 3.3) cause significant, globally and temporally synchronous increases in the production of atmospheric cosmogenic nuclides. The nuclides are deposited on the ocean, land and ice and eventually the production peaks are stored in sedimentary archives like ocean and lake sediments or ice cores. Thus, they may serve as global time markers to synchronise the age models of the individual records. The nuclide ^{10}Be is of particular interest in this context because the global production rate is relatively high (about 100 g ^{10}Be are produced every year) and because the atmospheric residence time of this aerosol-bound nuclide is low (about 1–2 yr).

Although, in practice, the quality of the recorded cosmogenic signals might be influenced by atmospheric circulation, oceanic transport, or any other climate or weather-induced transport process, there are techniques available to extract the production signal from the different archives (Christl et al., 2010; Steinhilber et al., 2012). The imprint, for example, of the most prominent geomagnetic excursion occurring in the INTIMATE time frame, the Laschamp excursion, is seen in many different archives and proxies (Fig. 5, for details see Section 3.3). These results show that peaks of cosmogenic ^{10}Be can be used to link different sediment archives, including ice cores and marine records. Also, in continental archives such as the Lake Lisan formation, the late Pleistocene precursor of the Dead Sea, a ^{10}Be peak was found at about 40 ka (Belmaker et al., 2008) indicating that lacustrine climate records can also be linked via ^{10}Be to their oceanic or glacial counterparts. Due to the long half-life of ^{10}Be (1.39 Ma) this technique is generally applicable for the past several millions of years.

3.3. Palaeomagnetic variations as isochrones

The Earth's magnetic field can be approximated by a geocentric axial dipole field with an axial dipole moment of 7.628×10^{22} Am^2 in 2010 (according to Finlay et al., 2010). Non-axial (non-dipole) components of the geomagnetic field are in the range of ~10% of the axial dipole field. The temporal variations of both the dipole and non-dipole terms are called secular variations, e.g. causing a permanent migration of the magnetic poles around the geographic poles at high latitudes. Major reversals of the Earth's magnetic (dipole) field are summarised in the Geomagnetic Polarity Time Scale (GPTS, e.g., Ogg and Smith, 2004). In stratigraphic sequences they act as isochronous tie points. Thus, magnetostratigraphy is a common method for dating long sedimentary records. The present day dipole field configuration is termed 'normal', whereas the opposite configuration, with field lines pointing from North to South, is termed 'reversed'. Commonly, phases of normal (reversed) polarity are marked as black (white) bars in magnetostratigraphic plots. This might suggest that the field is stable between its polarity reversals. The last major reversal occurred 780 ka ago, defining the onset of the Brunhes chron of normal polarity. Under certain conditions, sediments are able to record not only directional but also relative intensity changes of the geomagnetic field (e.g. Tauxe, 1993; Roberts and Winklhofer, 2004). The intensity of the natural remanent magnetisation (NRM) of a sedimentary magnetisation is

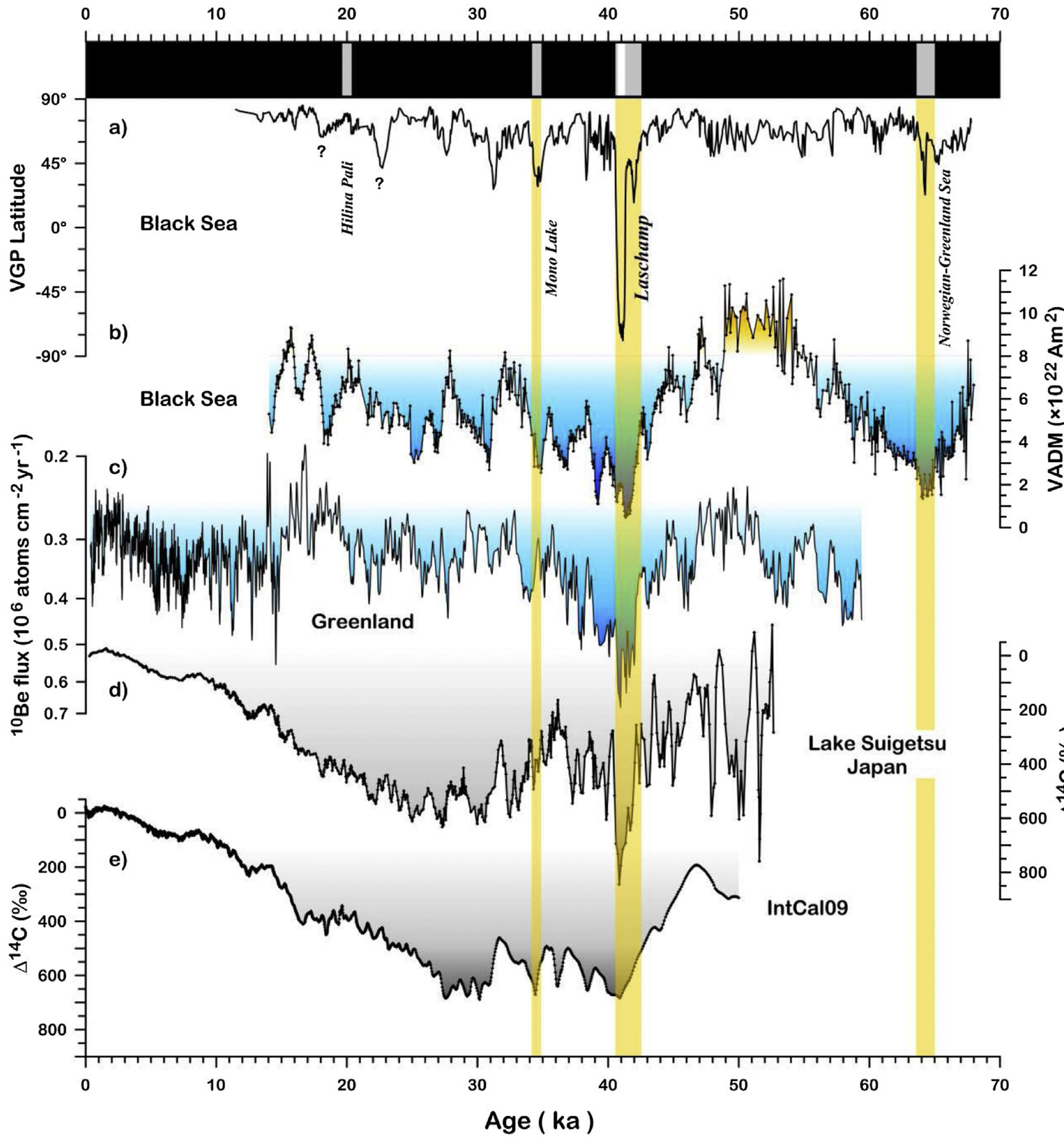

Fig. 5. High-resolution palaeomagnetic data from the Black Sea compared to records of cosmogenic radionuclides (Nowaczyk et al., 2012, 2013). Directional data in a) are shown as latitudes of virtual geomagnetic poles (VGP) and intensities b) as virtual axial dipole moments (VADM). The ^{10}Be flux in Greenland ice cores (Muscheler et al., 2005, pers. comm. 2013) are shown together with d) the Δ^{14}C record from Lake Suigetsu (Bronk Ramsey et al., 2012) and e) the older IntCal09 record (Reimer et al., 2009). In a) black (white) indicates normal (reversed) dipole polarity, whereas grey indicates phases of non-dipolar field geometry with associated geomagnetic excursions being marked by vertical yellow bars. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

then proportional to the concentration of magnetic particles and to the ambient field during their deposition. The latter relationship is linear for field strengths in the range of the geomagnetic field (up to ~70 μT). In order to isolate this field-dependent part of the NRM, its intensity has to be normalised by a parameter proportional to the concentration of magnetic particles (e.g., the anhysteretic remanent magnetisation, ARM). During the last two decades an increasing number of relative palaeointensity records have been used to create global composite records (e.g. Sint-200, Guyodo and Valet, 1996; PISO-1500, Channell et al., 2009). These stacks are mostly of intermediate temporal resolution (1 ky). A higher temporal resolution is provided by GLOPIS-75 (0.2 ky resolution; Laj et al., 2004). By comparison with absolute determinations from the thermomagnetic remanent magnetisation (TRM) of volcanic rocks, relative geomagnetic field variations obtained from sediments can be calibrated into variations of the virtual axial dipole moment (VADM). Thus, high-resolution sedimentary studies of geomagnetic field variations could show that the apparently 'stable' magnetic field, e.g. throughout the Brunhes chron, often underwent stability crises that can be detected in sediment records, thus allowing magnetostratigraphic methods to be applied also to sediment records within the INTIMATE time scale.

