High-resolution palaeoclimatology of the last millennium: a review of current status and future prospects

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Abstract: This review of late-Holocene palaeoclimatology represents the results from a PAGES/CLIVAR Intersection Panel meeting that took place in June 2006. The review is in three parts: the principal high-resolution proxy disciplines (trees, corals, ice cores and documentary evidence), emphasizing current issues in their use for climate reconstruction; the various approaches that have been adopted to combine multiple climate proxy records to provide estimates of past annual-to-decadal timescale Northern Hemisphere surface temperatures and other climate variables, such as large-scale circulation indices; and the forcing histories used in climate model simulations of the past millennium. We discuss the need to develop a framework through which current and new approaches to interpreting these proxy data may be rigorously assessed using pseudo-proxies derived from climate model runs, where the 'answer' is known. The article concludes with a list of recommendations. First, more raw proxy data are required from the diverse disciplines and from more locations, as well as replication, for all proxy sources, of the basic raw measurements to improve absolute dating, and to better distinguish the proxy climate signal from noise. Second, more effort is required to improve the understanding of what individual proxies respond to, supported by more site measurements and process studies. These activities should also be mindful of the correlation structure of instrumental data, indicating which adjacent proxy records ought to be in agreement and which not. Third, large-scale climate reconstructions should be attempted using a wide variety of techniques, emphasizing those for which quantified errors can be estimated at specified timescales. Fourth, a greater use of climate model simulations is needed to guide the choice of reconstruction techniques (the pseudo-proxy concept) and possibly help determine where, given limited resources, future sampling should be concentrated.

Key words: Palaeoclimatology, high-resolution, last millennium, tree rings, dendroclimatology, chronology, uncertainty, corals, ice-cores, speleothems, documentary evidence, instrumental records, varves, borehole temperature, marine sediments, composite plus scaling, CPS, climate field reconstruction, CFR, pseudo-proxy approach, time series, climate forcing.

Introduction and rationale

In its Fourth Assessment Report (AR4), Working Group 1 of the Intergovernmental Panel on Climate Change (IPCC, 2007) concluded, with respect to the palaeoclimate record of the last two millennia (Jansen *et al.*, 2007), that: 'Average Northern Hemisphere temperatures during the second half of the 20th century were very likely (> 90% certainty according to IPCC's definition) higher than during any other 50-year period in the last 500 years and likely (> 66% certainty) the highest in at least the past 1300 years'. A similar conclusion was also reached by the US National Academy of Sciences (National Research Council (NRC), 2006). Study of palaeoclimate of the last 1–2 millennia (late Holocene) has undergone dramatic developments in the last 15 years. Up to the early 1990s, there was little cross-disciplinary work and few studies attempted to bring together reconstructions from diverse proxies except at the subcontinental scale. Although there had been earlier qualitative attempts to compare different reconstructions (eg, Williams and Wigley, 1983), the first quantitative extension of the hemispheric-scale instrumental record was produced by Bradley and Jones (1993). This decadal-average curve showed that since about 1930 onwards, summer (June to August) Northern Hemisphere (NH) temperatures were warmer than they had been for any time since 1400. This work was reported in the Second Assessment Report (SAR, IPCC, 1995), replacing the schematic, conceptual temperature history that had been presented in the First Assessment Report (IPCC, 1990: figure 7.1c – see Appendix A for a discussion of the validity and likely source of this figure).

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Although the early results (Bradley and Jones, 1993) were received with considerable scientific interest, they were not allotted much prominence by the SAR, and did not receive much public attention. The availability of several independent or partially independent annually resolved NH average temperature reconstructions (Jones *et al*., 1998; Mann *et al*., 1998, 1999 – hereafter MBH98, MBH99; Briffa, 2000; Crowley and Lowery, 2000), along with the explicit representation of quantitative uncertainty estimates led to the reconstructions having greater prominence in the Third Assessment Report (TAR, IPCC, 2001). The MBH98 reconstruction was also prominently featured in the associated Summary for Policymakers (SPM). The lack of awareness of the 1993 paper and the SAR is evident in many studies which often only contrast the IPCC position in 1990 with that in 2001 (see also Appendix A), though this may not be for purely scientific reasons. Subsequent analyses since the TAR have not substantially changed the interpretation of recent palaeoclimatic data and, if anything, the earlier conclusions have been strengthened (SPM of IPCC, 2007). While the visibility of large-scale (average and spatially detailed) reconstructions stems from their ability to contextualize 'unprecedented' climate change in the twentieth century against a multicentury backdrop, such multiproxy reconstructions are critical to a variety of climate science studies. They provide a large-scale context with which to compare regional climate variability as reconstructed by single proxy records, which may ultimately help resolve the large-scale mechanisms of past low-frequency climate change. They also provide much-needed tests of the response of large-scale climate to a variety of climate forcings which occurred during the last millennium (most notably solar and volcanic forcing).

The purpose of this review is to consider what direction underlying scientific investigations might most profitably take in the immediate future, to reduce existing uncertainties, eg, those inherent in the basic proxy data and the reconstruction of past temperature variability, and to provide greater insight into the factors governing past climate change. More work is needed to better understand and more comprehensively quantify the sources of uncertainty in various climate proxy archives and errors in reconstructions, and then to improve them. However, we first need to understand the reasons for the differences among existing climate reconstructions which make use of different types or combinations of climate proxy data and different statistical methods to combine these data within a climate reconstruction. There is, as always, a clear need for more local and regional reconstructions from diverse proxies in as many parts of the world as possible. Climate model simulations can also play an important role here, acting as a surrogate climate, whose variability through time is perfectly known, and with which we may test the performance of alternative reconstruction approaches. Such tests, however, are tied to the underlying assumptions regarding the error structure of the climate proxy data.

To make significant progress there is a parallel need to improve our understanding of the nature of the processes, both climatic and non-climatic, that influence climate proxy data, and the need to recognize and account for these intrinsic limitations. It is also widely recognized that future work ought to extend beyond the reconstruction of simple hemispheric average temperature series and important large-scale circulation indices. Instead, further investigations should also attempt to resolve seasonally specific variations and identify the associated patterns of temperature, precipitation and circulation variability, perhaps examining specific anomalous periods in the past (selected from analysis of climate proxy observations, estimated forcing histories or model simulation results). This paper represents the outcomes of discussions at Wengen, Switzerland in June 2006 under the auspices of PAGES/CLIVAR held in an attempt to define how progress could be made on these issues.

Background

The purpose of the Wengen workshop was to synthesize the current state of late-Holocene climate reconstruction efforts, to assess the approaches to data–model comparisons and to elaborate on what possible or likely advances are expected in the coming years. The workshop discussions were organized into three principal subject areas:

(1) Proxy data availability and reliability
(2) Large-scale/regional reconstruction approaches and their uncertainties, and how these can be informed by climate modelling (from simple Energy Balance Models (EBMs) to fully coupled Atmosphere/Ocean General Circulation models (A/OGCMs), even including atmospheric chemistry)
(3) Factors that influence/force the climate system (both natural and anthropogenic external factors as well as internal variability) and how these are treated within climate models.

This review is an extensive and critical examination based on these discussions and is organized in a similar framework. A brief summary from the meeting has already been published (Mann *et al*., 2006). The section 'Proxy data uncertainty' below discusses aspects of the use of various types of proxy data, with a specific focus on their current availability and uncertainty. Although structured by discipline, each subsection addresses the most important issues: reducing uncertainties (improving both reconstruction reliability and, for less-than-annually resolved proxies, improving dating accuracy and resolution) and making best and full use of instrumental records for calibration and verification (particularly important for decadally to annually resolved proxies) of reconstructed climatic parameters. The issues discussed in this section differ in focus and detail, in part reflecting the different maturity of each discipline. Some weight and a corresponding amount of text is allotted to the dendroclimatic issues. The justification for this is the generally large proportion of tree-ring based proxies used in many of the large-scale reconstructions, but also because it is felt that much of the discussion on interpretational limitations of these data has relevance for the increasing use of other forms of high-resolution proxy data, particularly as they become available over increasingly longer periods of time.

The section 'Combining proxies to reconstruct large-scale patterns, continental and hemispheric averages' discusses the various approaches that have been developed to combine currently available proxy series into large-scale (continental to hemispheric) averages and to reconstruct internally consistent climate fields, representative of seasonal or annual conditions during the past. This section also describes the application of experiments using pseudo-proxies (derived from GCM output) to provide benchmark tests for assessing the performance of statistical reconstruction techniques. The following section, 'Climate forcing and histories' reviews the development of past climate forcing histories, discussing which are the most important for millennial-scale climate integrations, how different models implement the forcings, and discusses uncertainties in forcing histories. The final section concludes with a summary of major findings and a comprehensive set of recommendations for future work.

Proxy data uncertainty

Tree rings and the need for improved regional and temporal coverage

Tree-ring-derived records have played a prominent role in attempts to establish how climate has varied in the recent past. Networks of climatically sensitive tree-ring chronologies have long been used to reconstruct detailed spatial patterns of interannual climate variability on regional and near-hemispheric scales, typically extending observed climate records by several centuries (Fritts, 1991; Schweingruber *et al*., 1991; Briffa *et al*., 1994, 2002b; Cook *et al*., 2004b, 2007). Several chronologies extending over a longer time span, with variability displaying a strong and direct association with changing local temperatures, have been utilized in virtually all published studies aimed at reconstructing Northern Hemisphere (NH) or global average surface temperature changes during the millennium leading up to the present (Jansen *et al*., 2007).

We do not attempt here to discuss in detail the well-known positive attributes of dendroclimatology. Reviews of the scope and general strengths of different types of tree-ring data (that they are continuous, precisely defined with annual resolution or better, accurately dated on a calendar timescale, widely distributed and rigorously calibrated against observed climate data) are already well described in a number of books and general articles (Fritts, 1976b; Cook and Kairiukstis, 1990; Briffa, 1995; Schweingruber, 1996; Treydte *et al*., 2001, 2006; Hughes, 2002; McCarroll and Loader, 2004; Luckman, 2007) and the published proceedings of international conferences (eg, Hughes *et al*., 1982; Fritts and Swetnam, 1989; Bartholin *et al*., 1992; Dean *et al*., 1996). However, the continuing advancement of dendroclimatology as a discipline is not based solely on the exploitation of these strengths. It also involves an explicit appreciation of limitations or weaknesses. Hence, our purposes in this review article are best served by drawing attention to some of the general lessons learned in tree-ring research, with a focus on the shortcomings of this proxy:

those that are innate characteristics of tree-ring data themselves, but also those that may be identified in current dendroclimatic practise (Hughes, 2002; Esper *et al*., 2007b). The issues discussed do not relate solely to the most commonly used tree-ring 'width' data *per se*, but apply to all forms of tree-ring-derived proxy data, including densitometric (eg, Schweingruber, 1996), chemical or stable isotope (McCarroll and Loader, 2004; Treydte *et al*., 2007) data. The following discussion draws attention to some selected aspects of tree-ring data and dendroclimatology that are relevant to the continuing efforts to better understand late-Holocene climate variability. We first discuss the aspects of temporal and spatial coverage of tree-ring data, with a focus on the status of dendroclimatic studies in the tropics and the Southern Hemisphere. We then describe a number of issues relating to potential improvements in statistical methods used to assemble long tree-ring chronologies and how their statistical quality might be better represented. Finally, we draw attention to problems in the way chronologies are generally interpreted in terms of specific climate parameters. The points we raise about the need for improved ways to measure chronology confidence and possible limitations in the reconstruction of longer-timescale climate variability are described at some length here because we believe them to be equally relevant to the analysis and interpretation of other proxy records discussed in the following sections of this review.

Knowledge of the expanding geographic coverage and recent regional developments in dendroclimatology may be gleaned from various reviews (Hughes *et al*., 1982; Dean *et al.*, 1996; Hughes, 2002; Luckman, 2007). Here we focus on the most recent developments in those regions highlighted in the AR4, as virtually devoid of tree-ring data, specifically the tropics and the Southern Hemisphere (SH), though we include discussion of the state of 'long-chronology' development in these and other regions of the world.

Prospects for tropical dendroclimatology

Tropical dendrochronology was long considered impractical because the growth periodicity of most tropical tree species is seldom clearly and unambiguously defined (eg, Jacoby, 1989; Gourlay, 1995; Vetter and Wimmer, 1999; Worbes, 1995). However, this has turned out not to be entirely true, as seasonally dry regions have produced tree-ring records from tropical Asia (eg, Berlage, 1931; Buckley *et al*., 1995, 2005, 2007a, b; D'Arrigo *et al*., 1994, 1997, 2006a; Pumijumnong *et al*., 1995; Sano *et al*., 2008), from Africa (eg, Stahle *et al*., 1999; Tarhule and Hughes, 2002; Therrell *et al*., 2006) and from the American tropics (eg, Biondi, 2001; Schöngart *et al.*, 2004a, b; Therrel *et al*., 2004; Biondi *et al*., 2005; Brienen and Zuidema, 2005, 2006; D'Arrigo and Smerdon, 2008). Furthermore, the notion that tropical regions that lack clear seasonality pose an insurmountable problem has turned out not to be the case. Evans and Schrag (2004), Poussart *et al*. (2004) and Poussart and Schrag (2005) are among the first to have applied improved methods of stable isotope geochemistry that show the possibility of using many apparently 'ringless' species for dendroclimatic studies in tropical environments. Evans and Schrag (2004) demonstrated the application of these methods on species from Costa Rica, while Poussart *et al*. (2004) and Poussart and Schrag (2005) applied them to Thai and Indonesian trees. In all cases clear periodicity in $\delta^{18}O$ and $\delta^{13}C$ was demonstrated and determined to be annual, and relationships to rainfall were established. Poussart *et al*. (2006) also applied x-ray microprobe synchrotron analysis to a ringless species from Thailand, in an attempt to more easily define annual periodicity through chemical cycles of calcium, and this too holds promise for wider application.

In spite of the successes, however, formidable obstacles still restrict the development of tropical dendroclimatology, including the scarcity of suitable tree species with easily identifiable and measurable annual rings, and continued pressure on forest resources due to logging and disturbance that limits the availability of old-growth trees. A veritable absence of information regarding the ecophysiology and phenology of tree species further exacerbates problems of working in the tropics (Borchert, 1995). Even with the use of isotopic time series the problem of crossdating is still severe, given the nature of locally absent banding that is often severe in many tropical trees. Furthermore, the time and costs associated with isotopic geochemistry currently inhibit widespread application.

Temporal control remains an issue with tropical tree-ring analyses under many situations, and radiocarbon (^{14}C) measurements have been used to assess the annual nature of growth rings through detection of the radiocarbon 'bomb spike' (eg, Biondi and Fessenden, 1999; Hua *et al*., 1999) and to estimate the age and average growth rates of some tropical trees (eg, Poussart and Schrag, 2005). However, species-specific effects can limit the application of ^{14}C dating because of uncertainties associated with soil respiration, internal carbohydrate transfer and other ecophysiological factors (Worbes and Junk, 1989). Dendrometer studies have been employed (eg, Buckley *et al*., 2001; DaSilva *et al*., 2002), and these provide a useful alternative, along with cambium-wounding or 'pinning' methods that give a reference for growth from a time of known scarring of the cambium (eg, Mariaux, 1967; Nobuchi *et al*., 1995). Analysis of tropical wood chemistry may also reveal seasonal signatures of cambium activity, as exemplified by Poussart *et al*. (2004) in Indonesia, and Poussart and Schrag (2005) in Thailand, both of whom demonstrate that the generation of replicated subannual $\delta^{18}O$ and $\delta^{13}C$ records from ringless tropical trees is possible over several decades.

More traditional approaches to dendroclimatology in the tropics still have their place. Buckley *et al*. (2007b) produced an inferred reconstruction of Palmer Drought Severity Index (PDSI), a measure of soil moisture availability (Palmer, 1965), based on ring-widths of teak from northwestern Thailand that extended back nearly 500 years. This record illuminated periods of decadal-scale drought, the most significant of which persisted from 1690 to 1720 and 1735 to 1765, respectively, and coincided with periods of extreme social unrest. Sano *et al.* (2008) produced a 535-yr ring-width record from living *Fokienia hodginsii* (Dunn, A. Henry and H.H. Thomas) of the family Cupressaceae. The authors used this record to reconstruct PDSI for the pre-monsoon period of March to May over the past 500 years for northern Vietnam. Significantly this is the first record from the region that successfully calibrated and verified the reconstruction statistically, using the available instrumental records. When compared with the Buckley *et al*. (2007b) teak record from Thailand, a similar drought from 1750 to 1780 is revealed, suggestive of a 'megadrought' that may have extended from Burma to Vietnam. This period coincides with the outright collapse of all the major kingdoms in Southeast Asia (Leiberman, 2003), pointing to a possible direct climatic impact on the societies of the region.

In some instances it has been shown that the response of tropical tree species to climate is not as straightforward as for temperate regions. Rivera *et al.* (2002) showed that for a number of species the influence of subtle changes in photoperiodicity in the American and Asian tropics is more important than rainfall for producing annual flushing of leaves in shade-sensitive species. Buckley *et al.* (2007a) illustrated an apparent inverse response to drought in *Pinus merkusii* across southeast Asia, possibly linked to reduction of photosynthetically active radiation during times of intense convective rainfall. Saleska *et al*. (2007) found that Amazon forests significantly increased their biomass in response to a prolonged drought in 2005, counter to model predictions of forest collapse in response to drought (eg, Betts *et al*., 2004). At the peak of the 2005 Amazon drought, the authors found a significant increase in canopy 'greenness', indicating an ecological and physiological vegetation response that is opposite to a presumption that anomalously low rainfall would negatively influence these forests. Clark *et al*. (2003) found a negative relationship between minimum temperature and tropical tree growth at La Selva in Costa Rica, but no relationship with interannual variations in precipitation (Clark and Clark, 1994). These findings have implications for the interpretation of isotopic time series from tropical trees, illustrating the need for research into the physiological response of tropical forest species to climate, and a greater understanding of their anatomy.

Furthermore, the research of both Feely *et al*. (2007) and Clark *et al*. (2003) suggest a 'deceleration' in growth rates of tropical forests in Costa Rica, Panama and Malaysia, in opposition to some predictions that pan-tropical forests could experience increased growth induced by CO_2 fertilization or increases in water use efficiency (eg, Lewis *et al*., 2006). There continues to be much debate about the response of tropical forests to climate change and fundamental questions remain about whether or not tropical forests serve as carbon sink or source, and the extent to which any changes in biomass can be attributed to anthropogenic change (eg, Wright, 2006). Dendrochronology can help to answer many of these questions if it can be more widely applied, and the developments so far are encouraging. As more species are analysed for their eco-physiological characteristics and their response to climate, more old-growth species are discovered and improvements are made to geochemical methodologies, the prospects for widespread application of dendrochronology in the tropics look promising.