Such stability crises were characterised by more or less pronounced lows in geomagnetic field intensity with VADMs as low as

2.5×10^{22} Am^2. Some of these lows coincide with a distorted dipole geometry. Some of them have even been associated with a short-term, completely reversed dipole configuration, with minimum VADM values as low as 0.50×10^{22} Am^2, clearly less than 10% of the modern field (Nowaczyk et al., 2013). Such events are commonly termed 'geomagnetic excursions' (e.g., Laj and Channell, 2007), since the associated virtual geomagnetic pole (VGP) moves away from its usual migration area at high latitudes towards low latitudes, or in the case of a short-term reversal, to high latitudes of the opposite hemisphere, and then back again. The duration of these events, in terms of pronounced directional variations, is on the order of only a few hundred years (Nowaczyk et al., 2012) to a few thousand years (e.g. Laj et al., 2000; Knudsen et al., 2007; Bourne et al., 2012). The associated long-term palaeointensity lows (VADMs < 5×10^{22} Am^2) last, in general, 5–10 ky. On the other hand, there were also phases when the geomagnetic field was stronger than today by about 50%, that is, with VADMs reaching 12×10^{22} Am^2, e.g. at ~50 ka (Fig. 5).

The VADM maximum around 50 ka was preceded by a broad and deep low centred at 64.5 ka and is associated with the Norwegian-Greenland Sea excursion (Bleil and Gard, 1989). At high northern latitudes it is documented by a double swing to completely reversed directions (e.g., Nowaczyk and Frederichs, 1999). In sediments from the Black Sea, a mid-latitude site, anomalous directions only occur sporadically. The most pronounced geomagnetic feature during the past 100 ky is the Laschamp excursion at around 41 ka (Bonhommet and Babkine, 1967; Guillou et al., 2004; Plenier et al., 2007). It is characterised by a short-term full reversal of the geomagnetic field (Nowaczyk et al., 2012 and further references therein) and the lowest field intensities of the past 100 ky (Nowaczyk et al., 2013). Another palaeointensity low within the INTIMATE time frame at around 34.5 ka is linked with the Mono Lake excursion (Denham and Cox, 1971; Liddicoat and Coe, 1979; Laj and Channell, 2007; Kissel et al., 2011). This excursion is more an extreme secular variation feature rather than a short reversal. Another excursion, named Hilina Pali, is reported at around 20 ka and documented in lavas on Hawaii (e.g., Coe et al., 1978; Laj et al., 2002, 2011; Teanby et al., 2002) and probably in sediments of the high Arctic (Nowaczyk et al., 2003). It might be also related to one out of two inclination (VGP latitude) lows in Black Sea sediments (Fig. 5). A direct consequence of a weak magnetic field is an increased production of cosmogenic radionuclides (Fig. 5, see also chapter 3.2). The high-resolution palaeointensity record from the Black Sea (Nowaczyk et al., 2013) comprises many features that are also present in the ^{10}Be-flux in Greenland ice cores (Muscheler et al., 2005) and in the Δ^{14}C record from Lake Suigetsu (Bronk Ramsey et al., 2012).

4. Concluding remarks and recommendation

This review documents the great diversity of key dating techniques available for geological records of past climatic and environmental change covering the INTIMATE time frame 60–8 ka ago. Some of these methods, such as radiocarbon dating, have been applied for decades and across a wide range of archives, while others, such as cosmogenic nuclide dating, have only recently come into focus. Some methods are based on the decay of radiogenic isotopes measured on discrete samples using highly sophisticated and expensive instruments, while others count seasonal increments along core sequences with relatively easily accessible microscopes or even just by naked eye. All these methods have been improved and developed independently and with very specific foci on the materials to be dated. As a result of these diverse frameworks the respective dating communities have developed somewhat different traditions for reporting results and dealing with uncertainties. This often makes it difficult to compare the resulting chronologies on a solid and objective basis. However, since the demands of palaeoclimatic research are rapidly growing in terms of improved resolution and particularly in terms of precise integration of records from different regions, the dating communities are increasingly confronted with the need to harmonise their data outputs. The authors of this paper, belonging to nine different dating communities, regard harmonisation of chronologies for an improved integration of various proxy time series as a major challenge for the coming years. Nevertheless, we are also aware that the process of data harmonisation requires time and that those different communities may not yet be able to agree on common reporting standards. At this stage we, therefore, decline from recommending a standardised protocol for all dating techniques. However, we find it timely to recommend some basic requirements when reporting chronological data in publications in order to ensure reliable correlation of palaeoclimate records and to allow future re-evaluation of the data. The key recommendations from this paper are that:

- Proxy data should always be shown on both time and depth scales;
- Details of which age model has been applied (e.g. details on ^{14}C calibration procedure, name of ice core time scale, etc.) must be reported;
- The datum (reference year) must be specified, and if BP is used, it should be according to radiocarbon convention (present = A.D. 1950);
- Age uncertainties and counting error estimates should be reported together with all age estimates. It should be clearly stated which uncertainty contributions are included in the presented uncertainty measures and whether the reported uncertainties are Gaussian-style uncertainties or not, and if so, if 1σ or 2σ are given.

Last but not least, it is strongly recommended that chronologists should be included from the beginning (design) of a palaeoclimate project in order to choose the best dating methods, discuss an appropriate (sub)sampling strategy, and to synthesise, evaluate and if necessary recalculate published data.

Acknowledgements

This study is a contribution to the INTIMATE project, supported by INQUA and funded as the EU COST Action ES0907. Andreas Hendrich and Christine Gerschke (both GFZ Potsdam) helped with figure design and the reference list, respectively. We acknowledge helpful comments by Sabine Wulf. We would also like to thank Mike Walker and three anonymous reviewers for their valuable comments that helped us to improve an earlier version of this paper.

References

Aitchison, T.C., Leese, M., Michczynska, D.J., Mook, W.G., Otlet, R.L., Ottaway, B.S., Pazdur, M.F., Vanderplicht, J., Reimer, P.J., Robinson, S.W., Scott, E.M., Stuiver, M., Weninger, B., 1989. A comparison of methods used for the calibration of radiocarbon dates. Radiocarbon 31, 846–864.

Akçar, N., Deline, P., Ivy-Ochs, S., Alfimov, V., Hajdas, I., Kubik, P.W., Christl, M., Schlüchter, C., 2012. The 1717 AD rock avalanche deposits in the upper Ferret Valley (Italy): a dating approach with cosmogenic ^{10}Be. J. Quat. Sci. 27, 383–392. http://dx.doi.org/10.1002/jqs.1558.

Andersen, K.K., Svensson, A., Johnsen, S.J., Rasmussen, S.O., Bigler, M., Röthlisberger, R., Ruth, U., Siggaard-Andersen, M.-L., Steffensen, J.P., Dahl-Jensen, D., Vinther, B.M., Clausen, H.B., 2006. The Greenland Ice Core Chronology 2005, 15–42 ka. Part 1: constructing the time scale. Quat. Sci. Rev. 25, 3246–3257. http://dx.doi.org/10.1016/j.quascirev.2006.08.002.

Arnold, J.R., Libby, W.F., 1951. Radiocarbon dates. Science 113, 111–120.

Balco, G., Stone, J.O., Lifton, N.A., Dunai, T.J., 2008. A complete and easily accessible means of calculating surface exposure ages or erosion rates from ^{10}Be and ^{26}Al measurements. Quat. Geochronol. 3, 174–195. http://dx.doi.org/10.1016/j.quageo.2007.12.001.

Bayliss, A., 2009. Rolling out revolution: using radiocarbon dating in archaeology. Radiocarbon 51, 123–147.

Bazin, L., Landais, A., Lemieux-Dudon, B., Toyé Mahamadou Kele, H., Veres, D., Parrenin, F., Martinerie, P., Ritz, C., Capron, E., Lipenkov, V., Loutre, M.-F., Raynaud, D., Vinther, B., Svensson, A., Rasmussen, S.O., Severi, M., Blunier, T., Leuenberger, M., Fischer, H., Masson-Delmotte, V., Chappellaz, J., Wolff, E., 2013. An optimized multi-proxy, multi-site Antarctic ice and gas orbital chronology (AICC2012): 120–800 ka. Clim. Past 9, 1715–1731. http://dx.doi.org/10.5194/cp-9-1715-2013.

Belmaker, R., Lazar, B., Tepelyakov, N., Stein, M., Beer, J., 2008. ^{10}Be in Lake Lisan sediments – a proxy for production or climate? Earth Planet. Sci. Lett. 269, 448–457. http://dx.doi.org/10.1016/j.epsl.2008.02.032.

Blaauw, M., Heuvelink, G.B.M., Mauquoy, D., van der Plicht, J., van Geel, B., 2003. A numerical approach to C-14 wiggle-match dating of organic deposits: best fits and confidence intervals. Quat. Sci. Rev. 22, 1485–1500. http://dx.doi.org/10.1016/S0277-3791(03)00086-6.

Blaauw, M., Christen, J.A., 2005. Radiocarbon peat chronologies and environmental change. J. R. Stat. Soc. Ser. C (Appl. Stat.) 54, 805–816. http://dx.doi.org/10.1111/j.1467-9876.2005.00516.x.

Blauuw, M., Christen, J.A., 2011. Flexible paleoclimate age-depth models using an autoregressive gamma process. Bayesian Anal. 6, 457–464. http://dx.doi.org/10.1214/ba/1339616472.

Bleil, U., Gard, G., 1989. Chronology and correlation of Quaternary magnetostratigraphy and nannofossil biostratigraphy in Norwegian-Greenland Sea sediments. Geol. Rundsch. 78, 1173–1187.

Blockley, S.P.E., Lane, C.S., Lotter, A.F., Pollard, A.M., 2007. Evidence for the presence of the Vedde Ash in Central Europe. Quat. Sci. Rev. 26, 3030–3036. http://dx.doi.org/10.1016/j.quascirev.2007.09.010.

Blockley, S.P.E., Bronk Ramsey, C., Pyle, D.M., 2008. Improved age modelling and high-precision age estimates of late Quaternary tephras, for accurate palaeoclimate reconstruction. J. Volcanol. Geotherm. Res. 177, 251–262. http://dx.doi.org/10.1016/j.jvolgeores.2007.10.015.

Blockley, S.P.E., Lane, C.S., Hardiman, M., Rasmussen, S.O., Seierstad, I.K., Steffensen, J.P., Svensson, A., Lotter, A.F., Turney, C.S.M., Bronk Ramsey, C., INTIMATE members, 2012. Synchronisation of palaeoenvironmental records over the last 60,000 years, and an extended INTIMATE(1) event stratigraphy to 48,000 b2k. Quat. Sci. Rev. 36, 2–10. http://dx.doi.org/10.1016/j.quascirev.2011.09.017.

Blunier, T., Brook, E.J., 2001. Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. Science 291, 109–112. http://dx.doi.org/10.1126/science.291.5501.109.

Bonhommet, N., Babkine, J., 1967. Sur la présence d'aimantation inversée dans la Chaîne des Puys. C. R. l'Acad. Sci. Paris 264, 92–94.

Bourdon, B., Henderson, G.M., Lundstrom, C.C., Turner, S.P. (Eds.), 2003. Uranium-series Geochemistry. Reviews in Mineralogy and Geochemistry, vol. 52. Mineralogical Society of America, Washington D.C.

Bourne, A.J., Lowe, J.J., Trincardi, F., Asioli, A., Blockley, S.P.E., Wulf, S., Matthews, I.P., Piva, A., Vigliotti, L., 2010. Distal tephra record for the last ca 105,000 years from core PRAD 1-2 in the central Adriatic Sea: implications for marine tephrostratigraphy. Quat. Sci. Rev. 29, 3079–3094. http://dx.doi.org/10.1016/j.quascirev.2010.07.021.

Bourne, M., Mac Niocail, C., Thomas, A.L., Knudsen, M.F., Henderson, G.M., 2012. Rapid directional changes associated with a 6.5 kyr-long Blake geomagnetic excursion at the Blake-Bahama Outer Ridge. Earth Planet. Sci. Lett. 333–334, 21–34. http://dx.doi.org/10.1016/j.epsl.2012.04.017.

Brauer, A., Endres, C., Negendank, J.F.W., 1999a. Lateglacial calendar year chronology based on annually laminated sediments from Lake Meerfelder Maar, Germany. Quat. Int. 61, 17–25. http://dx.doi.org/10.1016/S1040-6182(99)00014-2.

Brauer, A., Endres, C., Günter, C., Litt, T., Stebich, M., Negendank, J.F.W., 1999b. High resolution sediment and vegetation responses to Younger Dryas climate change in varved lake sediments from Meerfelder Maar, Germany. Quat. Sci. Rev. 18, 321–329. http://dx.doi.org/10.1016/S0277-3791(98)00084-5.

Brauer, A., 2004. Annually laminated lake sediments and their palaeoclimatic relevance. In: Miller, H., Negendank, J.F.W., Flöser, G., von Storch, H., Fischer, H., Lohmann, G., Kumke, T. (Eds.), The KIHZ Project: Towards a Synthesis of Holocene Proxy Data and Climate Models. Springer-Verlag, Berlin Heidelberg, pp. 109–128.

Brauer, A., Allen, J.R.M., Mingram, J., Dulski, P., Wulf, S., Huntley, B., 2007. Evidence for last interglacial chronology and environmental change from Southern Europe. Proc. Natl. Acad. Sci. U. S. A. (PNAS) 104, 450–455. http://dx.doi.org/10.1073/pnas.0603321104.

Bronk Ramsey, C., 1995. Radiocarbon calibration and analysis of stratigraphy: the OxCal program. Radiocarbon 37, 425–430.

Bronk Ramsey, C., 1998. Probability and dating. Radiocarbon 40, 461–474.

Bronk Ramsey, C., 2000. Comment on 'the use of Bayesian statistics for ^{14}C dates of chronologically ordered samples: a critical analysis'. Radiocarbon 42, 199–202.

Bronk Ramsey, C., 2008. Deposition models for chronological records. Quat. Sci. Rev. 27, 42–60. http://dx.doi.org/10.1016/j.quascirev.2007.01.019.

Bronk Ramsey, C., 2009a. Bayesian analysis of radiocarbon dates. Radiocarbon 51, 337–360.

Bronk Ramsey, C., 2009b. Dealing with outliers and offsets in radiocarbon dating. Radiocarbon 51, 1023–1045.

Bronk Ramsey, C., Dee, M., Lee, S., Nakagawa, T., Staff, R.A., 2010. Developments in the calibration and modeling of radiocarbon dates. Radiocarbon 52, 953–961.

Bronk Ramsey, C., Staff, R.A., Bryant, C.L., Brock, F., Kitagawa, H., van der Plicht, J., Schlolaut, G., Marshall, M.H., Brauer, A., Lamb, H.F., Payne, R.L., Tarasov, P.E., Haraguchi, T., Gotanda, K., Yonenobu, H., Yokoyama, Y., Tada, R., Nakagawa, T., 2012. A complete terrestrial radiocarbon record for 11.2-52.8 kyr BP. Science 338, 370–374. http://dx.doi.org/10.1126/science.1226660.

Buck, C.E., Cavanagh, W.G., Litton, C.D., 1996. The Bayesian Approach to Interpreting Archaeological Data. Wiley, Chichester.

Buck, C.E., Christen, J.A., James, G.N., 1999. BCal: an on-line Bayesian radiocarbon calibration tool. Internet Archaeol. 7. http://intarch.ac.uk/journal/issue7/buck/toc.html.

Buck, C.E., Higham, T.F.G., Lowe, D.J., 2003. Bayesian tools for tephrochronology. Holocene 13, 639–647. http://dx.doi.org/10.1191/0959683603hl652ft.

Buylaert, J.P., Jain, M., Murray, A.S., Thomsen, K.J., Thiel, C., Sohbati, R., 2012. A robust feldspar luminescence dating method for Middle and Late Pleistocene sediments. Boreas 41, 435–451. http://dx.doi.org/10.1111/j.1502-3885.2012.00248.x.

Channel, J.E.T., Xuan, C., Hodell, D.A., 2009. Stacking paleointensity and oxygen isotope data for the last 1.5 Myr (PISO-1500). Earth Planet. Sci. Lett. 283, 14–23. http://dx.doi.org/10.1016/j.epsl.2009.03.012.

Chapot, M.S., Roberts, H.M., Duller, G.A.T., Lai, Z.P., 2012. A comparison of natural- and laboratory-generated dose response curves for quartz optically stimulated luminescence signals from Chinese Loess. Radiat. Meas. 47, 1045–1052. http://dx.doi.org/10.1016/j.radmeas.2012.09.001.

Cheng, H., Edwards, R.L., Hoff, J., Gallup, C.D., Richards, D.A., Asmerom, Y., 2000. The half-lives of U-234 and Th-230. Chem. Geol. 169, 17–33. http://dx.doi.org/10.1016/S0009-2541(99)00157-6.

Cheng, H., Edwards, R.L., Shen, C.-C., Polyak, V.J., Asmerom, Y., Woodhead, J., Hellstrom, J., Wang, Y., Kong, X., Spötl, C., Wang, X., Alexander, E.C., 2013. Improvements in ^{230}Th dating, ^{230}Th and ^{234}U half-life values, and U-Th isotopic measurements by multi-collector inductively coupled plasma mass spectrometry. Earth Planet. Sci. Lett. 371–372, 82–91. http://dx.doi.org/10.1016/j.epsl.2013.04.006.

Christl, M., Lippold, J., Steinhilber, F., Bernsdorff, F., Mangini, A., 2010. Reconstruction of global ^{10}Be production over the past 250 ka from highly accumulating Atlantic drift sediments. Quat. Sci. Rev. 29, 2663–2672. http://dx.doi.org/10.1016/j.quascirev.2010.06.017.