Recent developments in South America and southern Africa

For the interval 1890–2000, tree-ring width chronologies (from *Swietenia macrophylla* and *Cedrela odorata*) were developed by Dünisch *et al*. (2003) in Mato Grosso, Brazil (10°09′S, 59°26′W). Correlation analyses revealed a significant relationship between seasonal precipitation and the growth of both species. Several month-long inundation indices influence the formation of annual rings in trees growing in the seasonally flooded Amazon plains (3–4°S, 65°W). Ring widths are inversely related to duration of the flood. A 200-yr long Euphorbiaceae chronology (*Piranhea trifoliata*) has been used to estimate the length of the vegetation season, which is, in turn, related to ENSO events (Schöngart *et al*., 2004a, b). The results indicate that during the last two centuries, the severity of ENSO events in the Amazon basin has significantly increased.

In a related study, Schöngart *et al*. (2004b) developed tree-ring chronologies from *Macrolobium acaciifolium* in two different floodplains in Central Amazonia. Maximum tree age in the nutrient-poor 'black-water' was more than 500 years, contrary to the nutrient-rich 'white-water' floodplain, where ages are not older than 200 years. Ring-width variations in both floodplain forests were significantly correlated with the length of the vegetation period derived from the daily recorded water level at the port of Manaus since 1903. Both chronologies showed increased wood growth during El Niño events associated with negative precipitation anomalies and lower water discharge in Amazonian rivers.

Exploratory work has established the dendrochronological potential of several tropical lowland species in the Bolivian sector of the Amazon basin (11–15°S, 66–68°W). Brienen and Zuidema (2005) crossdated six rainforest species and established the influence of annual and seasonal rainfall on radial growth. Ongoing work in Argentina and Brazil has also identified several species typical of the dry tropical forest of southern Bolivia–northern Argentina (16–24°S) and southeastern Brazil (22–25°S) that show clear rings and potential for dendroclimatic studies.

A major advance in the effort to expand the spatial coverage of tree-ring records across the Americas has been the recent development of *Polylepis tarapacana* chronologies in the Bolivian Altiplano (Argollo *et al*., 2004). These records located between 16 and 22°S and above 4500 m elevation, are the closest to the Equator in the Andes and the highest-elevation chronologies in the world. Most *Polylepis* records cover the past three to four centuries, but some extend over seven centuries (Argollo *et al*., 2004; Soliz *et al*., 2008). Examination of interannual variations in ring width and climate in the Altiplano indicate that the growth of *Polylepis* is associated with summer water balance (Morales *et al*., 2004). In addition, *P. tarapacana* chronologies from the south-central tropical Andes provide high-resolution records that are extremely sensitive to ENSO in the tropical Pacific, and represent an important component to be considered in future multiproxy ENSO reconstructions (Christie *et al*., 2008).

A major requirement in Southern African dendroclimatology is the need to examine the many tropical and subtropical species to assess their dendrochronological potential (see the pioneer work of Lilly, 1977 and February, 1996). Stahle *et al*. (1997) have begun surveying the diverse indigenous forests in tropical Africa for species suitable for dendroclimatology. Two tropical chronologies of *Pterocarpus angolensis* from Zimbabwe are strongly correlated with total rainfall amounts during the wet season. Both chronologies reach back only to 1870 at present, but *P. angolensis* is the most important timber species in south tropical Africa, and old samples survive in buildings and other diverse sources that may permit the eventual development of 200–300 yr chronologies in southeastern Africa (Stahle *et al*., 1997).

Developing more 'long' tree-ring-based chronologies

There are very few tree-ring chronologies from around the globe that extend back 1000 years. This is very apparent in figure 6.11 of Jansen *et al*. (2007), which shows only 16 locations globally from which tree-ring data have been used in large-scale temperature reconstructions to date. Of these, three are in the SH. The remainder are virtually all restricted to the western edge of North America or the high latitudes of Eurasia, with the exceptions being two sites adjacent to the Mediterranean and one in Mongolia. The contribution of dendroclimatology to improved late-Holocene climate reconstruction must involve a geographic expansion of work developing long composite chronologies. Many areas, some with a proven history of (admittedly short-timescale) tree-ring research and demonstrably climatically sensitive trees, have yet to be systematically investigated for their potential to produce subfossil wood and hence much longer chronologies. More long records are needed in the northern mid latitudes though work is ongoing in Europe (Nicolussi and Schiessling, 2001; Grabner *et al*., 2001; Helama *et al*., 2005; Büntgen *et al*., 2006, 2008; Popa and Kern, 2008), N Africa (Esper *et al*., 2007a), N America (Barclay *et al*., 1999; Buckley *et al*., 2004; Luckman and Wilson, 2005) and Asia (Sidorova *et al*., 2006; Esper *et al*., 2007c).

Many more long chronologies are needed in the SH. Significant recent achievements in South America include the development of a 5666-yr-long composite chronology from *Fitzroya cupressoides*, currently the longest continuous chronology in the SH (Wolodarsky-Franke, 2002). Chronologies of this species (from Argentina and Chile) were used in the 1990s to develop the first millennium-long temperature reconstructions for South America. However, recent work suggests a lack of stability in the relationship between the low-frequency component of climate and *F. cupressoides* tree growth, particularly during recent decades. Studies using *Austrocedrus chilensis* have developed several new records. A 1864-yr-long chronology from northern Argentinean Patagonia is a major component of a preliminary SH ENSO reconstruction for the last 1300 years and collaborative studies between tree-ring laboratories in Chile, Argentina and USA have developed 800-yr-long precipitation reconstructions for central Chile (LeQuesne *et al*., 2006, 2008).

Development of millennia-long tree-ring chronologies from Australia and New Zealand has been based on three tree species:

Lagarostrobos franklinii from Tasmania, *Lagarostrobos colensoi* from the South Island of New Zealand and *Agathis australis* from the North Island of New Zealand. These are very long-lived species and provide abundant quantities of well-preserved wood. The *L. franklinii* record from a high-elevation site on Mount Read, Tasmania presently covers 4136 years (Cook *et al*., 2006). It contains a reliable warm-season temperature signal and its variability is significantly correlated with SSTs in the southern Indian Ocean. The *L. colensoi* record comes from low-elevation sites on the west coast of the South Island and is now 2327 years long. It too represents warm-season temperatures and correlates well with Tasman Sea SSTs. Finally, the *A. australis* record comes from mostly low-elevation sites on the North Island and now spans the past 3722 years. It contains a robust ENSO signal, but has not yet been used for climate reconstruction purposes (Cook *et al*., 2006).

Improved tree-chronology-production methods, measuring chronology confidence and statistical limitations

Despite the development of wider networks of various dendroclimatic proxies it is important to appreciate the essential requirement for continuing effort to improve the number and quality of current dendroclimatic resources, even in areas that currently possess widely utilized long chronologies. There remain major issues concerning statistical and interpretational confidence associated with building and interpreting existing long tree-ring chronologies. ‘Standardization’ of tree-ring data is required to remove biases in chronologies that arise during periods where the make-up of samples is dominated by a concentration of measurements from either relatively young-age trees (that typically have wider rings and higher density) or old-age trees (with narrow rings and lower density). Approaches used previously to overcome these potential biases tended to operate as high-pass filters of the data and resulting chronologies contained little, if any, evidence of climate variation on timescales longer than the average length of the sample trees, perhaps only on centennial timescales or even less (Cook *et al.*, 1995; Briffa *et al.*, 1996). In recent years new methods, or new applications of old methods, used in different contexts, have been proposed specifically for the processing of tree-ring data intended for long-period climate reconstruction work (Erlandsson, 1936; Becker, 1989; Briffa *et al*., 1992; Dupouey *et al*., 1992). These methods are generally referred to under the generic title ‘Regional Curve Standardization’ or RCS. RCS methods have the potential to preserve more low-frequency climate information. It is important to recognize that the chronologies used in many recent NH temperature reconstructions representing the last millennium were produced, either using only tree-ring data processed with various realizations of these RCS techniques (Briffa, 2000; Esper *et al.*, 2002; Cook *et al*., 2004a; D’Arrigo *et al*., 2006b) or, alternatively, included at least some tree-ring data produced with them (Mann *et al*., 1999; Mann and Jones, 2003; Hegerl *et al*., 2006).

Most long tree-ring chronologies must be constructed using a combination of data extracted from living, recently dead and somewhat older preserved subfossil wood. Processing of this material, using the RCS techniques, includes assumptions about the homogeneity of sample data through time (ie, that changes in the apparent average growth rates of trees in one region result solely from changes in local climate forcing and not because of a lack of homogeneity in the origins of the sample trees, whose average growth rates might reflect differences in elevation, aspect, soil type or competition pressures). These methods also require large replication of samples and that the samples together span a long period of time. In practical situations where these methods have been used, these requirements are unlikely to be entirely satisfied. So while the chronologies do exhibit greater variability on longer timescales than is apparent with earlier processing techniques (Briffa *et al*., 1996; Cook *et al*., 2006; D’Arrigo *et al.*, 2006b), this long-period variance may be associated with wide confidence limits. Hence, the possible improvement in preserved low-frequency variance must be viewed in the context of realistically assessed, and potentially large, uncertainty.

It is also known that various implementations of the basic RCS approach (as for other standardization approaches), at least in the way that they have been implemented until now, can impart systematic biases near the beginnings and ends of tree-ring chronologies (Cook and Peters, 1997; Briffa and Melvin, 2008; Melvin and Briffa, 2008). Recent statistical processing techniques go some way to correct for spurious variance inflation where sample replication is low, or where ring-width measurements are extremely small (Cook and Peters, 1997; Osborn *et al*., 1997; Frank *et al.*, 2007). Work is also continuing to explore other biases in the use of the RCS technique, but this work must be expanded to take in a wider range of geographical and ecological situations and to widen the scope of RCS application to situations where sample replication is limited, the data are known to be inhomogeneous or where they cover only a relatively short (often recent) period (Briffa and Melvin, 2008).

In the meantime, chronologies, even some based on similar measurement data from the same locations, can exhibit very different statistical properties because they were constructed using different techniques, or different implementations of similar techniques (eg, compare Hantemirov and Shiyatov, 2002 with Briffa, 2000; Naurzbaev *et al*., 2002 with Jacoby *et al.*, 2000 or Esper *et al*., 2002). This can cause confusion and may lead to a degree of arbitrariness in the choice of seemingly similar published data for inclusion in different hemispheric or global reconstruction efforts. This situation will also be improved if future work is concentrated on the provision of demonstrably robust regional chronologies and associated regional climate reconstructions.

Dendroclimatology has always paid attention to the important issue of quantifying chronology confidence, originally by calculating generalized statistics such as overall chronology standard error and through analysis of variance which was used to provide a measure of the strength of the common forcing inherent among the series of tree-ring indices that make up a chronology (Fritts, 1976b; Cook and Kairiukstis, 1990). Subsequent work demonstrated how a simple measure of the growth-forcing signal could be conveniently estimated (as the average of all correlation coefficients, $\bar{r}$, calculated from the multiple inter-series comparisons of standardized data) for different chronologies (Wigley *et al.*, 1984). More importantly, it was shown how $\bar{r}$ could be used to measure the time-dependent statistical quality of a chronology, in terms of its similarity to a hypothetical infinitely replicated series (ie, one with no expressed noise and where the variance of the underlying growth-forcing process is perfectly represented: the so-called expressed population signal, EPS). This measure of chronology quality has been widely adopted in dendroclimatic studies, to gauge the period over which a chronology can be considered to be of ‘acceptable’ statistical confidence, though a somewhat *ad hoc* EPS threshold criterion is often used for defining acceptance (Wigley *et al.*, 1984). However, what is perhaps not widely appreciated is that this use of EPS is often biased toward the measurement of short-timescale chronology confidence, because it is interannual variance that dominates the routine calculation of $\bar{r}$, particularly when this is done on the basis of a relatively short-period of common overlap between constitute chronology index series. So, while a chronology may be reasonably considered statistically robust in terms of its representation of relative year-to-year, or even decade-to-decade variability, assuming that high EPS values, as they are generally calculated, are a reliable measure of longer-timescale variance is questionable.

The common variability between low-pass-filtered index series within a chronology (eg, representing timescales >100 years) can be substantially less than that measured at medium-frequency (~ multidecadal timescales), and the common variability at long and medium timescales is invariably lower than that measured at the interannual (< 10 years) timescale. Early work aimed at assessing the level of sample replication required to achieve statistically robust chronologies for dendroclimatic studies was carried out in the southwest of the USA. Tree growth in this region is strongly influenced by a lack of available moisture and inter-tree ring-width variability displays strong common patterns, with large year-to-year variations. Measurements of common variance within these data led to the conclusion that ~20 trees were sufficient to provide a robust chronology (DeWitt and Ames, 1978). This estimate is unlikely to be generally valid where the focus is on relatively longer timescales than the short-period variability that dominates the growth patterns of drought sensitive conifers in the American southwest. In many other regions of the world, where the climate signal in tree-ring data is typically smaller, data from more trees are required to represent, even high-frequency, variability reliably and perhaps an order of magnitude more sample series will be required to produce a robust representation of long-timescale variance (see Cook and Kairiukstis, 1990). Many tree-ring chronologies, including some that make up well-known spatially expansive networks in parts of North America, Europe and Russia are not sufficiently internally replicated to ensure reliable individual representation of relatively low-frequency common forcing (ie, on timescales longer than several decades). Typically these networks are made of chronologies that span several centuries. However, for this time, intraregional similarities in chronology variability can be used to demonstrate the likely fidelity of inferred regional growth-forcing signals (Schweingruber and Briffa, 1996; Esper *et al*., 2002; D'Arrigo *et al*., 2006b; Esper *et al*., 2007c).

The value of large series replication in dendroclimatic reconstruction of centennial and longer timescales cannot be overstated. All proxy disciplines benefit from an appreciation of the need to provide multiple measurement series both within and between adjacent sampling locations. Reliable estimates of inherent time-dependent and timescale-dependent growth-forcing signals, and subsequent quantification of proxy series reliability are only possible where multiple sampling is routinely undertaken. Certainly, within dendroclimatology, even greater attention needs to be paid to the explicit quantification of chronology confidence for specific timescales of variability. Empirical, Monte Carlo-based approaches to assessing chronology confidence, such as by bootstrap techniques, applied at individual site and wider inter-regional spatial scales, could be usefully adopted much more widely (Cook, 1990; Esper *et al*., 2002; Cook *et al*., 2004a; Guiot *et al.*, 2005).

This brings us back to the particular issue of establishing the realism of low-frequency, regionally coherent, climate signals in existing high-resolution tree-ring series. In the same way that the parallel behaviour of different sample series within a localized chronology can be used explicitly to demonstrate the existence, and measure the strength, of a common climate signal, it is important to establish that different chronologies constructed within the same region display common features over relatively long, multi-centennial periods (Esper *et al*., 2002).

Instrumental climate records have been used to estimate the distances over which surface temperatures would be expected to co-vary in different parts of the globe (Briffa and Jones, 1993; Jones and Briffa, 1996). This covariance changes according to season and timescale: it is greater over tropical oceans compared with high-latitude land areas. It is greater in winter rather than summer and appears to be greater on decadal compared with interannual timescales. These results show that within 'reasonable' separation distances, different chronologies would be expected to exhibit common variability. At multidecadal, century and longer timescales, evidence of mutual variability is a minimum prerequisite for establishing whether or not chronologies provide genuine evidence of regional climate signals. There is as yet not enough evidence to demonstrate that the few relatively long chronologies that are generally incorporated within numerous different reconstruction studies robustly represent the low-frequency climate forcing of trees within their source regions. In the light of this uncertainty, there is, therefore, a requirement, not only for sampling 'new' areas, but also for substantially greater effort to provide a much improved density of samples within these critical regions, and over long periods, where there is already a proven potential for tree-ring data to contribute usefully to multimillennial-length, average hemispheric and global reconstruction efforts.

The various methods by which proxy climate data are transformed into estimates of past climate variability, on different time and space scales are discussed in the following section. However, several issues that arise in the context of interpreting tree-ring data are worthy of separate mention here.

The first is, in part, associated with possibly limited expression of long-timescale climate forcing discussed above: the potential biasing of regression weights. When tree-ring data are scaled by comparison with instrumental climate data there is generally no explicit consideration of whether or not the scaling should be equal for all timescales. Calculation of the optimum scaling factor takes account of the match in total variance, with no separate assessment made of the fit between growth and climate variability on short and long timescales. Tree-ring records from different tree species have very different variance spectra in terms of the ratio of short- to long-timescale variance, irrespective of the standardization procedure. For example, in northern Eurasia, pine species will display relatively low high-frequency variance (a redder spectrum) while larch series have a much greater proportion of year-to-year variability (bluer spectrum). Similarly, ring-width series in conifers have a redder spectrum than ring densities. Simple linear scaling or least squares regression will be strongly influenced by the year-to-year variability of the data, the more so where the comparison (calibration) involves data with little or no trend. Where the comparison period is short, the regression coefficients, besides being subject to large uncertainty, are invariably biased towards representing growth processes operating on year-to-year rather than century or longer timescales (Osborn and Briffa, 2004).

Opportunities to explore whether a single scaling is 'optimum' for different timescales have been limited by a general lack of long instrumental climate records, but what little work that has been done to test this assumption suggests that it may not be appropriate (Guiot, 1985; Osborn and Briffa, 2000). Whether this is a reflection of a genuine difference in the temperature control of tree-growth processes on different timescales, or merely a product of statistical sampling error is not known. What is known is that regressing climate data against the same tree-growth data, alternatively represented at annual resolution or after common low-pass smoothing of the climate and growth data, can result in very different scaling factors and, hence, reconstructed climate series with notably different long-term variability (Esper *et al*., 2005). This further complicates the comparison of dendroclimatic reconstructions where different scaling approaches have been used (Jansen *et al*., 2007).