Coe, R.S., Grommé, C.S., Mankinen, E.A., 1978. Geomagnetic paleointensities from radiocarbon-dated lava flows on Hawaii and the question of the Pacific non-dipole low. J. Geophys. Res. Solid Earth 83, 1740–1756.

Dansgaard, W., 1964. Stable isotopes in precipitation. Tellus 16, 436–468.

Davis, R., Schaeffer, O.A., 1955. Chlorine-36 in nature. Ann. N. Y. Acad. Sci. 62, 107–121.

Davies, S.M., Branch, N.P., Lowe, J.J., Turney, C.S.M., 2002. Towards a European tephrochronological framework for termination 1 and the early Holocene. Philos. Trans. R. Soc. Lond. A 360, 767–802. http://dx.doi.org/10.1098/rsta.2001.0964.

Davies, S.M., Wastegård, S., Abbott, P.M., Barbante, C., Bigler, M., Johnsen, S.J., Rasmussen, T.L., Steffensen, J.P., Svensson, A., 2010. Tracing volcanic events in the NGRIP ice-core and synchronising North Atlantic marine records during the last glacial period. Earth Planet. Sci. Lett. 294, 69–79. http://dx.doi.org/10.1016/j.epsl.2010.03.004.

De Geer, G., 1912. A geochronology of the last 12000 years. In: Compte rendu 11 Congress Geologique International, 11th International Geological Congress (1910), Stockholm, Sweden, vol. 1, pp. 241–253. German reprint in 1912 with English translation: Geologische Rundschau 3, 457–471.

Denham, C.R., Cox, A., 1971. Evidence that the Laschamp event did not occur 13300-30400 years ago. Earth Planet. Sci. Lett. 13, 181–190. http://dx.doi.org/10.1016/0012-821X(71)90122-1.

de Vivo, B., Rolandi, G., Gans, P.B., Calvert, A., Bohrson, W.A., Spera, F.J., Belkin, H.E., 2001. New constraints on the pyroclastic eruptive history of the Campanian volcanic Plain (Italy). Mineral. Petrol. 73, 47–65. http://dx.doi.org/10.1007/s007100170010.

Douville, E., Sallé, E., Frank, N., Eisele, M., Pons-Brachu, E., Ayrault, S., 2010. Rapid and accurate U–Th dating of ancient carbonates using inductively coupled plasma-quadrupole mass spectrometry. Chem. Geol. 272, 1–11. http://dx.doi.org/10.1016/j.chemgeo.2010.01.007.

Duller, G.A.T., 2008. Luminescence Dating: Guidelines on Using Luminescence Dating in Archaeology. English Heritage, Swindon.

Duller, G.A.T., 2011. What date is it? Should there be an agreed datum for luminescence ages? Anc. TL 29, 1–3.

Dunai, T.J., Stuart, F.M., 2009. Reporting of cosmogenic nuclide data for exposure age and erosion rate determinations. Quat. Geochronol. 4, 437–440. http://dx.doi.org/10.1016/j.quageo.2009.04.003.

Dunai, T.J., 2010. Cosmogenic Nuclides: Principles, Concepts and Applications in the Earth Surface Sciences. Cambridge University Press, Cambridge.

Edwards, R.L., Chen, J.H., Wasserburg, G.J., 1987. ^{238}U, ^{234}U, ^{230}Th, ^{232}Th systematics and the precise measurement of time over the past 500,000 years. Earth Planet. Sci. Lett. 81, 175–192. http://dx.doi.org/10.1016/0012-821X(87)90154-3.

Elsaesser, W., Ney, E.P., Winckler, J.R., 1956. Cosmic-ray intensity and geomagnetism. Nature 178, 1226–1227.

Feurdean, A., Persoiu, A., Tantau, I., Stevens, T., Markovic, S., Magyari, E.K., Onac, B.B., Andric, M., Connor, S., Galka, M., Hoek, W.Z., Lamentowicz, M., Sümegi, P., Persoiu, I., Kolaczek, P., Kunes, P., Marinova, E., Slowinski, M., Michczynska, D., Stancikaite, M., Svensson, A., Veski, S., Farcas, S., Tamas, T., Zernitskaya, V.,