In dendroclimatology, a long-recognized requirement to 'verify' the temporal stability of regression coefficients and to establish that multiple regression equations have not been overfitted, has led to the routine subdivision of climate series to allow the calculation of a suite of goodness-of-fit statistics, comparing climate estimates with independent observations (Cook *et al*., 1994). Typically, the period of overlap between tree-ring and climate data is split either in half or in the ratio two-thirds to one-third (eg, Neter *et al.*, 1990). The data in one period are used to derive the optimum scaling of the tree-ring data and the regression estimates are compared with the climate observations over the other period. In the case of multiple regression, this comparison provides more realistic estimates of the reconstruction uncertainty than provided by the calibration-period measures alone (Fritts, 1976a). Repeating the process with the periods reversed allows direct comparison of the time-stability of regression coefficients and the reconstruction calculated ranges. As this cross-calibration approach requires that the climate data be subdivided into even shorter series, the calculation of regression coefficients and the assessment of reconstruction quality are unavoidably strongly biased toward short-timescale variability. In an attempt to mitigate this bias, it has been common practice to repeat the calibration of the ultimate regression weights to be used for reconstruction, but using all of the available climate data (eg, Briffa *et al*., 1992). In such situations, however, no formal statistical verification of the quality of the long-timescale variance estimates is possible. Measurements of the similarity between estimated and observed climate variability at different timescales made over the full overlap period often reveal a higher association on interannual rather than on decadal and longer timescales. In future work, it would be desirable if reconstructions were more commonly presented with timescale-specific confidence limits: ideally in a way that indicates predictor (chronology) error and calibration error, separately and together. Where climate estimates are extrapolations beyond the range of the variance incorporated within the data used for calibration (or more strictly, those used for independent reconstruction verification) this should also be clearly indicated.

Recent divergence between tree-ring growth and temperature

The final aspect of tree-ring studies that needs to be highlighted is what has become known as the 'divergence' issue. This refers to the apparent failure of some (established as temperature-responsive) tree-ring data to follow the trend in instrumental temperatures observed over the latter part of the twentieth century. Chronology time series that vary largely in parallel with changing temperature in earlier periods progressively fail to show the increasing trends that would represent a continuing positive response to the strong warming observed during recent decades. Originally this was noted primarily in certain northern high-latitude areas for ring-width data in Alaska (Jacoby and D'Arrigo, 1995) and in ring-width and particularly ring-density data, in more extensive regions of northern Europe and Russia (Briffa *et al.*, 1998). In the earlier work, it was suggested that the cause of the North American observations was a shift from a direct dominant temperature control on tree growth to one where lack of available moisture becomes increasingly influential, possibly to an extent where the sign of the temperature influence becomes negative rather than positive (Jacoby and D'Arrigo, 1995; D'Arrigo *et al*., 2004).

Subsequently, various studies focused mainly on recent tree-growth in Alaska and Canada support the idea that current tree growth may no longer be responding positively to increased warming (Barber *et al.*, 2000; Lloyd and Fastie, 2002; Davi *et al*., 2003; Wilmking *et al*., 2004; Driscoll *et al*., 2005; Pisaric *et al.*, 2007). Other suggestions have been offered as the cause of the widely observed loss of temperature response over northern Eurasia. The increasing influence of drought has also been suggested as the cause (Jacoby *et al*., 2000), though other suggestions include possible reduced atmospheric clarity, localized persistence of spring snow cover and seasonal changes in ozone-related surface UV concentrations (Briffa *et al.*, 1998, 2004; Vaganov *et al.*, 1999; D'Arrigo *et al*., 2008).

The IPCC recently laid particular stress on this issue, pointing out that any significant shift in the recent growth response of trees would invalidate the assumptions that underlie the simple regression-based approach to reconstructing past temperature changes. This would imply an inability to recognize potential underestimates of the degree of warmth in earlier periods of reconstructions (Jansen *et al*., 2007). It is important to stress that not all high-latitude regions display this apparent decoupling between observed and dendroclimatically estimated temperatures (Briffa *et al*., 2007; Wilson *et al*., 2007). However, the issue remains a crucial one. Unfortunately, a comparative scarcity of recent (ie, post-1980) tree-ring data remains a major obstacle to further exploration of the extent and causes. Hence we stress the vital requirement for widespread updating of major tree-ring networks, as well as for the acquisition of data for new regions.

Corals

Massive reef-building corals provide high-resolution, continuous records of tropical ocean climate ranging from several decades to several centuries in length. Several characteristics of corals make them ideal tools for the reconstruction of tropical climate: (1) annual skeletal density banding provides chronological control, (2) high linear growth rates of 1–2 cm/yr combined with year-round growth provide annual to subannual resolution, (3) growth of colonies to several metres in height provides several hundred years of continuous growth, (4) location in shallow-water tropical ocean regions (~30°N to 30°S constrained by bathymetry and a variety of environmental factors, Kleypas *et al*., 1999) provides information from regions poorly represented by other sources of high-resolution palaeoclimatic data, and (5) the calcium carbonate skeleton incorporates a range of geochemical tracers that reflect local environmental conditions (see reviews by Gagan *et al*., 2000; Cole, 2003; Felis and Pätzold, 2003; Correge, 2006; Grottoli and Eakin, 2007). Coral skeletons that are well preserved after death allow generation of high-resolution proxy climate records during Uranium/Thorium (U/Th)-dated windows of the more distant past (eg, Tudhope *et al*., 2001; Cobb *et al*., 2003a, b; Felis *et al*., 2004). To date, proxy climate records derived from corals have mostly been used to reconstruct a specific feature of the tropical climate system, in particular ENSO and, to a lesser extent, decadal–interdecadal variability over the last several centuries (eg, Fleitmann *et al*., 2007). Here, after summarizing the basis for extracting climate information from corals, we focus on the broader objective of using coral records for large-scale climate reconstruction.

Specific data

The construction of long, regionally representative palaeoclimate records from massive corals requires careful site selection followed by close inspection of coral skeletal morphology and lithology. The process begins with the identification of a suitably large coral head growing in an open ocean reef environment. The most commonly used massive coral genera are *Porites* in the Indo-Pacific, *Montastraea* in the Caribbean/Atlantic with a couple of records obtained from *Pavona* and *Platygyra.* Recent studies have also demonstrated the palaeoclimatic potential of the more slowly growing species *Diploria* (Draschba *et al*., 2000), *Diploastrea* (Watanabe *et al*., 2003; Bagnato *et al*., 2004) and *Siderastrea* (Guzmán and Tudhope, 1998). Large several metre high coral colonies, although not uncommon, are not found in appreciable numbers on coral reefs (cf. dendroclimatology where there is often a whole forest from which to select suitable samples). The palaeoclimatic value of a particular core must be evaluated in the laboratory, where cores are cross-sectioned into ~5–10 mm thick slices for x-radiography and microscopic inspection. X-ray images reveal density variations associated with annual banding (when present) and growth direction, which guide the placement of transects for subsequent geochemical sampling. Ideally, a good coral slice will have a well-presented annual density banding pattern in which all years are clearly and consistently identifiable along the length of the coral core – this is rarely the case, especially in corals from sites with a small seasonal cycle in temperature. In these cases, other forms of annual banding, including luminescent banding (visible under illumination with longwave ultraviolet light in corals from some locations) and geochemical banding (in particular seasonality in skeletal $\delta^{18}O$ and $\delta^{13}C$) may be used to help establish the chronology. Coral growth characteristics (linear extension rate, skeletal density, calcification rate) can be measured from the x-rays and from gamma or x-ray densitometry (eg, Chalker and Barnes, 1990; Helmle *et al*., 2002), and may provide information related to environmental gradients in sea surface temperatures (SSTs) and water quality (eg, Lough and Barnes, 2000; Carricart-Ganivet and Merino, 2001). Microscopic analysis of thin coral sections is critical to detecting subtle diagenetic alteration (eg, pore infilling by secondary aragonite), which can compromise primary geochemical signals recorded in basal sections of long coral cores (Enmar *et al*., 2000; Muller *et al*., 2002; Quinn and Taylor, 2006). The occurrence and intensity of luminescent lines (observed under ultraviolet light) in near-shore corals from selected locations can provide robust reconstructions of past river flow and rainfall (eg, Isdale, 1984; Nyberg, 2002; Hendy *et al*., 2003; Lough, 2007).

Reconstructions of tropical ocean palaeoclimatic variability for the past several centuries are primarily based on geochemical analyses of the coral skeleton. This requires drilling of carbonate powder at regular intervals along a major coral growth axis. Sampling resolution is most commonly in the range of 4–12 samples per year (eg, Quinn *et al*., 1996). Homogenizing samples milled across entire growth bands yields records with annual to 5-yr resolution (eg, Druffel and Griffin, 1993; Cole *et al*., 2000; Hendy *et al*., 2002; Calvo *et al*., 2007). Commonly employed analysis techniques include using stable isotope ratio mass spectrometry for measuring oxygen and carbon isotopic composition ($\delta^{18}O$ and $\delta^{13}C$, respectively), and inductively coupled plasma mass spectrometry (ICP-MS), inductively coupled plasma atomic emission spectrometry (ICP-AES) and thermal ionisation mass spectrometry (TIMS) for measuring trace and minor element such as Sr/Ca, Mg/Ca, U/Ca and Ba/Ca (for examples, see Beck *et al*., 1992; Shen *et al*., 1992; Min *et al*., 1995; Mitsuguchi *et al*., 1996; Schrag, 1999; Gagan *et al*., 2000; Quinn and Sampson, 2002; McCulloch *et al*., 2003; Fleitmann *et al.,* 2007).

Most coral palaeoclimate proxy records are based on time series of $\delta^{18}O$ and/or Sr/Ca. While skeletal $\delta^{18}O$ is generally interpreted as a measure of SST, sea water salinity, or a combination of the two, coral Sr/Ca variability is considered a more direct measure of SST variations.

Coral skeletal: $\delta^{18}O$

The well established temperature-dependent fractionation of oxygen isotopes between sea water bicarbonate and calcium carbonate means that changes in water temperature are recorded in the coral skeleton, and there are now many empirical studies that show that the temperature dependent slope of coral aragonite $\delta^{18}O$ is close to that predicted from inorganic carbonate studies (eg, Craig, 1965) at 25°C, ie, about −0.2‰ $\delta^{18}O$ per 1°C increase in temperature. It is this temperature sensitivity of carbonate $\delta^{18}O$ that is most often used in the interpretation of coral $\delta^{18}O$ records. However, it is important to remember that changes in the isotopic composition of seawater in which the corals grow also directly impact the isotopic composition of the coral skeleton being precipitated. On seasonal and interannual timescales, the water isotopic composition ($\delta^{18}O_W$) is closely related to salinity, with isotopically enriched, or more positive, $\delta^{18}O_W$ values accompanying high salinity (eg, see Fairbanks *et al*., 1997). In detail, however, $\delta^{18}O_W$ is controlled by the combined effects of local precipitation/evaporation balances, regional sources and sinks of water vapour, ocean mixing and advection, and whole ocean $\delta^{18}O$ composition. Therefore, while the relationship between coral $\delta^{18}O$ and temperature does not vary, the relationship between salinity and $\delta^{18}O$ varies appreciably from site to site, and through time. Nonetheless, many of the published records of coral $\delta^{18}O$ have been used to reconstruct changes in rainfall associated with climatic phenomena such as ENSO (eg, Cole *et al*., 1993; Le Bec *et al*., 2000; Urban *et al*., 2000; Tudhope *et al*., 2001; Cobb *et al*., 2003b), though poor constraints on the relationship with seawater $\delta^{18}O$ on decadal-to-centennial timescales introduces significant uncertainty to climatic interpretations on long timescales.

Coral skeletal: Sr/Ca

The coral skeletal Sr/Ca ratio is considered to be a more direct measure of water temperature (Smith *et al*., 1979; Beck *et al*., 1992), an interpretation based on temperature-dependent changes in the Sr/Ca of inorganic aragonite (Kinsman and Holland, 1969) and an ever-growing body of empirical coral-based calibrations (eg, Alibert and McCulloch, 1997). Although this relationship appears to be strong and robust in some cases, the mechanisms responsible for the temperature relationship remain relatively poorly understood, and some non-temperature-related artefacts have been noted (eg, de Villiers *et al*., 1995; Cohen *et al*., 2002; Gaetani and Cohen, 2006). Poor Sr/Ca-temperature calibrations are observed in corals characterized by slow (less than 6 mm/yr) and/or variable growth rates. Subtle diagenesis has a large effect on coral Sr/Ca-based temperature reconstructions, owing to the shallow slope of the Sr/Ca-temperature relationship (~ 0.7%/K). To date there are relatively few published century-to-multicentury Sr/Ca coral records (Linsley *et al*., 2000; Hendy *et al*., 2002). Further study and more long Sr/Ca records, specifically multiple long records from the same site (eg, Hendy *et al*., 2002), are required to establish the strengths and weaknesses of this tracer for large-scale temperature reconstruction over centuries to millennia.

Description of specific measured proxies and analysed chronologies

There are now ~90 published coral geochemical records but only ~30 (predominantly $\delta^{18}O$ with its potentially mixed response to both SST and salinity) that provide proxy tropical climate records that extend earlier than 1900 (NOAA Paleoclimatology data base http://www.ncdc.noaa.gov/paleo/index.html). The majority (21) of these long coral records are from reefs in the tropical Pacific, six from the Indian Ocean, three from the Atlantic/Caribbean and two from the Red Sea. Several studies have demonstrated that even the current limited spatial network of coral records can capture useful amounts of large-scale climate variance (eg, Evans *et al*., 1998, 2000, 2002; D'Arrigo *et al*., 2005; Wilson *et al*., 2006; Gong and Luterbacher, 2008). The main limitation of these large-scale reconstructions is their reduced fidelity prior to ~1850 as the number of available coral records declines, with only four records by 1700, eight by 1750, 14 by 1800, 21 by 1850 and 32 by 1900. The longest records date back to 1560 for the central Great Barrier Reef (Hendy *et al*., 2002) and 1600 for the eastern equatorial Pacific (Dunbar *et al*., 1994).

Uncertainties in high-resolution coral proxy climate records

There are, as yet, no routine techniques (as used in dendroclimatology see the section 'Proxy data uncertainty') for determining the strength and reliability of climatic signals contained in these coral geochemical records (Lough, 2004). One common problem with calibration and verification of coral geochemical records is the lack of suitable long-term instrumental SST and environmental measurements at remote coral reef locations. Another major problem stems from the use of 'calibrations' based on annual cycles of coral $\delta^{18}O$ and SSTs to quantify interannual to multidecadal palaeo-SST variability. When sufficiently long SST records are available, band-specific calibrations of coral geochemical variations can be performed to check for non-linear coral proxy behaviours.

The largest source of uncertainty in the climatic interpretation of coral $\delta^{18}O$ records lies in the mixed temperature and sea water $\delta^{18}O$ signal contained within these records, and the possible changes in relative contributions of these factors over time or as a function of timescale. For example, $\delta^{18}O$ records can overestimate twentieth-century SST warming if the coral record is considered only as a temperature record and the standard temperature dependence relationship of about -0.2‰ $\delta^{18}O$/°C is applied, as regional freshening at some coral sites has likely contributed to coral $\delta^{18}O$ trends. Indeed, observed tropical SST warming of +0.2°C at 16 Indo-Pacific coral reef sites from 1951–1990 compares with a warming of +0.7°C estimated from coral $\delta^{18}O$ records at the same sites, when attributing $\delta^{18}O$ to temperature only (Lough, 2004). However, in some parts of the tropical oceans, such as the western and central equatorial Pacific warm pool region, temperature and salinity are tightly coupled on seasonal-to-interannual timescales, with high SST driving increased atmospheric convection and accompanying heavy rainfall. Therefore, it is possible to use calibration and verification of the coral $\delta^{18}O$ signal over the recent instrumental period to derive an empirical relationship with temperature, using methods developed for dendroclimatology (eg, Wilson *et al*., 2006). However, it is not clear how well these relationships, which are often biased to interannual variance, will hold for low-frequency changes on multidecadal–centennial timescales for which the SST versus sea-surface salinity relationships may be different. In these cases, an independent measure of SST variations (eg, from skeletal Sr/Ca from the same core) may provide a valuable check on the low-frequency SST variations and trends, as well as allowing the water composition (salinity) signal to be separately resolved. Another approach is to attempt to calibrate skeletal $\delta^{18}O$ to temperature independently for different frequency bands. The difficulty with the latter approach is the shortness of the instrumental record in most locations.

Additional uncertainty results from our limited understanding of the biological mechanisms leading to coral skeleton formation (calcification) and how the incorporation of geochemical tracers used for palaeoclimate reconstruction into the skeleton may be modified by biological processes. This biological overprint has been observed in studies of coral $\delta^{18}O$ and Sr/Ca, and can lead to significant uncertainties in coral palaeoclimate reconstructions in worst-case scenarios. For example, coral skeletal $\delta^{18}O$ values are significantly offset from those predicted by inorganic equilibrium calculations based on seawater $\delta^{18}O$ compositions and temperature. This isotopic disequilibrium or 'vital effect' has been attributed to kinetic isotopic effects (McConnaughey, 1989a, b, 2003). The magnitude of this offset from equilibrium can be substantially different in corals growing in the same location – amounting to mean coral $\delta^{18}O$ offsets of up to 0.4‰ (giving an uncertainty equivalent to ~2°C) in both modern and fossil corals (Linsley *et al*., 1994; Cobb *et al*., 2003b). This coral-to-coral offset poses no problem for interpreting relative changes in a single long coral $\delta^{18}O$ record, as the offset is relatively constant through time if the coral's major growth axis is sampled (Felis *et al*., 2003). Sampling transects that fall off the major growth axis are associated with larger kinetic effects (McConnaughey, 1989a; Hart and Cohen, 1996). However, it is a serious hindrance for interpretation of isolated fossil coral $\delta^{18}O$ sequences, as it is impossible to differentiate a change in mean climate conditions from an offset associated with coral-specific vital effects. One promising strategy for minimizing such errors involves averaging the mean coral $\delta^{18}O$ values across numerous fossil coral heads growing during the same time period (Cobb *et al.*, 2003b). Such coral-to-coral offsets likely exist for coral Sr/Ca records and other coral geochemical proxies, but have not been quantified as of yet. Another kinetic-related 'vital effect' concerns the observation that coral $\delta^{18}O$ and Sr/Ca vary with extension rate in slow-growing portions of corals (usually <6 mm/yr) (McConnaughey, 1989a, b; deVilliers *et al*., 1995; Felis and Patzold, 2003). By focusing on faster-growing coral skeletons, coral palaeoclimate researchers can minimize the contribution from growth-related artefacts to coral palaeoclimate records.

Improving coral-based palaeoclimate records

We now know considerably more about the wide range of potential palaeoclimatic and palaeoenvironmental records locked in the calcium carbonate skeletons of massive corals than when Knutson *et al*. (1972) suggested that they could be applied to determining 'water temperature variations'. Indeed, corals have made a significant contribution to understanding the nature and causes of climate variability in the tropical oceans, in particular interannual ENSO-related variability, but more recently also multidecadal to centennial timescale variability (eg, Linsley *et al*., 2004; D'Arrigo *et al*., 2005; Wilson *et al*., 2006; Gong and Luterbacher, 2008). This contribution to the global, high-resolution palaeoclimatic data base can be substantially extended by (1) improving the reliability of coral geochemical time series through increased replication, (2) increasing the temporal depth of coral records and (3) distinguishing between temperature and hydrological sources of variability in coral $\delta^{18}O$ records.

The fact that most published coral time series are based on records extracted from single coral cores prevents the quantitative separation of the desired climate signals from noise at a given site. Replication is essential to identify non-climatic artefacts in individual coral records, to ensure dating accuracy and to identify large-scale environmental signals that are shared among different coral records. Ideally, at least three cores from a reef location are necessary. Given the rarity of long, multicentury cores, it may be more practical to undertake the replication work on more common shorter cores that span the last several decades.

A revised globally coordinated sampling strategy may be necessary to increase the temporal depth of coral records to span the last millennium. To date, much of the sampling focus has been on obtaining proxy records from key locations influenced by ENSO events. However, future sampling should also target cooler reef sites, as average linear extension rates of massive corals vary directly with average annual SST so that corals in cooler waters grow more slowly than those in warmer locations (Lough and Barnes, 2000). This also means for the same-sized coral colony, one living in relatively cooler tropical waters will tend to contain more years of record than one living in warmer waters. Realistically, however, coral reconstructions that pre-date AD 1500 must be based on long-dead corals that are scattered on reef tops and beaches throughout the tropics. Such corals can be U/Th dated, analysed for $\delta^{18}O$ and Sr/Ca, and then spliced together to yield multicentury windows on tropical marine climate (Cobb *et al*., 2003a, b). Given appropriate material, such 'fossil' coral reconstructions can be spliced to the oldest portions of corresponding 'modern' coral reconstructions

Reducing the uncertainties surrounding the interpretation of coral $\delta^{18}O$ time series requires the implementation of multiple strategies. First, better characterization of the environments in which sampled coral colonies live would help reduce uncertainty in the climatic interpretation of records derived from them. For example, more routine use of *in situ* loggers (temperature and, ideally, conductivity) would allow the response of the coral to be better understood with respect to local and regional conditions. Even a few years of *in situ* logger data can help identify the extent to which the coral's local environment is representative of the more relevant regional-scale conditions. Such local monitoring should also include the generation of rainfall and seawater $\delta^{18}O$ time series, where practical, as the IAEA's Global Network of Isotopes in Precipitation contains very few ocean island sites, obscuring the relationship between large-scale climate and seawater $\delta^{18}O$ changes. A second, complementary approach involves increasing the number of coral Sr/Ca records from sites where coral $\delta^{18}O$ records already exist. This approach could be used to isolate the seawater $\delta^{18}O$ component of coral $\delta^{18}O$ records, which remains poorly constrained, especially on decadal-to-centennial timescales. Finally, as with any potential source of proxy climate information, there is still a need for detailed process studies of how to extract the most reliable palaeoclimate records from corals, both modern and fossil. This is particularly the case for the development of relatively new proxies such as Sr/Ca.

The coral palaeoclimatic community would also benefit from the development of a set of standard sampling, screening and reporting protocols that allow the quality and reliability of different coral records to be assessed. A meaningful step towards this goal would be to encourage the storage of more complete sets of metadata (including water depth and GPS locations of drilling, x-radiograph images of the cores and chronological methodologies) on national data servers such as NOAA/NCDC (see section 'Tree rings and the need for improved regional and temporal coverage'). Such information would also facilitate the informed use of coral records by researchers outside the coral community, particularly those interested in multiproxy climate field reconstruction.

High-resolution ice core climate records for the last millennium

Ice cores provide a rich record of environmental tracers trapped as snow progressively accumulates on ice sheets and glaciers. Climate processes leave an imprint on the delivery of moisture and contaminants that provides a basis for reconstructing climate parameters from the ever-increasing suite of measurable species.

The principal variable from ice cores that has been used in multiproxy climate reconstructions (eg, Mann and Jones, 2003; Pauling *et al.,* 2003, Jones and Mann, 2004; Luterbacher *et al.*, 2004; Schneider *et al.*, 2006) is the isotopic composition of the water, which is generally treated as a proxy for temperature. The main focus of this section is on these water isotope records: their interpretation, associated uncertainties and availability. However, the optimal use of the isotope records relies on other measurements, particularly high-resolution trace chemical measurements (eg, McConnell *et al.*, 2002; Kaspari *et al.*, 2004; Dixon *et al.*, 2005) which provide better resolved annual layers at lower accumulation sites where diffusion of water vapour extinguishes annual variability from the isotope record (see later discussion). In the future, it is likely that climate reconstructions will be enhanced by incorporating other ice core parameters. A growing range of such additional climatic indicators is available, from the local net annual snow accumulation to proxies for regional- and global-scale circulation modes (eg, ENSO, Meyerson *et al.*, 2002; Bertler *et al.*, 2004; NAO, Vinther *et al.*, 2003; Mosley-Thompson *et al.*, 2005), mid- and high-latitude sea-level pressures, the Southern Annular Mode and circulation indices (eg, Souney *et al.*, 2002; Goodwin *et al.*, 2004; Shulmeister *et al.*, 2004; Kaspari *et al.*, 2005), and sea-ice extent (Curran *et al.*, 2003). At certain sites, records of summer temperatures may be derived from the thickness and frequency of melt layers in the ice (Fisher *et al.*, 1995).

Ice cores have also provided significant information about global-scale climate forcing over recent millennia. They provide the main source of data on past concentrations of all the major long-lived greenhouse gases (MacFarling Meure *et al.*, 2006). They also provide annually resolved records of volcanic fallout that can be used to estimate past forcing by volcanic aerosols (eg, Robock and Free, 1995; Palmer *et al.*, 2001) and they offer one means to estimate the past strength of solar activity (eg, Bard *et al.*, 2000).

Clearly, ice core records are limited to ice-covered sites, but frequently such locations are precisely where other sources of proxy climatic data are not available. Many ice cores provide sufficient detail for 'high resolution' records, taken here to mean those with resolvable annual variations that enable a countable chronology with near-absolute precision (say a few percent). This precision is aided by reference to markers of known age from volcanic events and nuclear bomb testing (Mosley-Thompson *et al.*, 2001). The ability to resolve annual variations is determined by the amount of snow deposited, the stratigraphic disturbance by meltwater percolation (if any), postdepositional modification of the seasonal signal by processes such as vapour diffusion, and the removal, redeposition and/or mixing of near-surface snow by the wind.

The goal of obtaining records that extend over multiple millennia imposes a further constraint of sufficient ice sheet thickness. Sites which meet these combined requirements are found in Greenland, across temperate and tropical alpine regions and in higher accumulation zones in Antarctica.

The full geographic extent of potential sites has not yet been exploited. Significant initiatives within the ice core community, under the banner of IPICS (International Partnerships in Ice Core Sciences, Brook and Wolff, 2006), are expected to expand the network of ice cores, especially those that provide records for the last two millennia. This work extends other programmes, notably PARCA (Program for Arctic Regional Climate Assessment, Mosley-Thompson *et al.*, 2001; Thomas *et al.*, 2001) and SCAR-ITASE (Scientific Committee on Antarctic Research–International Trans Antarctic Scientific Expedition, Mayewski and Goodwin, 1997; Mayewski *et al.*, 2005), which have collectively extracted many ice core records from Greenland and Antarctica, respectively. In the context of dwindling alpine glaciers, the impetus for recovering records from these temperate locations is clear and urgent. These are briefly discussed in a later part of this section.

Ice core isotopic records: interpretation and uncertainties

The strong correlation between the isotopic composition of precipitation and local temperature is a well-known property at mid-to-high latitudes (Craig, 1961; Dansgaard, 1964). This provides an essential tool for reconstructing palaeotemperatures from ice core measurements of $\delta^{18}O$ or δD (hereafter abbreviated together as δ). In fact, the processes controlling fractionation are complex and reflect conditions at the moisture source, during atmospheric transport and at the points of condensation and final precipitation. However modelling studies confirm observations that at polar sites it is the final stages of moisture transport, and site temperature in particular, that dominates the δ signal (Helsen *et al.*, 2006). For alpine sites, depletion during transport and 'washout' (a dependence of isotopic fractionation on precipitation amount), together with seasonality in these effects, are often more important than local temperature in controlling fractionation. A significant body of literature is devoted to understanding and calibrating the 'isotope thermometer' (eg, Johnsen *et al.*, 1972; Jouzel *et al.*, 1997), key aspects of which are considered here.

Before continuing the discussion of water isotopes in ice cores, some general properties and constraints arising from the nature of ice cores need to be addressed. The most fundamental among these is the property that ice cores only record a δ signal during precipitation events, which individually do not represent the mean climate. Furthermore, seasonal variations in the timing and magnitude of precipitation events can lead to additional bias, and this may also be variable and subject to climatic modulation (Charles *et al.*, 1995). *In situ* studies, comparisons with meteorological data, model simulations and direct comparisons among different ice core species can help detect such bias and place limits on its influence.

In addition, non-climatic variability, or noise, results from surface roughness (sastrugi) and from postdepositional reworking, which lead to uneven stratification of the snow (Steffensen *et al.*, 1997). This depositional noise increases as the accumulation rate diminishes relative to surface relief, so that the highest quality records are expected from sites with higher accumulation rates. The signal-to-noise ratio (SNR) in highly resolved Greenland $\delta^{18}O$ data differs greatly across the ice sheet. SNR for annual $\delta^{18}O$ at the south Greenland Dye-3 drill site (annual accumulation 0.56 m ice) is ~2.5 (Fisher *et al.*, 1985, 1996), while the SNR is ~0.9 for the Greenland Ice Core Project (GRIP) drill site (annual accumulation 0.23 m ice) at the summit of the Greenland ice sheet (Johnsen *et al.*, 1997). Likewise, in Antarctica, the lowest noise is found for higher accumulation rate coastal areas, where SNR is comparable with that for sites with similar accumulation rates in Greenland (Jouzel *et al.*, 1997). Hence, while single ice core δ records are useful from sites with high accumulation rates such as Law Dome, Antarctica (0.7 m ice for the DSS site) and southern Greenland, for lower accumulation sites (eg, Greenland summit) replication is preferable and averaging of several annual δ records should be undertaken before a regional climatic interpretation is pursued (White *et al.*, 1997).

A further influence on the fidelity of high-resolution reconstructions from ice cores arises from the 'timing noise' in both environmental markers and precipitation events. Annual markers in the ice for defining years (typically mid-summer or mid-winter extrema in a measurable parameter) will not correspond to a fixed calendar date, and so computed annual averages will partition some portion of the signal into the adjacent year. While timing noise reduces the fidelity of annual and shorter-term reconstructions, the impact diminishes rapidly for averages over a few years. Timing noise can also be reduced by using multiple species to define annual horizons (eg, Palmer *et al.,* 2001).

Rapid ice flow can cause non-climatic trends in δ records. Such trends arise because ice cores drilled on a sloping surface will contain upstream ice originating from progressively greater elevations at progressively greater depths in the core. The Dye-3 $\delta^{18}O$ record is known to be affected by such upstream effects (Reeh *et al.*, 1985). It is essential to identify and correct ice core δ data for flow-related trends, if the cores are not drilled at a site with modest ice flow (ie, on an ice divide or on the summit of a dome). Even where flow is minimal, evolution of the ice sheet itself (eg, elevation changes) may introduce non-climatic temperature or isotope fractionation trends. This possibility needs to be considered before interpretation of the data.

Temporal resolution of ice records is limited by the accumulation rate, postdepositional noise, diffusive smoothing (see below), analytical sample size requirements and the finite size and number of precipitation events. Modern analytical techniques permit some species to be sampled at many tens of samples per year; however the actual number of precipitation events that can be distinguished above the surface reworking noise is much smaller. Glacier thickness can impose a further limit on high-resolution records because the progressive thinning of annual layers with depth, as a result of ice flow, ultimately reduces the layer thickness to the point where layers are not analytically resolvable. For alpine glaciers and thinner marginal regions of Antarctica and Greenland this rapid thinning can restrict annual records to the century-scale or less.

Turning specifically to issues affecting δ records, we first consider the limiting effect of diffusive smoothing on resolution. In the upper several tens of metres of the ice sheet where the ice matrix is permeable (the firn zone) water vapour is free to move diffusively. By the time the ice attains a state of impermeability, where vapour movement ceases, the integrated effect is sufficient to extinguish isotopic variations on vertical scales less than 20–25 cm of ice. This places a fundamental limit on the resolution of any reconstruction. It also limits the ability to use annual cycles in δ for layer counting, although signal processing techniques can be used to enhance cycles and improve countability (Johnsen and Vinther, 2007). Even at high accumulation sites, where seasonal variations survive diffusive smoothing, the amplitude of variations is progressively attenuated with depth in the firn. This introduces non-climatic trends in winter and summer season δ values which must be removed by prior correction for diffusion, if seasonally separated records are to be reconstructed (Vinther *et al.*, 2008).

Dating of most layer-counted cores uses a combination of annual cycles in isotopes, where these can be resolved, and annual cycles in trace chemicals. Most trace chemicals are typically (though not exclusively) free from movement in the firn, and with modern continuous analysis systems can be routinely sampled at the centimetre scale. This facilitates identification of well-resolved annual layers (sampled on the order of around ten times per year) for annual accumulation rates above 10 cm ice-equivalent. For such accumulation rates, vapour diffusion smoothes the water isotope signal over 2–3 years.

The observed strong correlation between δ and site temperature (*T*) suggests a straightforward approach for temperature reconstruction, by simply applying the observed δ–*T* regression slope (eg, Dansgaard, 1964). However, this spatially derived calibration need not apply to the temporal variations at a given site, and a more appropriate temporal calibration of the isotope thermometer has been determined either from direct calibrations using co-located data on water isotopes and site temperature (either instrumental or derived from a different proxy method), or from a modelling approach using atmospheric GCMs that have the capability to explicitly simulate water isotopes (eg, Cuffey *et al.*, 1995; Johnsen *et al*., 1995; Werner et *al*., 2000). Temporal regression slopes are generally difficult to compute directly from data since there are few records of site temperature of sufficient length at ice core sites to derive a calibration, and even where limited data are available, the temperature excursion in the record is typically small, limiting the calibration. At highly resolved sites, the larger subannual temperature variability can be used for calibration (van Ommen and Morgan, 1997), although the result will reflect seasonal variations in transport in addition to temperature.

One major issue that emerges from both modelling and data studies is that both the calibration slope and the quality of the δ–*T* relationship varies according to the time period covered. Thus a different value emerges from data studies comparing monthly values, annual values and decadal trends (eg, Peel *et al*., 1988; Rozanski *et al*., 1992; Kohn and Welker, 2005). It is, therefore, necessary to use a calibration in a manner appropriate to the study being conducted.

For coastal, high-accumulation Antarctic records, temporal calibration over recent years gives low δ–*T* slopes (Peel, 1992; McMorrow *et al.*, 2004). This result is also found in isotopic GCM studies (Werner and Heimann, 2002). Calibrations for high-elevation areas of inland Antarctica are more difficult, since few recent climatic data exist for these inland sites. An alternative approach is to investigate the δ–*T* relationship over very large past climate changes such as glacial–interglacial transitions. Even here, direct estimates are difficult in central Antarctica since the low-accumulation regime limits the utility of borehole temperature reconstructions, while the absence of abrupt temperature changes in Antarctica renders the thermal diffusion effect that has been used successfully in Greenland (see next paragraph) inapplicable. However, indirect evidence using palaeoprecipitation estimates to infer temperature and comparing with δ gives calibration slopes consistent with the spatially derived value (Jouzel *et al.*, 2003) and GCM modelling studies agree with this conclusion (Jouzel *et al*., 2007). In any case, it is the lower elevation records which provide the high-resolution chronologies considered here. So we are left with a dilemma since the longest-term calibration data from cores in the high interior of East Antarctica give a calibration slope similar to the spatial value, while the shorter-term calibrations possible from higher accumulation coastal sites give a much lower slope: whether these calibrations apply at these sites on different timescales, particularly for decadal and centennial changes is unclear, and needs further data and model-based calibration work.

For Greenland, long-term calibrations have been made by comparing borehole temperature data with δ values (eg, Cuffey *et al.*, 1995; Johnsen *et al.*, 1995). Another method is based on estimating the magnitude of past rapid temperature changes from combined gravitation and thermal diffusion of the air in the firn column (Severinghaus *et al.*, 1998). These two approaches give calibration slopes for δ–*T* at Greenland sites which are in good agreement with slopes estimated from seasonal oscillations in highly resolved central Greenland cores (Shuman *et al*. 1998), but are considerably lower than the spatially derived values, and values from simple Rayleigh-type distillation models. However, the two techniques apply particularly to the major climate changes of the past, and the reduced slope at least partly derives from a change in the seasonality of snowfall (less winter snow in very cold periods) (Werner *et al*., 2000; Krinner and Werner, 2003).

The different calibrations applicable at different sites and for different timescales of change underscore the importance of understanding the processes responsible. At higher accumulation sites, precipitation is typically dominated by the large atmospheric cyclonic systems that reach the Antarctic coast. Further inland, more of the precipitation arrives in the form of ice crystals, although infrequent penetration of large storms can still bring significant snowfall. The net effect of these different depositional processes on the isotope temperature calibration has yet to be quantified. It is not clear that the fractionation relationships for the two types of precipitation regime should be the same. It is clear, for example, that at Law Dome transport, as well as temperature, plays a considerable role on interannual timescales. On longer timescales, it is likely that the transport variability diminishes in importance and temperature correlation increases (van Ommen and Morgan, personal communication, 2008).