Timar, A., Tonkov, S., Toth, M., Willis, K.J., Plóciennik, M., Gaudeny, T., 2014. Climate variability and associated vegetation response throughout Central and Eastern Europe (CEE) between 8 and 60 kyrs ago. Quat. Sci. Rev. 106, 206–224.
Finlay, C.C., Maus, S., Beggan, C.D., Bondar, T.N., Chambodut, A., Chernova, T.A., Chulliat, A., Golovkov, V.P., Hamilton, B., Hamoudi, M., Holme, R., Hulot, G., Kuang, W., Langlais, B., Lesur, V., Lowes, F.J., Lühr, H., Macmillan, S., Mandea, M., McLean, S., Manoj, C., Menvielle, M., Michaelis, I., Olsen, N., Rauberg, J., Rother, M., Sabaka, T.J., Tangborn, A., Tøffner-Clausen, L., Thébault, E., Thomson, A.W.P., Wardinski, I., Wei, Z., Zvereva, T.I., 2010. International geomagnetic reference field: the eleventh generation. Geophys. J. Int. 183, 1216–1230. http://dx.doi.org/10.1111/j.1365-246X.2010.04804.x.
Francus, P., Keimig, F., Besonen, M., 2002. An algorithm to aid varve counting and measurement from thin-sections. J. Paleolimnol. 28, 283–286. http://dx.doi.org/10.1023/A:1021624415920.
Frappier, A.B., 2008. A stepwise screening system to select storm-sensitive stalagmites: taking a targeted approach to speleothem sampling methodology. Quat. Int. 187, 25–39. http://dx.doi.org/10.1016/j.quaint.2007.09.042.
Godwin, H., 1962. Half-life of radiocarbon. Nature 195, 984.
Goldstein, R.H., Stirling, C.H., 2003. Techniques for measuring U-series nuclides: 1992-2002. In: Bourdon, B., Henderson, G.M., Lundstrom, C.C., Turner, S.P. (Eds.), Uranium-series Geochemistry, Reviews in Mineralogy and Geochemistry, vol. 52. Mineralogical Society of America, Washington, pp. 23–57.
Gosse, J.C., Phillips, F.M., 2001. Terrestrial in situ cosmogenic nuclides: theory and application. Quat. Sci. Rev. 20, 1475–1560. http://dx.doi.org/10.1016/S0277-3791(00)00171-2.
Gosse, J.C., 2005. The contributions of cosmogenic nuclides to unraveling alpine paleo-glacier histories. In: Huber, U.M., Bugmann, H.K.M., Reasoner, M.A. (Eds.), Global Change and Mountain Regions. An Overview of Current Knowledge, Advances in Global Change Research, vol. 23. Springer, Dordrecht, pp. 39–50.
Guillou, H., Singer, B.S., Laj, C., Kissel, C., Scaillet, S., Jicha, B.R., 2004. On the age of the Laschamp excursion. Earth Planet. Sci. Lett. 227, 331–343. http://dx.doi.org/10.1016/j.epsl.2004.09.018.
Guyodo, Y., Valet, J.-P., 1996. Relative variations in geomagnetic intensity from sedimentary records: the past 200,000 years. Earth Planet. Sci. Lett. 143, 23–36. http://dx.doi.org/10.1016/0012-821X(96)00121-5.
Hajdas, I., 2008. The radiocarbon dating method and its applications in Quaternary studies. Quat. Sci. J. Eiszeitalt. Ggw. 57, 2–24.
Hajdas, I., Michczynski, A., 2010. Age-depth model of Lake Soppensee (Switzerland) based on the high-resolution ^{14}C chronology compared with varve chronology. Radiocarbon 52, 1027–1040.
Heiri, O., Koinig, K., Spötl, C., Barrett, S., Brauer, A., Drescher-Schneider, R., Gaar, D., Ivy-Ochs, S., Kerschner, H., Luetscher, M., Nicolussi, K., Moran, A., Preusser, F., Schmidt, R., Schoeneich, P., Schwörer, C., Sprafke, T., Terhorst, B., Willy Tinner, W., 2014. Palaeoclimate records 60-8 ka in the Austrian and Swiss Alps and their forelands. Quat. Sci. Rev. 106, 186–205.
Hellstrom, J., 2003. Rapid and accurate U/Th dating using parallel-ion counting multi-collector ICP-MS. J. Anal. At. Spectrom. 18, 1346–1351. http://dx.doi.org/10.1039/B308781F.
Hellstrom, J., 2006. U-Th dating of speleothems with high initial ^{230}Th using stratigraphical constraint. Quat. Geochronol. 1, 289–295. http://dx.doi.org/10.1016/j.quageo.2007.01.004.
Heyman, J., Stroeven, A.P., Harbor, J.M., Caffee, M.W., 2011. Too young or too old: evaluating cosmogenic exposure dating based on an analysis of compiled boulder exposure ages. Earth Planet. Sci. Lett. 302, 71–80. http://dx.doi.org/10.1016/j.epsl.2010.11.040.
Heyman, J., 2014. Paleoglaciation of the Tibetan Plateau and surrounding mountains based on exposure ages and ELA depression estimates. Quat. Sci. Rev. 91, 30–41. http://dx.doi.org/10.1016/j.quascirev.2014.03.018.
Hoffmann, D.L., 2008. ^{230}Th isotope measurements of femtogram quantities for U-series dating using multi-ion counting (MIC) MC-ICPMS. Int. J. Mass Spectrom. 275, 75–79. http://dx.doi.org/10.1016/j.ijms.2008.05.033.
Hoffmann, D.L., Spötl, C., Mangini, A., 2009. Micromill and in situ laser ablation sampling techniques for high spatial resolution MC-ICPMS U/Th dating of carbonates. Chem. Geol. 259, 253–261. http://dx.doi.org/10.1016/j.chemgeo.2008.11.015.
Hogg, A.G., Hua, Q., Blackwell, P.G., Niu, M., Buck, C.E., Guilderson, T.P., Heaton, T.J., Palmer, J.G., Reimer, P.J., Reimer, R.W., Turney, C.S.M., Zimmerman, S.R.H., 2013. SHCal13 southern hemisphere calibration, 0–50,000 cal yr BP. Radiocarbon 55, 1889–1903. http://dx.doi.org/10.2458/azu_js_rc.55.16783.
Hua, Q., Barbetti, M., Rakowski, A.Z., 2013. Atmospheric radiocarbon for the period 1950–2010. Radiocarbon 55, 2059–2072. http://dx.doi.org/10.2458/azu_js_rc.v55i2.16177.
Hughen, K., Baillie, M., Bard, E., Beck, J., Bertrand, C., Blackwell, P., Buck, C., Burr, G., Cutler, K., Damon, P., Edwards, R., Fairbanks, R., Friedrich, M., Guilderson, T., Kromer, B., McCormac, G., Manning, S., Ramsey, C., Reimer, P., Reimer, R., Remmele, S., Southon, J., Stuiver, M., Talamo, S., Taylor, F., van der Plicht, J., Weyhenmeyer, C., 2004. Marine04 marine radiocarbon age calibration, 0-26 cal kyr BP. Radiocarbon 46, 1059–1086.
Hunt, J.B., Hill, P.G., 2001. Tephrological implications of beam size – sample-size effects in electron microprobe analysis of glass shards. J. Quat. Sci. 16, 105–117. http://dx.doi.org/10.1002/jqs.571.
Huntley, D.J., Lamothe, M., 2001. Ubiquity of anomalous fading in K-feldspars and the measurement and correction for it in optical dating. Can. J. Earth Sci. 38, 1093–1106.