Greenland ice cores

During the past five decades, numerous ice core records have been retrieved from the Greenland ice sheet, many of them during the Greenland Ice Sheet Program (GISP) in the 1970s (Langway *et al.*, 1985) and PARCA in the 1990s. Records covering the past two millennia stem solely from ice cores drilled to bedrock during deep drilling campaigns, ie, Camp Century, Dye-3, GISP2, GRIP, Hans Tausen, NGRIP and Renland records (Langway, 1967; Dansgaard *et al.*, 1982, 1993; Johnsen *et al.*, 1992; Alley *et al.*, 1995; Hammer *et al.*, 2001; North Greenland Ice Core Project (NGRIP) Members, 2004). Of these seven ice cores, the Dye-3, GRIP and NGRIP cores have been measured at sufficient resolution to provide annually resolved $\delta^{18}O$ data over the whole of the past two millennia, and two shallow cores drilled near the GRIP drill site also provide such data for the past millennium (Johnsen *et al.*, 1997). The GISP2 core has been measured with annual resolution back ~1200 years (Grootes and Stuiver, 1997).

Annually and seasonally resolved $\delta^{18}O$ data have been shown to be closely related to Greenland coastal temperatures (Barlow *et al.*, 1993; White *et al.*, 1997; Rogers *et al.*, 1998; Vinther *et al.*, 2003) and a comparison with a new Greenland coastal temperature record extending back to 1784, showed persistent and highly significant correlations between winter season $\delta^{18}O$ from seven Greenland ice cores and winter temperatures in SW Greenland (Vinther *et al.*, 2006b). A new study based on seasonal data from multiple Greenland shallow ice cores indicates that winter season $\delta^{18}O$ can also be used as a proxy for local and coastal annual average temperatures (Vinther *et al.*, 2008). Indeed winter season $\delta^{18}O$ seem to be a significantly better proxy than annual average $\delta^{18}O$ for both coastal Greenland annual average temperatures and local temperatures reconstructed from borehole temperature profiles (Vinther *et al.,* 2008). Greenland winter season $\delta^{18}O$ has also been found to contain a clear NAO signal (Vinther *et al*., 2003).

Recently the Dye-3, GRIP and NGRIP cores have been cross-dated throughout the Holocene using volcanic reference horizons as tie-points (Rasmussen *et al.*, 2006; Vinther *et al.*, 2006a). It is expected that the common dating for the three cores will facilitate extraction of regional-scale Greenland climatic signals. Seasonal variations can be resolved for the last two millennia in the Dye-3 and GRIP $\delta^{18}O$ records, while diffusional processes in the top 60 m of the ice sheet destroys the seasonal oscillations at the lower accumulation NGRIP drill site (Johnsen, 1977; Johnsen *et al.*, 2000; Vinther *et al*., 2008). The upcoming North Greenland Eemian (NEEM) deep drilling project in NE Greenland is expected to produce at least one additional seasonally resolved δ record spanning the last two millennia.

Antarctic ice cores

There are only a few long, annually resolved ice core records from Antarctica dated with better than decadal precision for millennial or longer timescales. Currently, the longest high-resolution continental temperature reconstruction from ice cores (Schneider *et al.*, 2006) draws upon cores at five sites to construct a two-century record. Of these five sites, the longer records come from Dronning Maud Land (1000 years; Graf *et al.*, 2002), Law Dome (700 years; Morgan and van Ommen, 1997) and ITASE cores 2000-01 (~350 years) and 2000-02 (~300 years; Steig *et al.*, 2005). Annually resolved records are also available from the 302 m core (~550 years) drilled at Siple Station at the base of the Antarctic Peninsula (Mosley-Thompson *et al.*, 1990) and the upper 190 m (~480 years) of a core from the Dyer Plateau in the Antarctic Peninsula (Thompson *et al.*, 1994). When compared with the highest resolution δ histories available at that time, the Siple and Dyer δ records strongly suggest that the seesaw in temperature between the Peninsula region and the East Antarctic Plateau evident in the available meteorological data (1945 to 1985) likely persisted for the last five centuries (Mosley-Thompson, 1992). Further high-resolution measurements of Law Dome core material (V. Morgan, personal communication, 2008) will provide annually resolved δ data back 2000 years.

New cores emerging from the West Antarctic Ice Sheet (WAIS) Ice Divide deep ice coring site and Berkner Island may potentially give interannual- to decadally resolved records. Earlier data from Berkner Island provide annual cycles countable with subdecadal precision back ~750 years. The accumulation rates at the WAIS and Berkner sites are not sufficient for preservation of annually resolved δ, because of diffusion, but trace chemical data may provide dating (J. McConnell, personal communication, 2008) with high precision.

Lower latitude ice cores

Ice cores have also been recovered from high-elevation ice fields at lower latitudes. The interpretation of δ histories is challenging for lower latitude, high-elevation cores, particularly those in the tropics and subtropics where large seasonal differences dominate the precipitation regime. Calibration efforts have been limited by the lack of contemporaneous instrumental temperature climatic records as most drill sites are in very remote locations. There has been vigorous discussion in the literature as to whether the δ records from the highest tropical and subtropical ice caps and glaciers reflect local temperature or precipitation (via the amount effect) or large-scale patterns of tropical climate. Rozanski *et al.* (1992, 1993, 1997) reviewed *in situ* data and suggested that to a first approximation, temperature is the primary control on isotopic fractionation in the high- to mid-latitudes but the amount of precipitation is the more dominant process in the tropics. More recent investigations of the processes controlling δ in snowfall at very high elevations in low latitudes suggest a more complicated picture (Thompson and Davis, 2005; Mosley-Thompson *et al.*, 2006).

One of the first efforts to quantify the controls on δ in snowfall over the Tibetan Plateau was conducted by Yao *et al.* (1996) who measured δ in fresh precipitation collected at selected meteorological stations where ambient temperature and precipitation data were available. They found that on timescales ranging from a discrete precipitation event to an average over several months the trends in δ reflect trends in temperature more strongly than in precipitation amount at sites in northern Tibet. In southern Tibet where monsoonal precipitation dominates, they found trends in δ more closely linked to precipitation amount than temperature. On annual timescales the δ–*T* relationship reflected atmospheric dynamical processes (more depletion in the warm, wet season); however, on longer timescales atmospheric temperature appeared to be the more dominant control on δ.

Another process affecting δ in precipitation on the high ice fields in the Himalayas and the South American Andes is the intensity of atmospheric convection. Increases in intense convection result in condensation at higher elevations and hence at colder temperatures (Thompson *et al.*, 2000). The resulting precipitation is more isotopically depleted. The interpretation of δ in Andean precipitation on shorter timescales has been addressed by Grootes *et al.* (1989), Henderson *et al.* (1999), Bradley *et al*. (2003) and Vuille *et al.* (2003); however, no simple interpretation has emerged. Bradley *et al.* (2003) reported a strong link between sea surface temperatures across the equatorial Pacific and the δ histories in ice cores from the tropical Andes and the Dasuopu Glacier in the Himalayas. They also reported a strong link between trends in δ and ENSO variability, which is not surprising as ENSO integrates all the mechanisms mentioned above. On multicentury to millennial timescales there is strong evidence that δ is more strongly controlled by temperature than precipitation (Thompson *et al.*, 2000, 2003). Even on shorter timescales (multiannual to centuries) there is abundant evidence that precipitation amount is not the primary control on δ in precipitation over ice fields in the tropics and subtropics (see Thompson *et al.*, 2006: figure 5). However, as with the polar ice cores, a better understanding of the differential roles these various processes play in controlling δ on high-elevation, low-latitude ice fields remains an area ripe for further investigation (eg, Schmidt *et al*., 2007).

Ice cores offer a wealth of information that extends well beyond that provided by stable isotopic ratios. They provide records of regional changes in the climate and environment as well as histories of larger, global-scale changes. A more complete picture of the Earth system's climate history requires the extraction and accurate interpretation of high-resolution (ideally annual) ice core records from the polar ice caps to the highest mountains in the tropics and subtropics. These histories must extend back many centuries to multiple millennia. Such a collection of ice-core-derived histories is closer to becoming a reality with the IPICS initiative that calls for a global array of 2000-yr cores (Brook and Wolff, 2006).

Documentary and early instrumental data

Documentary data include all forms of written historical information about past climate and the weather. The sources are numerous and varied ranging from direct human observations (of sea ice, frozen lakes and rivers or snow lines), through the effects of the weather on important aspects of crop growth, such as yields and harvest dates, to paintings of glaciers in the Alps. The sources encompass all possible aspects of written history that were routinely noted in diaries and also because it was required for state, legal and tax reasons. The use of documentary data in past climate reconstructions is, therefore, restricted to locations where there are long written histories (such as Europe, China, Japan and Korea, where some extensive analyses have been undertaken, but also potentially includes the Islamic World, where little has been analysed and published). Brázdil *et al*. (2005) provide an extensive summary of the type of proxy evidence used, European documentary sources and available reconstructions. Luterbacher *et al*. (2006) complement this with a review of the documentary evidence from the Mediterranean area. Information from eastern Asia and Japan is not summarized in a single recent paper, but is generally split into those looking at China, Japan or Korea (Wang and Zhao, 1981; Zhang and Crowley, 1989; Song, 1998, 2000; Mikami, 1999, 2002; Wang *et al*., 2001; Qian *et al*., 2002; 2003; Yang *et al*., 2002; Ge *et al*., 2003, 2005, 2008).

Although most analyses have been undertaken in these regions, there is scope for some geographic expansion using early European colonial archives (for Iceland and Greenland involving sea-ice records in particular, Ogilvie, 1992, 1996; Ogilvie and Jónsdóttir, 1996; for the Americas, eg, Bradley and Jones, 1995; Overland and Wood, 2003; Druckenbrod *et al*., 2003 for North America and Quinn and Neal, 1992; Prieto and Herrera, 1999; Prieto *et al*., 2000, 2004; Ortlieb, 2000; García-Herrera *et al*., 2008 for South America) and in barely accessed archives of the Ottoman Empire in Turkey (for the Middle East region). Further, records from Spanish galleons crossing the Pacific Ocean during the sixteenth to eighteenth centuries provide a description of secular changes in the wind circulation (García *et al*., 2001). The systematic abstraction of naval and merchant ship logbooks has made available a new data base with daily wind observations over the Atlantic and Indian Oceans for the period 1750–1850 (eg, Jones and Salmon, 2005; Wheeler, 2005; García-Herrera *et al*., 2005a; Gallego *et al*., 2005) and for the later decades will provide instrumental measurements of SSTs and atmospheric pressure. The existence of possible secular variations in the occurrence of Atlantic hurricanes in the sixteenth to eighteenth centuries is also suggested from records obtained from the same sources (García-Herrera *et al*., 2005b). There are also a large number of as yet unexplored ship logbooks (mainly British) that have great potential to improve and extend sea level pressure reconstructions and SSTs further back in time (eg, Jones and Salmon, 2005; García-Herrera *et al*., 2005a).

Arabic documentary sources have been widely used in astronomy and geophysics to date phenomena such as eclipses and auroras. However, they have not yet been explored from a climatic perspective. J.M. Vaquero (University Extremadura, Spain, personal communication, MedClivar Meeting, Carmona, 2006) provided several examples of their use to construct series of extreme events such as droughts and floods in southern Iberia. Significant collections have been preserved in Spain, frequently translated into Spanish that could provide the basis for the reconstruction of extreme events in south-central Iberia for the period AD 750–1200. The actual availability of these sources in the Mediterranean Arabic countries is currently unknown.

Documentary sources, by their very nature, will tend to emphasize extreme events, primarily in the winter and summer seasons and to a lesser extent spring, as these are the types of records that are more likely to have been recorded because of their socio-economic relevance. Records indicating conditions in past autumns tend to be more rare (Pfister, 1992; Xoplaki *et al*., 2005). Luterbacher *et al*. (2007) endorse this latter conclusion but still show that the warmth in autumn 2006 in Central Europe was very likely exceptional within the last 500 years. Until the early 1970s most documentary evidence was collected into compendiums about past weather with little regard for whether the reporter was local or contemporary in time (Brázdil *et al*., 2005 and references therein). Since the mid-1970s, much greater rigour has been used, with long series developed combining a whole array of documentary sources into indices of past seasonal temperature and precipitation conditions (eg, Pfister *et al*., 1998, 1999; Glaser, 2001; Meier *et al*., 2007 for Switzerland and Central Europe; van

Engelen *et al*., 2001 for the Low Countries and Le Roy Ladurie, 2004; Chuine *et al*., 2004 for France). Similar series have also been developed for China, Japan and Korea (see earlier references).

Documentary data are the only kind of palaeoclimatic information based on direct observations of meteorological variables in terms of narrative descriptions. They are highly temporally resolved with detailed descriptive precision. Important uses of documentary data are the verification of extreme values in natural proxies such as tree rings and for the detailed descriptions of past weather conditions. Most importantly, they are the only evidence that is directly related to the socio-economic impacts of rare but significant disasters in the period prior to the organization of the network of instrumental observations (eg, Brázdil *et al*., 2005). In Europe, the most difficult problem is that the types of historical sources available prior to the middle of the eighteenth century are markedly less widely available since that time (Pfister, 1992; Pfister *et al*., 1999; Jones *et al*., 2003; Luterbacher *et al*., 2004; Xoplaki *et al*., 2005; Brázdil *et al.*, 2005 and references therein). The widespread use of new meteorological instruments at that time meant the death knell for many types of documentary sources (eg, diaries) as the observers (who were often scientists or medical doctors) quickly moved to using the new instruments. The development of long documentary-based records extending to the present has led to a variety of approaches to fill the recent gap. The most common approach is to use degraded instrumental records for much of the last 200 years (Jones *et al*., 2003). In Europe, this issue means that it is difficult to assess the accuracy of the true documentary part of the long series for an accurate calibration, as there are only a few areas with short periods with an overlap of documentary and instrumental data (see Dobrovolný *et al.,* 2008 and references therein). This problem is less of an issue in the Far East, as here documentary sources continue well into the twentieth century, particularly in China, and instrumental records only really begin there in earnest in the late-nineteenth century.

Early instrumental data

Instrumental recording began in Europe in the late-seventeenth century, but only a few long series exist from this time (see Jones and Briffa, 2006). In most other parts of the world, the widespread use of instruments began much later (see eg, Jones and Thompson, 2003). It is often believed that instrumental records only extend back to the founding of National Meteorological Services (NMSs), but for most countries there are generally much longer records available. Sometimes they will take some tracking down, but there are considerable amounts of early data for Africa, South America and southern Asia in European colonial archives. Recently early Japanese records have been found (in Dutch archives) from the late-eighteenth century, taken by Dutch traders at Nagasaki (Können *et al*., 2003) and data for other locations are beginning to be found in Japanese archives (Zaiki *et al*., 2006).

Instrumental series are an essential part of palaeoclimatology, as they provide the basis for assessing the usefulness and reliability of all proxy records. For almost all sites, though, it is rare for the temperature and precipitation records to be fully homogeneous over their full length of record. Much work must generally be undertaken to ensure that adequate adjustments (for changes in sites, instruments, recording hours, etc.) are made to all instrumental records to ensure their long-term homogeneity. Numerous methods and approaches have been applied (see Jones and Thompson, 2003 and references therein). Because longer instrumental records enable better proxy calibration and verification exercises, ensuring the earliest parts of series are correct is vital. The most difficult aspect of ensuring homogeneity relates to thermometer exposure issues before the widespread use of variants of Stevenson screens in the late-nineteenth century, affecting mostly European records. The issue is particularly important for summer temperatures, which might be too warm because of the possibility of direct solar influences on the early thermometers (see the discussion in Moberg *et al*., 2003; Böhm *et al*., 2009). Rebuilding pre-Stevenson screens (from mid-nineteenth century diagrams) or using data loggers (at the locations of the earliest sites) has recently enabled the issue to be addressed in Spain and Sweden (Brunet *et al*., 2006 and Klingbjer and Moberg, 2003, respectively), but much more work is needed. The recent intensive digitization of many of the long daily European records is providing new insights into long-term homogeneity by additionally looking at air pressure and cloudiness observations (Camuffo and Jones, 2002 and Moberg *et al*., 2003). New approaches for homogeneity assessment of long daily series are also important advances (Della-Marta and Wanner, 2006).

Regional climate reconstructions

Temporal and regional high-resolution temperature reconstructions show important climatic features, such as regionally very hot or cool summers or autumns and very mild or cold winters or springs that may be masked in hemispheric or global reconstructions (eg, Luterbacher *et al*., 2004, 2007; Brázdil *et al*., 2005; Casty *et al*., 2005, Guiot *et al*., 2005; Xoplaki *et al*., 2005). Thus, regional studies and reconstructions of climate are important when climate impacts are evaluated. Seasonal extremes at regional-to-continental scales, such as the hot European summer of 2003 (Luterbacher *et al*., 2004), the mild winter of 2006/2007 (Luterbacher *et al*., 2007) and warm spring of 2007 (Rutishauser *et al*., 2008) exhibit much larger amplitudes than extremes at the global or NH scale, thus they may markedly affect the local-to-regional natural environment and impact societies and economies.

In their large-scale Northern Hemisphere (NH) temperature reconstructions (MBH98) Mann *et al*. (2000) pointed out that there is a decrease in the number of independent patterns of variability (eg, principal components) that can be reliably reconstructed prior to 1760. This leads to greater spatial smoothing and thus to less high-frequency variance at continental spatial scales. Consequently, to accurately reproduce the spatial pattern within regions, an extensive and dense proxy network is required to capture several of the principal components. By combining long homogenized station temperature series and temperature-derived indices from documentary evidence as well as a few seasonally resolved natural proxies, Luterbacher *et al*. (2004) and Xoplaki *et al.* (2005) were able to map seasonally resolved temperature anomalies across European land areas for individual years back to 1500, along with area-specific error estimates. Multivariate principal component (PC) regression (later in section ‘Combining proxies to reconstruct large-scale patterns, continental and hemispheric averages’ also termed truncated Empirical Orthogonal Functions, or EOFs, regression) was used for the monthly (back to 1659) and seasonal (four values per year; 1500–1658) European land surface air-temperature fields (at 0.5° by 0.5° resolution). This method places a greater weight on specific proxy series that exhibit a greater affinity with the twentieth-century large-scale instrumental record (see Luterbacher *et al*., 2002b for a detailed description of the methodology). The mapping permits a rigorous assessment of the spatial coherency of past annual-to-decadal climatic changes at a subcontinental scale, and also allowed Pauling *et al*. (2003) to calculate the ‘best’ predictors of boreal winter and summer temperatures from the available array of different proxy climate data for different parts of Europe and the North Atlantic Ocean (particularly involving documentary evidence from sea-ice extents north of Iceland). It shows, for instance, as could be intuitively expected, that tree rings have been a good predictor of past summer temperatures across northern and central Europe, whereas documentary sources are more reliable for reconstructing winter temperatures. Brönnimann *et al*. (2007) recently investigated the influence of the ENSO phenomenon on European climate using these gridded continental-scale reconstructions. After removing years following tropical volcanic eruptions, they find a consistent and statistically significant ENSO signal in European climate in late winter and spring.