Huntley, D.J., Hutton, J.T., Prescott, J.R., 1993. The stranded beach-dune sequence of south-east South Australia: a test of thermoluminescence dating, 0-800 ka. Quat. Sci. Rev. 12, 1–20.
Ivy-Ochs, S., Kerschner, H., Schlüchter, C., 2007. Cosmogenic nuclides and the dating of Lateglacial and early Holocene glacier variations: the Alpine perspective. Quat. Int. 164–165, 53–63. http://dx.doi.org/10.1016/j.quaint.2006.12.008.
Ivy-Ochs, S., Kerschner, H., Reuther, A., Preusser, F., Heine, K., Maisch, M., Kubik, P.W., Schlüchter, C., 2008. Chronology of the last glacial cycle in the European Alps. J. Quat. Sci. 23, 559–573. http://dx.doi.org/10.1002/jqs.1202.
Ivy-Ochs, S., Kober, F., 2008. Surface exposure dating with cosmogenic nuclides. Quat. Sci. J. 57, 157–189.
Ivy-Ochs, S., Schaller, M., 2010. Examining processes and rates of landscape change with cosmogenic radionuclides. In: Froehlich, K. (Ed.), Environmental Radionuclides – Tracers and Timers of Terrestrial Processes. Elsevier, Amsterdam, pp. 232–294.
Ivy-Ochs, S., Dühnforth, M., Densmore, A.L., Alfimov, V., 2013. Dating fan deposits with cosmogenic nuclides. In: Dating torrential processes on fans and cones. Springer Verlag, The Netherlands, pp. 243–263.
Johnsen, S.J., Clausen, H.B., Cuffey, K.M., Hoffmann, G., Schwander, J., Creyts, T., 2000. Diffusion of stable isotopes in polar firn and ice: the isotope effect in firn diffusion. In: Hondoh, T. (Ed.), Physics of Ice Core Records. Hokkaido University Press, Sapporo, pp. 121–140.
Jones, M., Nicholls, G., 2002. New radiocarbon calibration software. Radiocarbon 44, 663–674.
Kaufman, A., Broecker, W.S., 1965. Comparison of Th^{230} and C^{14} ages for carbonate materials from lakes Lahontan and Bonneville. J. Geophys. Res. Lett. 70, 4039–4054. http://dx.doi.org/10.1029/JZ070i016p04039.
Kelts, K., Hsü, K.J., 1978. Freshwater carbonate sedimentation. In: Lerman, A. (Ed.), Lakes. Geology, Chemistry, Physics. Springer-Verlag, New York, pp. 295–323.
Kissel, C., Guillou, H., Laj, C., Carracedo, J.C., Nomade, S., Perez-Torrado, F., Wandres, C., 2011. The Mono Lake excursion recorded in phonolitic lavas from Tenerife (Canary Islands): paleomagnetic analyses and coupled K/Ar and Ar/Ar dating. Phys. Earth Planet. Inter. 187, 232–244. http://dx.doi.org/10.1016/j.pepi.2011.04.014.
Knudsen, M.F., Henderson, G.M., Mac Niocail, C., West, A.J., 2007. Seven thousand year duration for a geomagnetic excursion constrained by $^{230}Th_{xs}$. Geophys. Res. Lett. 34, L22302. http://dx.doi.org/10.1029/2007GL031090.
Laj, C., Kissel, C., Mazaud, A., Channell, J.E.T., Beer, J., 2000. North Atlantic palaeointensity stack since 75 ka (NAPIS75) and the duration of the Laschamp event. Philos. Trans. R. Soc. Lond. A 358, 1009–1025. http://dx.doi.org/10.1098/rsta.2000.0571.
Laj, C., Kissel, C., Scao, V., Beer, J., Thomas, D.M., Guillou, H., Muscheler, R., Wagner, G., 2002. Geomagnetic intensity and inclination variations at Hawaii for the past 98 kyr from core SOH-4 (Big Island): a new study and a comparison with existing contemporary data. Phys. Earth Planet. Inter. 129, 205–243. http://dx.doi.org/10.1016/S0031-9201(01)00291-6.
Laj, C., Kissel, C., Beer, J., 2004. High resolution global paleointensity stack since 75 kyr (GLOPIS-75) calibrated to absolute values. In: Channell, J.E.T., Kent, D.V., Lowrie, W., Meert, J.G. (Eds.), Timescales of the Paleomagnetic Field, Geophysical Monograph Series, vol. 145. American Geophysical Union, Washington, D.C, pp. 255–265.
Laj, C., Channell, J.E.T., 2007. Geomagnetic excursions. In: Schubert, G., Bercovici, D., Dziewonski, A., Hering, T., Kanamori, H., Kono, M., Olson, P.L., Price, G.D., Romanowicz, B., Spohn, T., Stevenson, D., Watts, A.B. (Eds.), Treatise on Geophysics, Geomagnetism, vol. 5. Elsevier, B.V, Amsterdam, pp. 373–416.
Laj, C., Kissel, C., Davies, C., Gubbins, D., 2011. Geomagnetic field intensities and inclination records from Hawaii and the Réunion island: geomagnetic implications. Phys. Earth Planet. Inter. 187, 170–187. http://dx.doi.org/10.1016/j.pepi.2011.05.007.
Lal, D., Peters, B., 1967. Cosmic ray produced radioactivity on the earth. In: Handbuch der Physik/Encyclopedia of Physics. Kosmische Strahlung II/Cosmic Rays II, vol. 9/46/2. Springer-Verlag, Berlin Heidelberg New York, pp. 551–612.
Lambert, A.M., Hsü, K.J., 1979. Varve-like sediments of the Walensee, Switzerland. In: Schlüchter, C. (Ed.), Moraines and Varves; Origin, Genesis, Classification. A.A. Balkema, Rotterdam, pp. 287–294.
Lamoureux, S.F., 2001. Varve chronology techniques. In: Last, W.M., Smol, J.P. (Eds.), Developments in Paleoenvironmental Research (DPER), Tracking Environmental Change Using Lake Sediments: Basin Analysis, Coring, and Chronological Techniques, vol. 1. Kluwer, Dordrecht, pp. 247–260.
Lane, C.S., Andrič, M., Cullen, V.L., Blockley, S.P.E., 2011. The occurrence of distal Icelandic and Italian tephra in the Lateglacial of Lake Bled, Slovenia. Quat. Sci. Rev. 30, 1013–1018. http://dx.doi.org/10.1016/j.quascirev.2011.02.014.
Lane, C.S., Blockley, S.P.E., Mangerud, J., Smith, V.C., Lohne, O.S., Tomlinson, E.L., Matthews, I.P., Lotter, A.F., 2012. Was the 12.1 ka Icelandic Vedde Ash one of a kind? Quat. Sci. Rev. 33, 87–99. http://dx.doi.org/10.1016/j.quascirev.2011.11.011.
Lane, C.S., Brauer, A., Blockley, S.P.E., Dulski, P., 2013. Volcanic ash reveals time-transgressive abrupt climate change during the Younger Dryas. Geology. http://dx.doi.org/10.1130/G34867.1.
Libby, W.F., Anderson, E.C., Arnold, J.R., 1949. Age determination by radiocarbon content: world-wide assay of natural radiocarbon. Science 109, 227–228.
Liddicoat, J.C., Coe, R.S., 1979. Mono Lake geomagnetic excursion. J. Geophys. Res. Solid Earth 84, 261–271.
Lotter, A.F., Lemcke, G., 1999. Methods for preparing and counting biochemical varves. Boreas 28, 243–252. http://dx.doi.org/10.1111/j.1502-3885.1999.tb00218.x.