Since some of the most important potential consequences of past climate changes have been linked to changes in regional/continental circulation patterns (eg, Shindell *et al.*, 2001; Luterbacher *et al*., 2002a; Casty *et al*., 2007), the frequency and intensity of droughts and floods (eg, Jacobeit *et al*., 2003; Mudelsee *et al*., 2003; Pfister *et al*., 2006) and hurricane activity (eg, Gárcia-Herrera *et al*., 2005b), regional and large-scale reconstructions of changes in other climatic variables, such as precipitation, over the last centuries provide a valuable complement to those for temperature (NRC, 2006). In recent work, Pauling *et al*. (2006) used a multiproxy data set including precipitation measurements, precipitation indices derived from documentary sources as well as precipitation-sensitive natural proxies to derive European-scale precipitation maps with associated uncertainties, back to 1500. Casty *et al*. (2005) applied multivariate PC regression to a similar data set as Luterbacher *et al*. (2004) to reconstruct seasonal temperature and precipitation variability over the greater-Alpine area. Both studies found that precipitation at regional and continental scales is much more difficult to estimate back in time than temperature, and explained variances are almost always lower than for temperature.

Other potential high-resolution proxies

The most important feature of the high-resolution proxies already discussed is that they involve as near to absolute dating as possible. This is a fundamental aspect of the use of tree rings and documentary information, where it is ensured through crossdating and source assessment and replication, respectively. For both ice cores and corals, there is a growing emphasis in recent years on the analysis of multiple sample data sets and the use of tree-ring techniques to improve dating accuracy. Better reconstructions of climate variability have been achieved, for both these sources, when data from more than one core have been integrated. Other potential sources of absolutely dated records include those from varved sediments (lacustrine and marine) and speleothems. As with ice cores and corals, as well as other archives assumed to have absolute time control, extra effort must always be made to confirm the exact nature of the temporal resolution and accuracy before using varved sediment and speleothem records for reconstructing the high-resolution palaeoclimatology of the last millennium.

Varved sediment archives

Over the last couple of decades, quantitative palaeoclimatic reconstructions from terrestrial (eg, lake and wetland) and marine sediments have been widely generated and used. An important, and growing, subset of these reconstructions includes records where time control is provided by annually laminated, or ‘varved’, sediments. Varved sediments have been found in lakes on almost all continents, and from the tropics to high-latitudes, as well as in some restricted marine basins (eg, the Cariaco Basin and the Santa Barbara Basin) and high-latitude fjords, where bottom-waters usually low in oxygen inhibit bioturbation. There are many types of varved sediment, some being biogenic in origin, others chemical or physical (ie, sedimentological). In most cases where varves have palaeoclimatic utility, they are made up of two or more seasonal, or subseasonal, laminae that can be recognized as together constituting an annual period of sediment accumulation. Varves are rarely of use for reconstructing the high-resolution palaeoclimatology of the last millennium unless both their chronology and climate sensitivity is well understood.

In terms of chronology, the most important step in using varved sediments is to document that the sediment laminations are indeed reliable annual varves. Ideally, this takes the form of three mutually confirming analyses (eg, Lamoureux, 2001; Shanahan *et al*., 2008). First, multiple independent age estimates (eg, from ^{210}Pb, ^{14}C, bomb nuclides, tephras, etc.) need to be used to show that the laminated sediments in question are annually laminated. Second, it is also helpful when the nature of the sedimentation is well understood and known to be consistent with a documented varve formation mechanism. This can be explored in a great variety of ways depending on the lake or marine basin setting, but usually involves both a detailed subannual characterization (eg, biology, chemistry, physical stratigraphy) of the varved sediments, as well as knowledge of corresponding environmental variability in and around the lake or basin; the two should be consistent in a well-documented fashion. Third, and where possible, a good correspondence between reconstructed varved-based environmental (ie, climatic) variability and that observed in an instrumental or other independent record, should also be documented. Modern varved sediment chronology building has also evolved to utilize multiple replicate cores from a single basin, rather than only one or two, as a way to reduce chronological uncertainty resulting from lake, sediment and coring processes that can disturb sediments. Even with such efforts, however, chronological error can often be as much as 2–5%, perhaps more, so users of varved sediment data should pay particular attention to making sure that they do not overestimate temporal accuracy.

Once a varve-based chronology is available, then a wealth of available sediment-based quantitative climate reconstruction methods can be used to produce a varved-based palaeoclimate record. As with chronology, the key is to confirm that the resulting record represents what it is purported to represent, usually via a time-series comparison with instrumental or other independent climatic data. Some approaches to varve-based palaeoclimate reconstruction use variations in the varves themselves calibrated against instrumental data (eg, Hughen *et al*., 2000; Besonen *et al*., 2008; Romero-Viana *et al*., 2008) or in the chemical and isotopic variations of the sediments calibrated against instrumental data (eg, Field and Baumgartner, 2000). A wider set of varved sediment reconstructions have used spatial relationships between proxy and climate to generate a ‘transfer-function’ analogue or other means to convert temporal sequences of varved sediment proxy data into quantitative time series (including errors) of inferred palaeoclimate. This has been done with pollen data (eg, Gajewski, 1988), diatoms (eg, Bradbury *et al*., 2002; Stewart *et al*., 2008), foraminifera (eg, Field, D.B. *et al*., 2006), and other biotic proxies. With these reconstructions, many of the usual concerns must be evaluated (eg, Telford and Birks, 2005; Batterbee *et al*., 2008), including the possibility that there is a temporal lag (eg, Overpeck *et al*., 1990; Overpeck, 1996), or frequency band-pass, bias between climate variability and the proxy response (eg, vegetation as reflected by fossil pollen data) to that climate variability. Single references focused on methods used with varved sediments tend to evolve quickly given the rapidly evolving nature of the field, so readers should consult the extensive varved sediment literature regularly for the latest developments.

Speleothem archives

Speleothems are cave deposits, such as stalactites and stalagmites, formed when calcium carbonate (usually calcite) precipitates from degassing solutions seeping into limestone caves. Palaeoclimatic reconstruction efforts typically focus on stalagmites, whose growth patterns are more regular than stalactites. Growth rates vary between ~ 0.05 and 0.4 mm/yr and provided annual bands are present and well preserved, the combination of annual band counts and Uranium/Thorium-series (U/Th) dating results in absolute chronologies, but with age uncertainties (Fleitmann *et al*., 2004). Dating errors arise from several sources, including analytical error in U/Th dates, uncertainty in corrections for detrital thorium in most speleothems, and the fact that minor hiatuses in stalagmite growth often go unresolved. Chronological uncertainties can be greatly improved by obtaining records from multiple overlapping stalagmites (via crossdating), and the generation of multiple estimates of contaminant thorium composition. The thickness of the bands can be used to reconstruct climate as well as trace metal concentrations and $\delta^{18}O$ and $\delta^{13}C$ isotopic ratios, with the latter being the most frequently used climate proxies. Very few studies (eg, Smith *et al.*, 2006) have focused on the reconstruction of continental-scale climate variability during the past 500 to 1000 years using speleothems, mainly because of the difficulty in devising high-resolution sampling strategies. There are a number of well-known reconstructions (Polyak and Asmerom, 2001; Fleitmann *et al*., 2003, 2004, 2007; Proctor *et al*., 2004; Mangini *et al*., 2005; Partin *et al*., 2007), but, at present, dating limitations appear to preclude reliable comparison with higher-resolution proxies such as tree-rings, corals and documentary data.

Lower-resolution proxy records

There are a wide range of terrestrial and palaeoclimate archives, and two types provide the continuous and quantitative records of use for reconstructing the high-resolution palaeoclimatology of the last millennium: lake/wetland sediments and temperature–depth profiles measured in boreholes. Because annually laminated, varved sediments have the potential for annual, even seasonal, resolution of both terrestrial and marine palaeoclimate, these archives are discussed above in the section ‘Other potential high-resolution proxies’.

Non-varved sediments are common from both lakes and wetlands from all of the continents of Earth, and thus are extremely valuable for reconstructing a wide range of terrestrial palaeoenvironmental information. When it comes to the last millennium, however, available time control limits the utility of these non-varved sediments. On the positive side, the combination of ^{210}Pb, bomb-nuclide and other geochronological tools available for dating the last *c*. 150 years of sediment accumulation make it possible to generate proxy records that can be compared directly with instrumental data for proxy calibration and verification purposes (see section ‘Other potential high-resolution proxies’ for more detail). However, below *c*. 150 years, the need to rely on radiocarbon and other radiometric dating methods results in temporal errors that likely exceed 10%, or even 20% in many cases. Exceptions exist, for example, where volcanic tephra of known ages can be used to constrain age models, but any use of sediments for understanding the high-resolution palaeoclimatology of the last millennium needs to provide clear evidence that the sediment geochronology is sufficiently accurate, and that bioturbation is not distorting the sediment record. For example, one way is to demonstrate that the sediment-based reconstruction is capable of reproducing climate variations observed in the instrumental record, and in this case, not just a monotonic trend that is observed in the instrumental record.

One new approach being attempted employs Bayesian hierarchical methods that rigorously entrain explicit statistical models of proxy–climate relationships across multiple-frequency bands (cf. Ammann *et al*., 2007a).

Borehole archives

Within the context of millennial climate reconstructions, borehole profiles have been one source of information that has significantly contributed to our understanding of centennial temperature changes. Temperature–depth profiles measured in boreholes contain a record of temperature changes at the Earth’s surface. Various approaches to deriving past temperature at local- and hemispheric-scales based on the information recorded in temperature logs have become well established within the last decade (eg, Huang *et al*., 2000; Harris and Chapman, 2001; Beltrami, 2002; Beltrami and Bourlon, 2004; Pollack and Smerdon, 2004; Gonzalez-Rouco *et al*., 2008). Reconstructions have been developed using the large number of borehole logs and derived-temperature reconstructions (Huang and Pollack, 1998) that in 2004 included 695 sites in the NH and 166 in the SH (see Jansen *et al*., 2007: figure 6.11).

As the solid Earth acts as a low-pass filter to downward propagating surface temperature signals, borehole reconstructions lack annual resolution, so they typically portray multidecadal-to-centennial timescale changes. Temperature reconstruction based on borehole profiles depends on the assumption that surface air temperature (SAT) changes are coupled to ground surface temperature (GST) changes and propagate to the subsurface by thermal conduction merging with the background geothermal gradient field. The downward propagation of the temperature disturbance is a function of the relatively small thermal diffusivity of rock strata, so that the first few hundred metres below the surface store the integrated thermal signature of the last millennium (eg, Gonzalez-Rouco *et al*., 2008). Long-period events such as the temperature changes associated with postglacial warming will still produce significant signal amplitudes at depths greater than 1000 m. A palaeoclimatic reconstruction derived from borehole temperatures is characterized by a decrease in temporal resolution as a function of depth and time. Typically, individual climatic events can be resolved from borehole temperatures only if their duration was at least 60% of the time since its occurrence, ie, an event that occurred 500 years before present will be seen as a single event if it lasted for at least 300 years; otherwise events are incorporated into the long-term average temperature change (Gonzalez-Rouco *et al*., 2008).

Borehole reconstructions have an advantage in that they arguably do not have to be calibrated against the instrumental record because temperature itself is measured directly. However, a disadvantage is that, as discussed above, they measure GST rather than the desired quantity (SAT), and under certain conditions there may be substantial differences between the them. Important quantitative uncertainties exist, but general trends in these reconstructions are likely robust. Estimates of the magnitude of recent warming as hemispheric (continental) averages, or averages over the middle latitudes, are approximately 0.7–0.9°C, from the mid-nineteenth century to the late-twentieth century (Pollack and Huang, 2000; Huang *et al*., 2000; Harris and Chapman, 2001), which is similar to the increase estimated from the instrumental record. Some borehole-based reconstructions also indicate an earlier persistent but smaller warming of roughly 0.3°C from 1500 to 1850 (Pollack and Huang, 2000; Huang *et al*., 2000). Regional estimates for the warming for the last 150 years range from 2°C to 4°C for northern Alaska (Lachenbruch and Marshall, 1986; Pollack and Huang, 2000) to 0.5°C for Australia (Pollack and Huang, 2000). Estimates for the western and eastern sectors of North America are rather different, 0.4–0.6°C and 1.0–1.3°C, respectively (Pollack and Huang, 2000). Such a reconstruction (for the NH) is similar to the cooler multiproxy reconstructions in the sixteenth and seventeenth centuries, but sits in the middle of the multiproxy range in the nineteenth and early-twentieth centuries (eg, Jansen *et al*., 2007).

Borehole reconstructions based on the available data base generally yield somewhat muted estimates of the twentieth-century trend because of the relatively sparse representation of borehole data north of 60°N and the fact that many were logged more than 20 years ago. The assumption that the reconstructed GST history is a good representation of the SAT history has been examined with both observational data and model studies. The mean annual GST differs from the mean annual SAT in regions where there is snow cover and/or seasonal freezing and thawing (eg, Smerdon *et al*., 2004; Taylor *et al*., 2006). The long-term coupling between SAT and GST has been addressed by simulating both air and soil temperatures in GCMs. Mann and Schmidt (2003) suggested that GST estimates during the winter season are biased by seasonal influences related to changing snow cover, and that less that 50% of the total spatiotemporal variance in GST is explained by SAT variations during the cold half of the year. Chapman *et al*. (2004) contest the implications this has for recent temperature trends (however, see Schmidt and Mann, 2004), and some long-term simulations (eg, by González-Rouco *et al*., 2003, 2006, using the ECHO-G model) suggests the possibility that seasonal differences in coupling might be of little significance over long timescales as long as temperature trends are similar in different seasons. However, in cases where there are large seasonal differences in climate trends (a possibility that remains for the past few centuries, see Mann *et al*., 2003) such seasonal bias issues could lead to misleading inferences regarding long-term SAT trends from indicators of past GST change.

Marine sedimentary archives

Palaeoclimatic reconstructions generated from marine sedimentary archives offer a unique view of ocean climate change, even extending to a three-dimensional view of past states of the ocean through the use of depth transects across ocean basins. The greatest progress has been achieved in the reconstruction of SSTs based on proxies that make use of the chemical composition or abundance of planktonic organisms. Depending on the region of interest, seasonal temperature estimates can be obtained by using different proxies (eg, alkenone unsaturation index, faunal assemblages and Mg/Ca ratio of foraminiferal shells).

The classical approach has been to use the species abundance of plankton, most commonly planktonic foraminifera and diatoms, using transfer functions calibrated by statistically comparing the surface sediment distributions to modern surface ocean hydrography. A critical assessment of these spatially based transfer functions (Telford and Birks, 2005) indicates that some early claims of the accuracy of this approach were overstated. The optimal reproducibility of transfer function estimates appears to be at best ±1°C (Telford and Birks, 2005). Progress over the past years has been particularly good concerning reconstruction of sea-surface temperatures based the chemical composition of shells of planktonic organisms (Cronin *et al*., 2003) or biomarker molecules originating from specific plankton groups. Depending on the region of interest and availability of microfossils in the sediments, seasonal temperature estimates can also be obtained by using different chemical proxies such as an alkenone unsaturation index or the Mg/Ca ratio of foraminiferal shells. The uncertainty of the reconstructions is on the order of ±0.5–1°C for these methods, but the uncertainty increases with lower temperature and lower salinities.

Examples of high-resolution SST records with a resolution between 2 and 30 years over recent millennia, based on transfer functions and geochemical temperature proxies, include surface layer temperatures and dynamics in the low- and high-latitude regions of the North Atlantic (eg, Andersson *et al*., 2003; Andersen *et al*., 2004, Black *et al*., 2007). Also recently published records indicate an imprint of solar irradiance on sea-surface temperature in the North Atlantic (Jiang *et al*., 2005) and upwelling intensity variations off northwest Africa (McGregor *et al*., 2007). High-resolution proxy records of sea-ice cover off Iceland are published with decadal resolution, comparing well with historical data (Moros *et al*., 2006). In general, major century-scale cold episodes, apparent in Northern Hemisphere multiproxy temperature reconstructions, are detected in the marine high-latitude North Atlantic SST reconstructions (eg, a cooling centred around 1450, cooling in the seventeenth century and the first half of the nineteenth century, a generally warmer period 1000–1100 and the twentieth century warming).

Changes in large-scale ocean circulation (eg, in the rate of the Atlantic Meridional Overturning Circulation, also referred to as Atlantic Thermohaline Circulation), and the associated poleward heat transport are of primary interest for a comprehensive assessment of natural climate variability. Using vertical profiles of the oxygen-isotope composition of planktonic and benthic foraminifera it is possible to infer changes in the density structure of water masses. This approach has been successfully employed to quantify past variations in Gulf Stream transport through the Florida Strait during the last millennium (Lund *et al*., 2006). So far this approach has only produced records of at best century-scale resolution, but it seems feasible to improve on this given optimal conditions. Associated with the changes in Florida Strait throughflow were changes in ocean salinity as recorded at multidecadal resolution by the combination of Mg/Ca palaeothermometry and $\delta^{18}O$ data. The results show that reduced Florida Strait throughflow occurred with enhanced salinities of the Gulf Stream waters (Lund *et al*., 2006).

Sedimentological proxies based on grain size analyses are being used to estimate changes in deep-water flow speed. Although the method awaits calibration to quantify changes in velocity, some initial results are very promising. A bi-annual record covering the last 230 years suggests variations in flow speed along the deep-water pathway in the North Atlantic that parallel changes in the phase of the North Atlantic Oscillation (Boessenkool *et al*., 2007).

A new method based on the protactinium–thorium ratio of foraminifera shells (McManus *et al*., 2004) holds great promise for assessing variations in large-scale mixing and overturning of the ocean, which is of great relevance for oceanic heat transport as well as the global carbon cycle. So far this method has not been employed successfully on high-resolution sediments, and there is a need for improved calibrations and refinement to be able to detect smaller-scale changes in the overturning circulation than those of the last deglaciation, and to detect changes occurring on shorter timescales than a century.

In some areas marine sediments provide records of terrestrial and atmospheric processes. The Cariaco basin in the waters off Venezuela contains decadal-scale records of runoff variability associated with north/south migrations of the Inter Tropical Convergence Zone, obtained through the titanium content of the sediments which are related to local riverine inputs (Haug *et al*., 2001, 2003). Such records can be linked to terrestrial palaeoclimatic records, documenting drought patterns in the Caribbean (eg, Hodell *et al*., 2005).

In the Mediterranean Sea, considerable attention has been paid recently to obtaining high-resolution records of SST, salinity and water chemistry for the Holocene, using new archives such as vermetid reefs (living reefs, with data spanning 500–600 years and with resolution every 30–50 years), non-tropical corals (ie, living *Cladocora caespitosa* for the last 100–150 years, eg, Silenzi *et al*., 2004, 2005; Montagna *et al*., 2006 and references therein) and deep-sea corals (ie, living *Desmophyllum dianthus*, *Lophelia pertusa*, *Madrepora oculata* for last 100–150 years). Vermetids are thermophile and sessile gastropods living in intertidal or shallow subtidal zones, forming dense aggregates of colonial individuals. These new archives potentially complement and improve the information derived from the main climate indicators such as foraminifera, alkenones, dinoflagellate cysts, calcareous nanoplankton and, especially for the Mediterranean Sea, serpulid overgrowth on submerged speleothems (Antonioli *et al*., 2001). All these latter marine markers enable longer palaeoclimate reconstructions but with a much coarser resolution (usually lower than 100–200 years). However, in areas of extremely high sedimentation rates (>80 cm/ka), such as in the southern Levantine Basin, a resolution of 40–50 years may be possible during the last millennium (Schilman *et al*., 2001).

Combining marine sedimentary proxy records with reconstructions from tree rings, ice cores and corals can be very challenging because of the uncertainty associated with the dating of the marine records. The dating error is generally significantly larger than for the other types of proxy archives. Nevertheless, significant progress has been made using well-dated ash layers and ^{210}Pb dating for the most recent time span as well as improved marine radiocarbon chronologies. Future developments in producing records of the short-term variations in the radiocarbon reservoir age of surface waters will help improve the situation in coming years (eg, using schlerochronology, Scourse *et al*., 2006). Owing to the dynamical nature of the ocean surface and its variable seasonal stratification, major uncertainties remain in terms of ascribing the proxy data to the correct season and water depth when the carrier signal was derived. As an example, studies of Holocene high-resolution records from the same core have revealed major differences in how the proxies responded to the increased summer insolation and seasonal contrasts of the early Holocene. Some proxies, recording the temperature during the summer season of the euphotic zone (typically the upper 50 m) document strong warming in response to increases in insolation, whereas proxies that derive their chemical signature deeper in the surface layer did not respond to the insolation changes (Jansen *et al*., 2008). Thus improved knowledge about the inclusion of the climatic signal into the carrier organisms is required. On the other hand, the differences point to improved possibilities to capture seasonal differences using multiproxy approaches.

At present only a small number of marine sedimentary archives have been retrieved that allow potential interannual resolution. An increasing number of records with temporal control that are adequate to assess multidecadal-to-century scale variations have been produced and concerted efforts to further improve this situation are underway in the marine palaeoclimate community. Detailed surveys of the seafloor in recent years have revealed many potential sites, where such high-resolution sedimentary archives could be obtained. Acquiring such archives would be a major step towards better integration of marine sedimentary proxy data with other proxy data. To reduce the uncertainties associated with marine proxies it will be necessary to conduct open-ocean culturing experiments to optimize the calibration of palaeoenvironmental proxies. A co-ordinated effort of the palaeoceanographic community, therefore, could generate a marine equivalent to the land-based temperature reconstructions for the last few millennia, developing much improved records for integration with land-based records.

Combining proxies to reconstruct large-scale patterns, continental and hemispheric averages

The various approaches to combination

Most previous proxy-based reconstructions of large-scale climate have targeted hemispheric or global average temperature, though some studies have also attempted to reconstruct the underlying spatial patterns of past surface temperature changes at global (eg, MBH98/99; Rutherford *et al*., 2005; Wahl and Ammann, 2007; Mann *et al*., 2007) and regional (eg, Fritts *et al*., 1979; Briffa *et al*., 1988, 1994; Luterbacher *et al*., 2004, 2006, 2007; Casty *et al*., 2005; 2007; Xoplaki *et al*., 2005; Meier *et al*., 2007; Riedwyl *et al*., 2008a) scales, and other fields such as precipitation (Casty *et al*., 2005; Luterbacher *et al*., 2006; Pauling *et al*., 2006), Palmer Drought Severity Index (PDSI; eg, Cook *et al*., 2004b, 2007; Nicault *et al*., 2008), sea level pressure (Briffa *et al*., 1986; Luterbacher *et al.*, 2002b) and SST (eg, Evans *et al*., 2002; Wilson *et al*., 2006). Still other studies have focused on the reconstruction of particular climate indices that represent coherent large-scale patterns of variability, such as the North Atlantic Oscillation (NAO) and the related Northern Annular Mode (NAM), the Pacific Decadal Oscillation (PDO), the Atlantic Multidecadal Oscillation (AMO), the Southern Annular Mode (SAM) and indices that describe the variability of the ENSO phenomenon. An extensive review is provided by Jones and Mann (2004).

A comparison of methods: CPS and CFR

Most reconstructions of large-scale mean climate (eg, hemispheric-average temperature) or of individual climate indices have employed the 'Composite Plus Scaling' (CPS) method. A selection of climate proxy records are first standardized, then averaged (composited) and finally centred and scaled to provide an estimate of the temperature series for the region or hemisphere represented by the proxy records (the 'target' time series).

The CPS method has been implemented in a variety of ways, contributing to the wide range of possible reconstructions. The proxy-selection process may make use of information about the correlation between individual proxy records and their local climate (eg, Briffa *et al*., 2001; Mann and Jones, 2003; Osborn and Briffa, 2006), or might be based on more general prior expectations of the climate signal contained within the candidate proxy records (eg, season and climate variable they represent) or of their retention of low-frequency variability (eg, Esper *et al*., 2002). The averaging stage may be unweighted (eg, Jones *et al.*, 1998) or weighted, with weights determined according to some intrinsic assessment of reliability (eg, Briffa *et al.*, 2001), according to an assumed area of representation (eg, Esper *et al*., 2002; Mann and Jones, 2003) or in some cases determined to give a best-fit to the target climate record (eg, the inverse regression results of Juckes *et al*., 2007). In the latter case, the CPS method may have some parallels with the climate field reconstruction (CFR) methods discussed below, because in some CFR techniques (eg, Briffa *et al*., 1986), the reconstructed fields (and, therefore, their spatial average) can be expressed as a weighted sum of the proxy records used in the reconstruction. A worthwhile future endeavour would be to examine the relationships between individual proxy records and large-scale climate patterns that arise in the context of CFR-based approaches.

After forming the (weighted or unweighted) average of the individual proxies, the composite time series is centred and scaled to achieve a quantitative estimate of the target climate series in appropriate physical units. The composite time series is centred by the addition of a constant, chosen so that (over a defined period during which they overlap) the time-mean of the composite time series is equal to the time-mean of the instrumental target series. Finally, the centred composite time series is multiplied by a scaling coefficient. The scaling coefficient has been determined in a variety of ways, including (i) regression of the proxy composite onto the temperature (eg, Briffa and Osborn, 2002), (ii) regression of the temperature onto the proxy composite (eg, the inverse regression results of Juckes *et al.*, 2007), (iii) matching the variance of the composite to that of the temperature record (eg, Jones *et al.*, 1998; Crowley and Lowery, 2000; Mann and Jones, 2003; Moberg *et al.*, 2005; D'Arrigo *et al.*, 2006b; among others) or, more recently, (iv) by using total least squares regression and apportioning the regression error between both the temperature and proxy records (Hegerl *et al.*, 2007). Another potentially useful extension to the CPS method has been to combine high-pass filtered tree-ring data with low-pass filtered information from low-resolution (decadal- or centennial-scale) proxies that might be better-suited to reconstructing low-frequency climate variability (Moberg *et al*., 2005). This latter method does not, however, overcome the limitation that low-resolution information cannot be calibrated by temporal regression against the relatively short instrumental temperature record. By applying the same scaling coefficient to the filtered low-resolution composite as to the high-pass filtered tree-ring data, Moberg *et al.* (2005) implicitly rely on the standardizing process to ensure that the low-frequency component is not artificially inflated relative to the higher-frequency component (Mann *et al*., 2005).

An alternative to CPS is the climate field reconstruction (CFR) approach, which assimilates proxy records into reconstructions of the underlying spatial patterns of past climate change (eg, MBH98; Luterbacher *et al*., 2002a, 2004; Casty *et al.*, 2005, 2007; Rutherford *et al*., 2005; Xoplaki *et al.,* 2005; Pauling *et al*., 2006; Ammann and Wahl, 2007; Mann *et al*., 2007; Riedwyl *et al*., 2008a,b). A number of statistical methods have been used for CFR. All of these methods make use of the covariance between proxy series and instrumental series during a calibration period, but there are key differences in how they use covariance within the proxy data set or within the instrumental data set to combine records into a smaller number of variables. Some examples are as follows.

(1) Luterbacher *et al.* (2004) implement separate multiple regression equations between each leading principal component (PC) time series of the proxy records and all the leading PC time series of the instrumental data, thus explicitly making use of covariance within the separate proxy and instrumental data sets. This approach has been given various names, including principal components regression, multivariate truncated-EOF regression, or orthogonal spatial regression (Cook *et al.*, 1994). The climate field is then reconstructed by multiplying each reconstructed instrumental PC time series (or RPC) by the corresponding instrumental data EOF pattern, and then summing the resulting patterns (termed 'EOF expansion').
(2) The core of the MBH98 method is based on separate multiple regressions between each 'individual' proxy record and all the retained PCs of the instrumental data. Covariance within the proxy data set is utilized, both because some 'individual' proxy records are actually principal components of regional tree-ring chronology networks and because the least squares simultaneous solution of these multiple regressions implicitly depends on covariability between proxy records. Once the RPCs are obtained in this method, the climate field is reconstructed from the instrumental EOF expansion as in (1) above.
(3) Cook *et al.* (2004b) use covariance within the tree-ring data to define PCs of the proxy data, but then use these in separate multiple regressions for each grid-box in the instrumental climate domain (the point-by-point regression approach).
(4) The Regularized Expectation Maximization (RegEM, Schneider, 2001) method makes use of covariance within the proxy data, within the instrumental data and between proxy and instrumental data, in the form of regression equations between all proxy and instrumental variables considered. The regression equations are fitted to the available data, taking into account the covariance between variables and the estimated covariance of the regression residuals. The regressions are then used to reconstruct the earlier climate variables, and then the procedure is iterated until convergence. The regularization step is necessary to avoid overfitting these regression equations. Rutherford *et al.* (2005) used ridge regression to regularize the algorithm; effectively this downweights variability that is present in relatively few series compared with variations that are coherent across many series. Mann *et al.* (2007) regularize the algorithm by substituting all proxy and instrumental series by the leading PCs of the combined proxy and instrumental data set, truncating to exclude less important PCs. In this latter implementation of RegEM, the climate field is not directly reconstructed, rather the RegEM algorithm is used to determine the RPCs of the climate field. Once the RPCs are obtained, the climate field is reconstructed from the instrumental EOF expansion as in (1) above.

Compared with CPS reconstructions, CFRs allow a more detailed investigation of past climate by highlighting spatial differences, thereby also approaching the socially relevant local-to-regional scale. Hemispheric or global averages, as well as any climate indices of interest, can then be computed directly from the reconstructions of the underlying spatial field. Another potential advantage of the CFR approach is that the reconstructed spatial patterns can be directly compared with model-predicted patterns of climate responses to forcing (Shindell *et al*., 2001, 2004; Waple *et al*., 2002), provided that the details that discriminate different patterns can be reconstructed with sufficient skill. In this context, CFR methods are used to validate model simulations while model simulations are used as testing ground for the robustness and reliability of reconstructions (pseudo-proxy experiments, see below).

CFR methods make use of both the local proxy–climate relationship and non-local climate teleconnections by relating the long-term proxy climate data to the temporal variations (PCs) in the large-scale patterns of the spatial field of interest (generally defined as the empirical orthogonal functions, or EOFs, of that field). Examples of these teleconnected, or non-local, relationships include the close tie between drought (and therefore drought-sensitive tree rings) over the western USA and SST patterns in the Pacific Ocean (eg, Stahle *et al.*, 1998), or ice core records from Greenland which are closely tied to the behaviour of the NAO (Appenzeller *et al.*, 1998; Vinther *et al*., 2003, 2006b), which influences cold-season temperature patterns over North America and Eurasia. The CFR approach takes fuller advantage of the potential climate information in the proxy data set by exploiting these relationships. Investigations using synthetic proxy data (so called 'pseudo-proxies'), suggest that these assumptions are reasonable for the range of variability inferred over the past one or two millennia (see section 'Introduction to the pseudo-proxy approach'). A more severe restriction of CFR (and also of the CPS approach) is, however, the usually rapidly decreasing number of proxies prior to the instrumental period (~1850), limiting the ability of this approach to properly capture the entire climate field in earlier years. A detailed discussion of this issue is also provided in the section 'Experiments with pseudo-proxies'.

For both CPS and CFR methods, the utility of the reconstructions for, eg, detecting unusual climate change or comparing with climate model simulations (Collins *et al.*, 2002), is greatly enhanced by the estimation of reconstruction uncertainties. The magnitude of the uncertainties will generally vary with timescale, although this is not always explicitly considered (exceptions are the examination of the power spectrum of calibration residuals by MBH99, and the presentation of uncertainties for different timescales in figure 5 of Briffa *et al.*, 2001). Often the uncertainties are estimated from the residuals between the actual and reconstructed values during the calibration period, though this will likely be an underestimate of real uncertainties because (i) uncertainties in the scaling coefficient (for CPS) or in the linear model coefficients (for CPS with individual proxy weights determined by regression against the climate data and for CFR methods) will inflate the error outside of the calibration period (or, equivalently, there will be a tendency to over-optimize the linear model coefficients by using chance correlations between proxy errors and climate data, thus artificially reducing the magnitude of the residuals); and (ii) there may be some long-term inhomogeneities in the proxy records or deterioration in the proxy reliability that cannot be identified within the calibration period. The former source of additional error can be estimated by using known properties of the calibration algorithm (such as standard formulae for estimating the standard error of regression coefficients – used, eg, by Briffa *et al.*, 2002a) though pseudo-proxy studies (see below) can provide an alternative estimate of the total calibration error and should be used more in future work. Both sources of additional error can, at least partly, be addressed by derived error estimates from residuals over an independent verification period or cross-validation experiments (eg, Rutherford *et al.*, 2005; Mann *et al.*, 2007). Finally, uncertainties can also be underestimated if prior analysis steps – such as the selection of proxy records – have already used the calibration (or verification) climate data (Bürger, 2007; Osborn and Briffa, 2007).

Introduction to the pseudo-proxy approach

Experiments involving synthetic climate proxy records, or 'pseudo-proxies', derived from climate model simulations (or, in some cases, from instrumental climate data) can be used to investigate many different aspects of reconstruction sensitivity, such as the impact of different proxy networks (not just temporal and spatial coverage – eg, Mann and Rutherford, 2002 and Rutherford *et al.*, 2003 – but also type of climate proxy – eg, Zorita and González-Rouco, 2002 and Küttel *et al.*, 2007), different levels of individual proxy errors (eg, Mann *et al.*, 2007; Lee *et al.*, 2008), and the behaviour/performance of different reconstruction algorithms (eg, Bürger *et al.*, 2006; Ammann and Wahl, 2007; Lee *et al.*, 2008; Riedwyl *et al.*, 2008a,b; von Storch *et al.*, 2008) at different timescales and for different sets of calibration data (eg, Mann *et al.*, 2007; Lee *et al.*, 2008). The pseudo-proxy approach selects only a limited sample of grid boxes from the complete global fields of relevant GCM integrations (principally those incorporating all important forcings), often according to the location of real proxy data, and degrades these grid-box time series by the addition of noise, chosen to represent the unknown error in real proxy data. In all the above cases, multiple repetitions may be made with different realizations of random noise or even of random pseudo-proxy selection. In each case the reconstruction target is exactly known (being the model-generated full climate field or the subset/average that is of interest). A much more detailed and systematic investigation into the potential performance of climate reconstructions can, therefore, be undertaken using the pseudo-proxy approach than can be obtained by simple calibration and verification tests with real-world proxy and instrumental climate data.

Apparent success within the context of individual and specific pseudo-proxy tests does not, of course, guarantee the reliability of a particular method in all reasonable situations, including the real-world situation. Two particular limitations of the pseudo-proxy approach are that (i) our knowledge of the characteristics of real proxy records, including a full understanding of their response to climate variations and a full statistical model of their error, is still incomplete, which prevents the generation of completely realistic pseudo-proxies; and (ii) the climate model simulations may not replicate real-world behaviour, either because of model deficiencies or because of unrealistic forcing or initialization (Osborn *et al.*, 2006). The former limitation has been partly addressed by using a wide range of noise models to generate multiple sets of pseudo-proxies mimicking the specific temporal characteristics of real world proxies (Mann *et al.*, 2007; Moberg *et al.*, 2008; Riedwyl *et al.*, 2008a,b). The latter limitation is being partly addressed by using simulated data from more than one climate model (eg, von Storch *et al.*, 2004; Küttel *et al.*, 2007; Mann *et al.*, 2007; Lee *et al.*, 2008; Riedwyl *et al.*, 2008a,b), though it has also been argued (Osborn and Briffa, 2004) that even a partially unrealistic model simulation is still a useful test-bed for pseudo-proxy climate reconstructions – ie, a well-behaved reconstruction algorithm should work in a relatively wide range of plausible situations. The latter argument is less applicable if particular modes of climate variation, such as ENSO, are of key importance to a particular study and are poorly simulated.

The pseudo-proxy approach can, subject to these limitations, provide useful insight into the relative performance of various reconstruction techniques, under different conditions such as calibration period and proxy quality. A current initiative of the PAGES/CLIVAR Intersection is a 'Paleoclimate Reconstruction Challenge' (Ammann, 2008; see also http://www.pages.unibe.ch/science/prchallenge/index.html), which will allow multiple groups to test different algorithms within the same pseudo-proxy setting. In particular, the core activity will involve a double-blind experiment, where those attempting to reconstruct the simulated climate will be provided with pseudo-proxy and pseudo-instrumental data, but will be unaware of the actual simulated climate or the actual errors imposed on the pseudo-proxy data. This mimics the information available when attempting real-world climate reconstructions.

Before considering some specific results of recent pseudo-proxy studies, it is worth emphasizing that, while they can provide important insights, they can never prove the veracity of a particular real-world reconstruction; in particular, their results are dependent on the quality of the pseudo proxies and the structure of noise and bias in real-world proxies cannot be determined with full confidence.