Lowe, D.J., 2011. Tephrochronology and its application: a review. Quat. Geochronol. 6, 107–153.
Lowe, J., Ammann, B., Birks, H., Bjorck, S., Coope, G., Cwynar, L., Debeaulieu, J., Mott, R., Peteet, D., Walker, M., 1994. Climatic changes in areas adjacent to the North Atlantic during the last glacial-interglacial transition (14-9 ka BP) – a contribution to IGCP-253. J. Quat. Sci. 9, 185–198.
Luo, X.Z., Rehkämper, M., Lee, D.C., Halliday, A.N., 1997. High precision $^{230}Th/^{232}Th$ and $^{234}U/^{238}U$ measurements using energy filtered ICP magnetic sector multiple collector mass spectrometry. Int. J. Mass Spectrom. 171, 105–117.
Mangili, C., Brauer, A., Plessen, B., Moscariello, A., 2007. Centennial-scale oscillations in oxygen and carbon isotopes of endogenic calcite from a 15,500 varve year record of the Piànico interglacial. Quat. Sci. Rev. 26, 1725–1735. http://dx.doi.org/10.1016/j.quascirev.2007.04.012.
Marshall, M., Schlolaut, G., Nakagawa, T., Lamb, H., Brauer, A., Staff, R., Bronk Ramsey, C., Tarasov, P., Gotanda, K., Haraguchi, T., Yokoyama, Y., Yonenobu, H., Tada, R., 2012. A novel approach to varve counting using μXRF and X-radiography in combination with thin-section microscopy, applied to the Late Glacial chronology from Lake Suigetsu, Japan. Quat. Geochronol. 13, 70–80. http://dx.doi.org/10.1016/j.quageo.2012.06.002.
Masarik, J., Beer, J., 1999. Simulation of particle fluxes and cosmogenic nuclide production in the Earth's atmosphere. J. Geophys. Res. Atmos. 104, 12099–12111.
Millard, A., 2014. Conventions for reporting radiocarbon determinations. Radiocarbon 56 (2), 555–559. http://dx.doi.org/10.2458/56.17455.
Mingram, J., Negendank, J.F.W., Brauer, A., Berger, D., Hendrich, A., Köhler, M., Usinger, H., 2007. Long cores from small lakes – recovering up to 100 m-long lake sediment sequences with a high-precision rod-operated piston corer (Usinger-corer). J. Paleolimnol. 37, 517–528. http://dx.doi.org/10.1007/s10933-006-9035-4.
Mook, W.G., van der Plicht, J., 1999. Reporting C-14 activities and concentrations. Radiocarbon 41, 227–239.
Moreno, A., Svensson, A., Brooks, S., Connors, S., Engels, S., Fletcher, W., Genty, D., Heiri, O., Labuhn, I., Peyron, O., Sadori, L., Wulf, S., Giovanni Zanchetta, G., data contributors, 2014. A compilation of Western European terrestrial records 60–8 ka BP: towards an understanding of latitudinal climatic gradients. Quat. Sci. Rev. 106, 167–185.
Mortensen, A.K., Bigler, M., Grönvold, K., Steffensen, J.P., Johnsen, S.J., 2005. Volcanic ash layers from the Last Glacial Termination in the NGRIP ice core. J. Quat. Sci. 20, 209–219. http://dx.doi.org/10.1002/jqs.908.
Muhs, D.R., Bettis III, E.A., Roberts, H.M., Harlan, S.S., Paces, J.B., Reynolds, R.L., 2013. Chronology and provenance of last-glacial (Peoria) loess in western Iowa and paleoclimatic implications. Quat. Res. 80, 468–481. http://dx.doi.org/10.1016/j.yqres.2013.06.006.
Müller, H., 1974. Pollenanalytische Untersuchungen und Jahresschichtenzählungen an der eem-zeitlichen Kieselgur von Bispingen/Luhe. Geol. Jahrb. A 21, 149–169.
Murray, A.S., Wintle, A.G., 2000. Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. Radiat. Meas. 32, 57–73. http://dx.doi.org/10.1016/S1350-4487(99)00253-X.
Murray, A.S., Olley, J.M., 2002. Precision and accuracy in the optically stimulated luminescence dating of sedimentary quartz: a status review. Geochronometria 21, 1–16.
Muscheler, R., Beer, J., Kubik, P., Synal, H.-A., 2005. Geomagnetic field intensity during the last 60,000 years based on ^{10}Be and ^{36}Cl from Summit ice cores and ^{14}C. Quat. Sci. Rev. 24, 1849–1860. http://dx.doi.org/10.1016/j.quascirev.2005.01.012.
Neugebauer, I., Brauer, A., Dräger, N., Dulski, P., Wulf, S., Plessen, B., Mingram, J., Herzschuh, U., Brande, A., 2012. A Younger Dryas varve chronology from the Rehwiese palaeolake record in NE-Germany. Quat. Sci. Rev. 36, 91–102. http://dx.doi.org/10.1016/j.quascirev.2011.12.010.
Nowaczyk, N.R., Frederichs, T.W., 1999. Geomagnetic events and relative paleointensity variations during the past 300 ka as recorded in Kolbeinsey Ridge sediments, Iceland Sea: indications for a strongly variable geomagnetic field. Int. J. Earth Sci. 88, 116–131.
Nowaczyk, N.R., Antonow, M., Knies, J., Spielhagen, R.F., 2003. Further rock magnetic and chronostratigraphic results on reversal excursions during the last 50 ka as derived from northern high latitudes and discrepancies in precise $AMS^{14}C$ dating. Geophys. J. Int. 155, 1065–1080. http://dx.doi.org/10.1111/j.1365-246X.2003.02115.x.
Nowaczyk, N.R., Arz, H.W., Frank, U., Kind, J., Plessen, B., 2012. Dynamics of the Laschamp geomagnetic excursion from Black Sea sediments. Earth Planet. Sci. Lett. 351-352, 54–69. http://dx.doi.org/10.1016/j.epsl.2012.06.050.
Nowaczyk, N.R., Frank, U., Kind, J., Arz, H.W., 2013. A high-resolution paleointensity stack of the past 14 to 68 ka from Black Sea sediments. Earth Planet. Sci. Lett. 384, 1–16. http://dx.doi.org/10.1016/j.epsl.2013.09.028.
O'Brien, K., 1979. Secular variations in the production of cosmogenic isotopes in the Earth's atmosphere. J. Geophys. Res. Space Phys. 84, 423–431.
Ogg, J.G., Smith, A.G., 2004. The geomagnetic polarity time scale. In: Gradstein, F., Ogg, J., Smith, A. (Eds.), A Geological Timescale 2004. Cambridge University Press, Cambridge, pp. 63–86.
Ojala, A.E.K., Saarinen, T., Salonen, V.P., 2000. Preconditions for the formation of annually laminated lake sediments in southern and central Finland. Boreal Environ. Res. 5, 243–255.
Ojala, A.E.K., Francus, P., Zolitschka, B., Besonen, M., Lamoureux, S.F., 2012. Characteristics of sedimentary varve chronologies – a review. Quat. Sci. Rev. 43, 45–60. http://dx.doi.org/10.1016/j.quascirev.2012.04.006.
Olsson, I.U., 2009. Radiocarbon dating history: early days, questions, and problems met. Radiocarbon 51, 1–43.
Parrenin, F., Petit, J.R., Masson-Delmotte, V., Wolff, E., Basile-Doelsch, I., Jouzel, J., Lipenkov, V., Rasmussen, S.O., Schwander, J., Severi, M., Udisti, R., Veres, D., Vinther, B.M., 2012. Volcanic synchronisation between the EPICA Dome C and Vostok ice cores (Antarctica) 0-145 kyr BP. Clim. Past 8, 1031–1045. http://dx.doi.org/10.5194/cp-8-1031-2012.
Plenier, G., Valet, J.-P., Guérin, G., Lefèvre, J.-C., LeGoff, M., Carter-Stiglitz, B., 2007. Origin and age of the directions recorded during the Laschamp event in the Chaîne des Puys. Earth Planet. Sci. Lett. 259, 414–431. http://dx.doi.org/10.1016/j.epsl.2007.04.039.
Porter, S.C., Swanson, T.W., 2008. 36Cl dating of the classic Pleistocene glacial record in the northeastern Cascade Range, Washington. Am J Sci 308 (2), 130–166.
Povinec, P., Litherland, A., von Reden, K., 2009. Developments in radiocarbon technologies: from the Libby counter to compound-specific AMS analyses. Radiocarbon 51, 45–78.
Putnam, A.E., Schaefer, J.M., Barrell, D.J.A., Vandergoes, M., Denton, G.H., Kaplan, M.R., Finkel, R.C., Schwartz, R., Goehring, B.M., Kelley, S.E., 2010. In situ cosmogenic ^{10}Be production-rate calibration from the Southern Alps, New Zealand. Quat. Geochronol. 5, 392–409. http://dx.doi.org/10.1016/j.quageo.2009.12.001.
Rasmussen, S.O., Andersen, K.K., Svenson, A.M., Steffensen, J.P., Vinther, B.M., Johnsen, S.J., Clausen, H.B., Larsen, L.B., Bigler, M., Röthlisberger, R., Andersen, M.-L.S., Fisher, H., Ruth, U., Goto-Azuma, K., Hansson, M., 2006. A new Greenland ice core chronology for the last glacial termination. J. Geophys. Res. Atmos. 111, D06102. http://dx.doi.org/10.1029/2005JD006079.
Rasmussen, S.O., Bigler, M., Blockley, S.P.E., Blunier, T., Buchardt, S.L., Clausen, H.B., Cvijanovic, I., Dahl-Jensen, D., Johnsen, S.J., Fischer, H., Gkinis, V., Guillevic, M., Hoek, W.Z., Lowe, J.J., Pedro, J., Popp, T., Seierstad, I.K., Steffensen, J.P., Svensson, A.M., Vallelonga, P., Vinther, B.M., Walker, M.J.C., Wheatley, J.J., Winstrup, M., 2014. A stratigraphic framework for naming and robust correlation of abrupt climatic changes during the last glacial period based on three synchronized Greenland ice core records. Quat. Sci. Rev. 106, 14–28.
Reimer, P., Reimer, R., 2001. A marine reservoir correction database and on-line interface. Radiocarbon 43, 461–463.
Reimer, P., Brown, T., Reimer, R., 2004. Discussion: reporting and calibration of post-bomb C-14 data. Radiocarbon 46, 1299–1304.
Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Bronk Ramsey, C., Buck, C.E., Burr, G.S., Edwards, R.L., Friedrich, M., Grootes, P.M., Guilderson, T.P., Hajdas, I., Heaton, T.J., Hogg, A.G., Hughen, K.A., Kaiser, K.F., Kromer, B., McCormac, F.G., Manning, S.W., Reimer, R.W., Richards, D.A., Southon, J.R., Talamo, S., Turney, C.S.M., van der Plicht, J., Weyhenmeyer, C.E., 2009. IntCal09 and marine09 radiocarbon age calibration curves, 0-50,000 years cal BP. Radiocarbon 51, 1111–1150. https://journals.uair.arizona.edu/index.php/radiocarbon/article/view/3569.
Reimer, P.J., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Bronk Ramsey, C., Grootes, P.M., Guilderson, T.P., Haflidason, H., Hajdas, I., Hatté, C., Heaton, T.J., Hoffmann, D.L., Hogg, A.G., Hughen, K.A., Kaiser, K.F., Kromer, B., Manning, S.W., Niu, M., Reimer, R.W., Richards, D.A., Scott, E.M., Southon, J.R., Staff, R.A., Turney, C.S.M., van der Plicht, J., 2013. IntCal13 and Marine13 radiocarbon age calibration curves 0-50,000 years cal BP. Radiocarbon 55, 1869–1887. http://dx.doi.org/10.2458/azu_js_rc.55.16947.
Rhodes, E.J., Bronk Ramsey, C., Outram, Z., Batt, C., Willis, L., Dockrill, S., Bond, J., 2003. Bayesian methods applied to the interpretation of multiple OSL dates: high precision sediment ages from Old Scatness Broch excavations, Shetland Isles. Quat. Sci. Rev. 22, 1231–1244. http://dx.doi.org/10.1016/S0277-3791(03)00046-5.
Rhodes, E.J., 2011. Optically stimulated luminescence dating of sediments over the past 200,000 years. Annu. Rev. Earth Planet. Sci. 39, 461–488. http://dx.doi.org/10.1146/annurev-earth-040610-133425.
Richards, D.A., Dorale, J.A., 2003. U-series chronology and environmental applications of speleothems. In: Bourdon, B., Henderson, G.M., Lundstrom, C.C., Turner, S.P. (Eds.), Uranium-series Geochemistry, Reviews in Mineralogy and Geochemistry, vol. 52. Mineralogical Society of America, Washington D.C., pp. 407–460.
Roberts, A.P., Winklhofer, M., 2004. Why are geomagnetic excursions not always recorded in sediments? Constraints from post-depositional remanent magnetization lock-in modelling. Earth Planet. Sci. Lett. 227, 345–359. http://dx.doi.org/10.1016/j.epsl.2004.07.040.
Roberts, H.M., 2008. The development and application of luminescence dating to loess deposits: a perspective on the past, present and future. Boreas 37, 483–507. http://dx.doi.org/10.1111/j.1502-3885.2008.00057.x.
Rose, J., 2007. The use of time units in Quaternary Sciences Reviews. Quat. Sci. Rev. 26, 1193. http://dx.doi.org/10.1016/j.quascirev.2007.04.002.
Ruff, M., Fahrni, S., Gaggeler, H.W., Hajdas, I., Suter, M., Synal, H.A., Szidat, S., Wacker, L., 2010. On-line radiocarbon measurements of small samples using elemental analyzer and MICADAS gas ion source. Radiocarbon 52, 1645–1656.
Schlolaut, G., Marshall, M.H., Brauer, A., Nakagawa, T., Lamb, H.F., Staff, R.A., Bronk Ramsey, C., Bryant, C.L., Brock, F., Kossler, A., Tarasov, P.E., Yokoyama, Y., Tada, R., Haraguchi, T., 2012. An automated method for varve interpolation and its application to the Late Glacial chronology from Lake Suigetsu, Japan. Quat. Geochronol. 13, 52–69. http://dx.doi.org/10.1016/j.quageo.2012.07.005.
Seierstad, Inger K., Abbott, Peter M., Bigler, Matthias, Blunier, Thomas, Bourne, Anna J., Brook, Edward, Buchardt, Susanne L., Buizert, Christo, Clausen, Henrik B., Cook, Eliza, Dahl-Jensen, Dorthe, Davies, Siwan M., Guillevic, Myriam,

Johnsen, Sigfús J., Pedersen, Desiree S., Popp, Trevor J., Rasmussen, Sune O., Severinghaus, Jeff, Svensson, Anders, Vinther, Bo M., 2014. Consistently dated records from the GRIP, GISP2 and NGRIP ice cores for the past 104 ka. Quat. Sci. Rev. 106, 29–46.

Shen, C.-C., Wu, C.-C., Cheng, H., Edwards, R.L., Hsieh, Y.-T., Gallet, S., Chang, C.-C., Li, T.-Y., Lam, D.D., Kano, A., Hori, M., Spötl, C., 2012. High-precision and high resolution carbonate ^{230}Th dating by MC-ICP-MS with SEM protocols. Geochim. Cosmochim. Acta 99, 71–86. http://dx.doi.org/10.1016/j.gca.2012.09.018.

Smith, V.C., Isaia, R., Pearce, N.J.G., 2011. Tephrostratigraphy and glass compositions of post-15 kyr Campi Flegrei eruptions: implications for eruption history and chronostratigraphic markers. Quat. Sci. Rev. 30, 3638–3660. http://dx.doi.org/10.1016/j.quascirev.2011.07.012.

Southon, J., 2004. A radiocarbon perspective on Greenland ice-core chronologies: can we use ice cores for ^{14}C calibration? Radiocarbon 46, 1239–1259.

Spötl, C., Mattey, D., 2012. Scientific drilling of speleothems – a technical note. Int. J. Speleol. 41, 29–34. http://dx.doi.org/10.5038/1827-806X.41.1.4.

Staff, R.A., Nakagawa, T., Schlolaut, G., Marshall, M.H., Brauer, A., Lamb, H.F., Bronk Ramsey, C., Bryant, C.L., Brock, F., Kitagawa, H., van der Plicht, J., Payne, R.L., Smith, V.C., Mark, D.F., Macleod, A., Blockley, S.P.E., Schwenninger, J.-L., Tarasov, P.E., Haraguchi, T., Gotanda, K., Yonenobu, H., Yokoyama, Y., Suigetsu 2006 Project Members, 2013. The multiple chronological techniques applied to the Lake Suigetsu SG06 sediment core, central Japan. Boreas 42, 259–266. http://dx.doi.org/10.1111/j.1502-3885.2012.00278.x.

Steffensen, J.P., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Fischer, H., Goto-Azuma, K., Hansson, M., Johnsen, S.J., Jouzel, J., Masson-Delmotte, V., Popp, T., Rasmussen, S.O., Röthlisberger, R., Ruth, U., Stauffer, B., Siggaard-Andersen, M.-L., Sveinbjörnsdóttir, A.E., Svensson, A., White, J.W.C., 2008. High-resolution Greenland ice core data show abrupt climate change happens in few years. Science 321, 680–684. http://dx.doi.org/10.1126/science.1157707.

Steinhilber, F., Abreu, J.A., Beer, J., Brunner, I., Christl, M., Fischer, H., Heikkilä, U., Kubik, P.W., Mann, M., McCracken, K.G., Miller, H., Miyahara, H., Oerter, H., Wilhelms, F., 2012. 9,400 years of cosmic radiation and solar activity from ice cores and tree rings. Proc. Natl. Acad. Sci. U. S. A. (PNAS) 109, 5967–5971. http://dx.doi.org/10.1073/pnas.1118965109.

Stuiver, M., Polach, H.A., 1977. Reporting of C-14 data – discussion. Radiocarbon 19, 355–363.

Svensson, A., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Davies, S.M., Johnsen, S.J., Muscheler, R., Parrenin, F., Rasmussen, S.O., Röthlisberger, R., Seierstad, I., Steffensen, J.P., Vinther, B.M., 2008. A 60 000 year Greenland stratigraphic ice core chronology. Clim. Past 4, 47–57. http://dx.doi.org/10.5194/cp-4-47-2008.

Swierczynski, T., Lauterbach, S., Dulski, P., Delgado, J., Merz, B., Brauer, A., 2013. Mid- to Late Holocene flood frequency changes in the northeastern Alps as recorded in varved sediments of Lake Mondsee (Upper Austria). Quat. Sci. Rev. 80, 78–90.

Synal, H.A., Stocker, M., Suter, M., 2007. MICADAS: a new compact radiocarbon AMS system. Nucl. Instrum. Methods Phys. Res. Sect. B Beam Interact. Mater. Atoms 259, 7–13. http://dx.doi.org/10.1016/j.nimb.2007.01.138.

Synal, H.A., Wacker, L., 2010. AMS measurement technique after 30 years: possibilities and limitations of low energy systems. Nucl. Instrum. Methods Phys. Res. Sect. B Beam Interact. Mater. Atoms 268, 701–707. http://dx.doi.org/10.1016/j.nimb.2009.10.009.

Tauxe, L., 1993. Sedimentary records of relative paleointensity of the geomagentic field: theory and practice. Rev. Geophys. 31, 319–354. http://dx.doi.org/10.1029/93RG01771.

Teanby, N., Laj, C., Gubbins, D., Pringle, M., 2002. A detailed palaeointensity and inclination record from drill core SOH1 on Hawaii. Phys. Earth Planet. Inter. 131, 101–140. http://dx.doi.org/10.1016/S0031-9201(02)00032-8.

Thomsen, K.J., Murray, A.S., Jain, M., Bøtter-Jensen, L., 2008. Laboratory fading rates of various luminescence signals from feldspar-rich sediment extracts. Radiat. Meas. 43, 1474–1486. http://dx.doi.org/10.1016/j.radmeas.2008.06.002.

Tomlinson, E.L., Thordarson, T., Müller, W., Thirlwall, M., Menzies, M.A., 2010. Microanalysis of tephra by LA-ICP-MS – strategies, advantages and limitations assessed using the Thorsmork ignimbrite (Southern Iceland). Chem. Geol. 279, 73–89. http://dx.doi.org/10.1016/j.chemgeo.2010.09.013.

Veres, D., Bazin, L., Landais, A., Toyé Mahamadou Kele, H., Lemieux-Dudon, B., Parrenin, F., Martinerie, P., Blayo, E., Blunier, T., Capron, E., Chappellaz, J., Rasmussen, S.O., Severi, M., Svensson, A., Vinther, B., Wolff, E.W., 2013. The Antarctic ice core chronology (AICC2012): an optimized multi-parameter and multi-site dating approach for the last 120 thousand years. Clim. Past 9, 1733–1748. http://dx.doi.org/10.5194/cp-9-1733-2013.

WAIS Divide Project Members, 2013. Onset of deglacial warming in West Antarctica driven by local orbital forcing. Nature 500, 440–444. http://dx.doi.org/10.1038/nature12376.

Walker, M., 2005. Quaternary Dating Methods. John Wiley & Sons Ltd, Chichester.

Wastegård, S., Rasmussen, T.L., Kuijpers, A., Nielsen, T., van Weering, T.C.E., 2006. Composition and origin of ash zones from Marine Isotope Stages 3 and 2 in the North Atlantic. Quat. Sci. Rev. 25, 2409–2419. http://dx.doi.org/10.1016/j.quascirev.2006.03.001.

Wheeler, G.C.W.S., 1988. Optically stimulated phosphorescence and optically transferred TL as a tool for dating. Quat. Sci. Rev. 7, 407–410. http://dx.doi.org/10.1016/0277-3791(88) 90037-6.

Wintle, A.G., 1973. Anomalous fading of thermoluminescence in mineral samples. Nature 245, 143–144. http://dx.doi.org/10.1038/245143a0.

Wintle, A.G., Murray, A.S., 2006. A review of quartz optically stimulated luminescence characteristics and their relevance in single-aliquot regeneration dating protocols. Radiat. Meas. 41, 369–391. http://dx.doi.org/10.1016/j.radmeas.2005.11.001.

Wolff, E.W., 2007. When is the “present”? Quat. Sci. Rev. 26, 3023–3024. http://dx.doi.org/10.1016/j.quascirev.2007.10.008.

Wulf, S., Kraml, M., Brauer, A., Keller, J., Negendank, J.F.W., 2004. Tephrochronology of the 100 ka lacustrine sediment record of Lago Grande di Monticchio (southern Italy). Quat. Int. 122, 7–30. http://dx.doi.org/10.1016/j.quaint.2004.01.028.

Wulf, S., Kraml, M., Keller, J., 2008. Towards a detailed distal tephrostratigraphy in the Central Mediterranean: the last 20,000 yrs record of Lago Grande di Monticchio. J. Volcanol. Geotherm. Res. 177, 118–132. http://dx.doi.org/10.1016/j.jvolgeores.2007.10.009.

Wulf, S., Ott, F., Slowinski, M., Noryskiewicz, A.M., Dräger, N., Martin-Puertas, C., Czymzik, M., Neugebauer, I., Dulski, P., Bourne, A., Blaszkiewicz, M., Brauer, A., 2013. Tracing the Laacher See Tephra in the varved sediment record of the Trzechowskie palaeolake in central Northern Poland. Quat. Sci. Rev. 76, 129–139. http://dx.doi.org/10.1016/j.quascirev.2013.07.010.

Young, N.E., Schaefer, J.M., Briner, J.P., Goehring, B.M., 2013. A ^{10}Be production – rate calibration for the Arctic. J. Quat. Sci. 28, 515–526. http://dx.doi.org/10.1002/jqs.2642.

Zolitschka, B., Brauer, A., Stockhausen, H., Lang, A., Negendank, J.F.W., 2000. An annually dated late Weichselian continental palaeoclimate record from the Eifel, Germany. Geology 28, 783–786. http://dx.doi.org/10.1130/0091-7613(2000)28%3c783:ADLWCP%3e2.0.CO;2.

Zolitschka, B., 2003. Dating based on freshwater and marine laminated sediments. In: Mackay, A., Battarbee, R., Birks, J., Oldfield, F. (Eds.), Global Change in the Holocene. Edward Arnold Publishers, London, pp. 92–106.