Experiments with pseudo-proxies

Pseudo-proxy intercomparison of multiple reconstruction methods

Pseudo-proxy experiments have been used to test the performance of both the CPS (Mann *et al.*, 2005; Ammann and Wahl, 2007) and CFR methods (Mann and Rutherford, 2002; Rutherford *et al.*, 2003; Zorita *et al.*, 2003; von Storch *et al.*, 2004; Mann *et al.*, 2005; Bürger *et al.*, 2006; Ammann and Wahl, 2007; Smerdon and Kaplan, 2007; Küttel *et al.*, 2007; Mann *et al.*, 2007; Riedwyl *et al.*, 2008b), though in many cases only a single reconstruction method is considered (sometimes with a range of modifications). Lee *et al.* (2008) compared the skill of several different CFR and CPS reconstruction methods using millennial climate model simulations. Their intercomparison included evaluation of CPS techniques using simple variance matching, forward and inverse regression, total least squares regression, and an elaboration of these techniques based on a state-space time series model. The Lee *et al.* (2008) evaluation also considered two CFR reconstruction approaches: MBH98 and RegEM (Schneider, 2001; Rutherford *et al.*, 2005). Note, however, that regularization in RegEM was performed by means of ridge regression (Rutherford *et al.*, 2005), rather than the truncated total least squares (TTLS) regression that is used in Mann *et al.* (2007). Riedwyl *et al.* (2008b) compare multivariate PC regression (used by Luterbacher *et al.*, 2004, see also section 'Regional climate reconstructions') and RegEM (with TTLS regression) to reconstruct European summer and winter surface air temperature over the past millennium. They use pseudo-proxy data derived from two climate model simulations (ECHO-G 4 and NCAR CSM 1.4, von Storch *et al.*, 2004 and Ammann *et al.*, 2007b, respectively) to test the sensitivity and performance of the two CFR techniques in a specific virtual experimental set-up.

Lee *et al.* (2008) assessed reconstruction technique performance by constructing pseudo-proxies from these two long climate model simulations and then comparing the reconstructed NH average temperatures with the known values simulated by the climate models, forced with reconstructions of historical anthropogenic and natural external forcing. Tests considered randomly configured 15- and 100-location pseudo-proxy networks selected from the NH locations for which proxies are currently available, so that conclusions as to the relative performance of the different techniques would be independent of decisions made by the practitioner regarding the proxy-network configuration. Pseudo-proxies were constructed using two signal-to-noise ratios (SNRs), and either white noise or mildly persistent red noise. In addition, the study also varied the length of the calibration period somewhat, considered the effect of smoothing the proxies and target temperatures prior to calibration, and considered the effect of including a detrending step prior to calibration. Performance was assessed mostly using decadally smoothed series, though some assessment at annual timescales was undertaken.

Figure 1 displays some of the results obtained by Lee *et al.* (2008). The ranges and medians of root-mean-squared reconstruction errors of hemispheric temperatures from multiple pseudo-proxy realizations, using white-noise (two different SNRs) pseudo-proxies are shown. Figure 1 shows that conclusions regarding the relative performance of the different techniques are not affected by the SNR, size of network or particular choice of millennial climate simulation. Conclusions were also not affected by the length of the calibration period, by the use of red-noise rather than white-noise proxies or if detrending was undertaken prior to calibration (not shown here, but see Lee *et al.*, 2008), although in the case of the latter in particular, the reconstruction quality was reduced in many cases, especially for the MBH method (this issue was also considered by von Storch *et al.*, 2006; Wahl *et al.*, 2006; and Mann *et al*., 2007).

The state–space model and RegEM approaches provided better reconstructions at the annual timescale and for the Northern Hemisphere, than the other techniques evaluated by Lee *et al.* (2008), but most methods provided satisfactory and similar results at decadal timescales. Exceptions were the MBH technique, and the non-smoothed forward regression and non-smoothed variance matching methods. The results comparing multivariate PC regression (also referred to as forward multivariate truncated-EOF regression) and RegEM (implemented with TTLS as described by Mann *et al*., 2007, see section 'A comparison of methods: CPS and CFR') for European summer and winter average temperature reconstructions (Riedwyl *et al*., 2008a,b) show that more skilful results are generally achieved with RegEM, as low frequency variability is better preserved, particularly for summer. However, RegEM showed a noted tendency to introduce strong low-frequency artefacts in the winter average temperature reconstructions at the lowest signal-to-white noise ratios and at the strongest red noise additions examined. Focusing on reconstruction results of the whole climate field, differences between the two CFR techniques are very small and barely detectable (Riedwyl *et al*., 2008a,b). Note, however, that these are general results representative of performance for arbitrarily selected proxy networks. The pseudo-proxy network in Riedwyl *et al.* (2008b) is chosen according to site locations of published European data and series that will be potentially available from current European research projects. Riedwyl *et al*. (2008b) argue that if the techniques already fail, using input data covering the full length of 1000 years (as is assumed in the paper for the pseudo-proxies at these locations), it is certain they will when the number and spatial distribution are reduced and change through time. In this sense, performance for a specific fixed proxy network, such as that used by MBH, was not considered (see section 'Pseudo-proxy experiments to test single time series reconstructions' and also Wahl and Ammann, 2007 and Ammann and Wahl, 2007, for discussion of the robustness of the MBH reconstruction). When analysed with the decadally smoothed real-world palaeoclimate proxy data from Hegerl *et al.* (2007), all CPS methods considered provided almost identical results. This similarity in performance suggests that the difference between many real-world reconstructions are more likely to be due to the choice of the proxy series, their quality, spatial and temporal distribution, or the use of different target seasons or latitudes than to the choice of statistical reconstruction method (see also Rutherford *et al*., 2005; Juckes *et al.*, 2007).

A further extension and optimization of the CPS approach considered by Lee *et al.* (2008) is the use of a Kalman filter and smoother algorithm to incorporate a state equation into the reconstruction paradigm. This approach combines the 'observing equation', that relates the observed proxy composite to the hemispheric mean temperature with a 'state equation' that describes the evolution of the hemispheric mean temperature over time. The latter can be either a simple autoregressive time series model describing the temporal persistence of the hemispheric mean temperature, or something more complex that describes the dynamics of the hemispheric mean and how it responds to external forcing. The advantage of this approach is that it allows explicit representation of the source and nature of the error in the observed proxy composite, together with a representation of the temporal behaviour of the reconstruction target. A potential disadvantage is that reconstructions that incorporate information about the response to estimated forcing histories cannot be considered as truly independent for the purposes of assessing climate model simulations that have been driven by the same (or similar) forcings. Both state-space model approaches provide better reconstructions than existing CPS methods at annual timescales, though differences are insignificant at longer timescales. Lee *et al.* (2008) show that the state-space model approach does not produce substantially different reconstructions when the estimated responses to prescribed external forcing time series are included as extra information (see, for example, Figure 1). Including forcing response terms in the state-space model does, however, allow a simultaneous reconstruction and detection analysis, and also enables an extension that allows forecasts of future climates (given prescribed forcing). Consistent with the results of Hegerl *et al.* (2003, 2007), Lee *et al.* (2008) detect the effects of anthropogenic forcing (greenhouse gas and aerosol forcing combined) and volcanic forcing in real-world palaeoclimatic proxy data.

Pseudo-proxy experiments to test single time series reconstructions

There remains a need for a more systematic exploration of the performance of reconstruction methods under a much wider spread of conditions than has yet been attempted. To further this aim, we outline here a sample protocol for organizing the wide variety of pseudo-proxy tests to systematically investigate the potential performance of climate reconstructions, and then illustrate this protocol with example applications. Our example applications specifically examine the performance of climate reconstruction methods given a fixed proxy network. The test protocol nests the experiments in a hierarchy, starting with temporal orderings, and progressing to calibration periods and noise levels in turn. This hierarchy was developed as an example of a systematic way to organize testing of the kinds of parameters that have been studied so far in pseudo-proxy experiments, and to provide a structure for future studies. The protocol is *not* intended to be prescriptive: modifications and extensions are anticipated as research progresses.

Separate, though very similarly structured, protocols are outlined for testing single time series reconstructions and for testing whole-field reconstructions. The reason for separating tests of single time series from tests of entire fields is to allow a more engineering-oriented testing approach to be taken in the single series case (for both CPS and the spatial-mean of CFR approaches). Engineering-oriented testing is designed to include extreme 'end-member' situations that enable examination of reconstruction performance under both optimal and worst-case conditions beyond what would occur in actual applications. What such end-member cases might be is more easily defined in the single series case. The example protocol for such tests is shown in Figure 2.

This protocol can be used to evaluate many aspects of climate reconstructions; here we illustrate its application by focusing on one crucial issue – how sensitive are reconstruction methods to the extent to which the full climatological range of a variable being reconstructed is represented in the calibration period. In a generalized regression context it is well-established that forecasting (or 'hindcasting' in the case of climate reconstructions) is less well-constrained when the predictor values lie outside the range of data

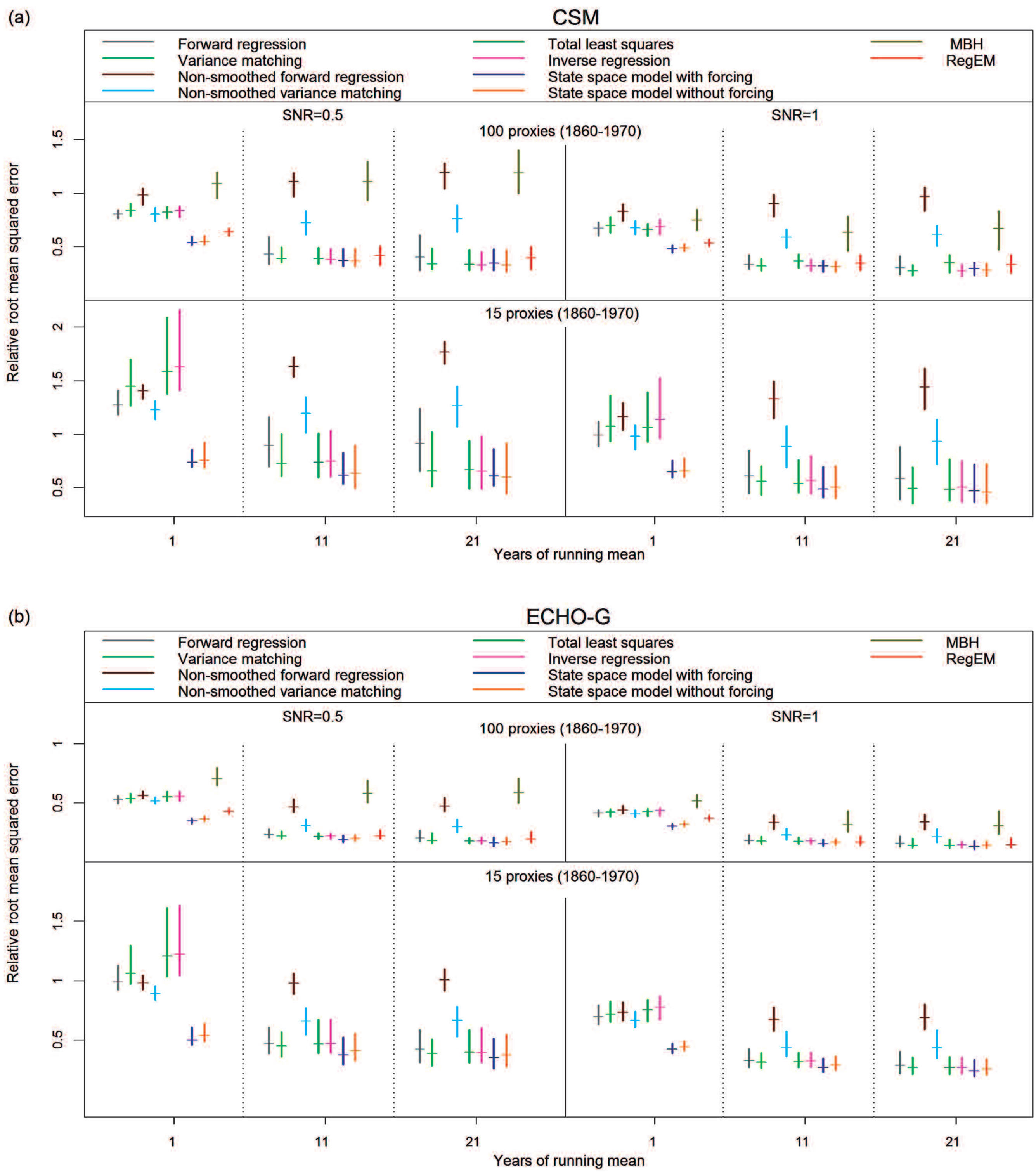

Figure 1 Relative root mean squared error (RRMSE) of the reconstruction error, expressed relative to the variability of the simulated hemispheric temperature. The median RRMSE from multiple realizations of randomly selected pseudo-proxy networks (100 repetitions for the CPS techniques, 40 for the CFR techniques because of the greater computational cost involved) is indicated with horizontal bars and the estimated 5–95% range of the RRMSEs is shown with vertical lines. Results using the 1860–1970 calibration period with different SNRs and varying number of pseudo-proxies are shown for two millennial climate model simulations: (a) CSM and (b) ECHO-G. From Lee *et al.* (2008: figure 2)

employed in calibration. One way to test this sensitivity is to explore specific 'end-member' sets of model simulated climates. Such test sets would use re-orderings of model output by year (or other time step) that proceed from lowest values to highest, from highest values to lowest, and including both highest and lowest values in the calibration period (cf. Figure 2). Re-ordering the model output will severely modify climate persistence and change over multiple years, but few CPS and CFR methods in fact make use of such information. Methods that work with (for example) decadal-mean values could still be tested, by re-ordering the decadally averaged model output. Obviously, data re-ordering does not affect the spatial relationships across all model grid boxes *within* each year or other time step. It should be noted that some reorderings of data will (by design) artificially diminish the range

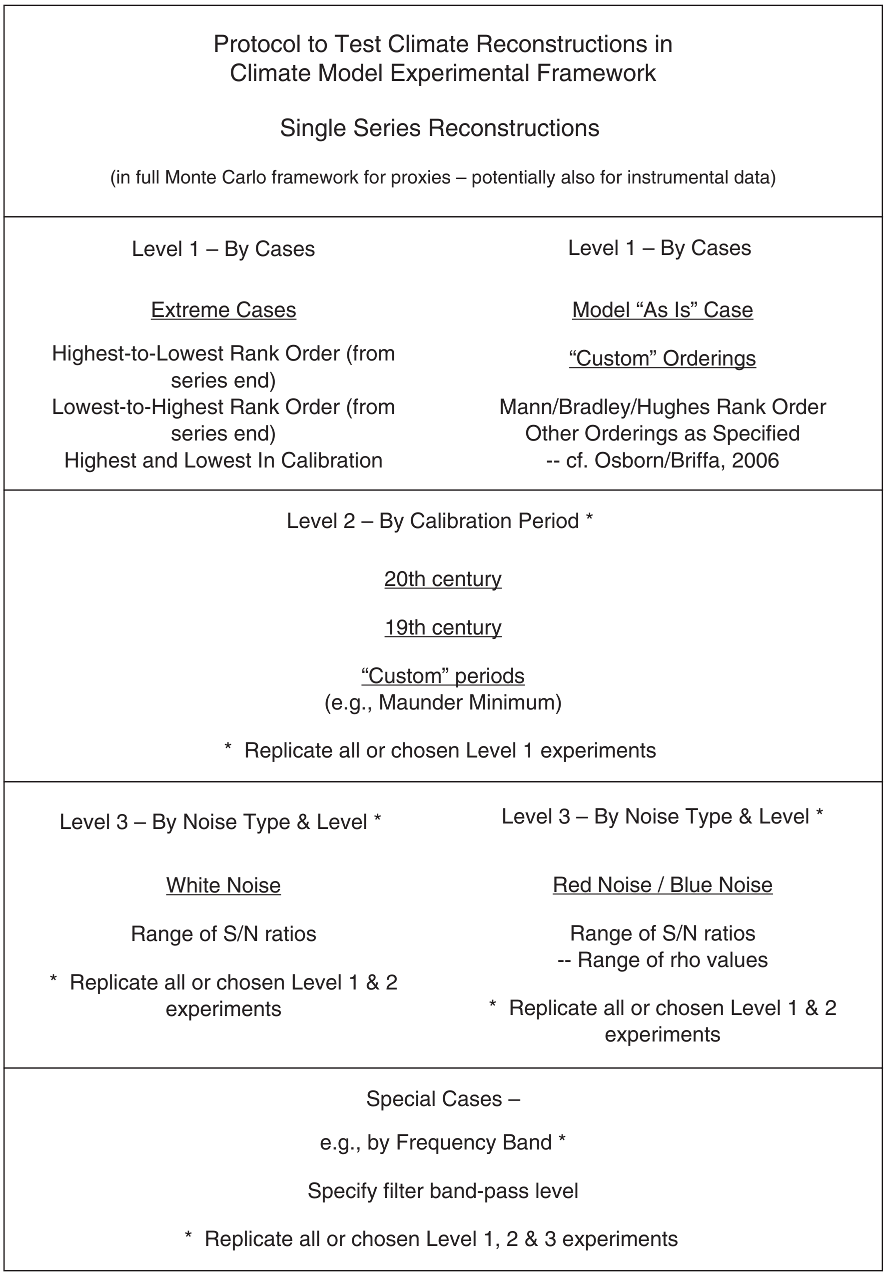

Figure 2 Protocol for testing single-series climate reconstructions in model-based pseudo-proxy experiments

of data variation during the calibration period, thus introducing a far larger extrapolation error than is likely to be encountered in the real world. Results of 'end member' experiments must therefore be carefully interpreted in that context.

Examinations of this kind (Figure 3a,b and further experiments not shown) demonstrate that both the truncated-EOF CFR method used by MBH98 (employing inverse regression) and the simple regression methods that have been widely used by many other palaeoclimate researchers exhibit some degree of sensitivity to the range of climatological information used during calibration. The MBH98 method performs very well in reconstructing NH temperature (Figure 3b) when the full climatological range over the reconstruction period is represented in the calibration period, whereas it systematically loses amplitude the further the range of temperature in the hindcast period moves outside the range occurring within the calibration period (Figure 3a). (Note that a simple CPS reconstruction, not shown, exhibits no systematic loss in this situation.) Other model-based testing in which significant portions of the full climatological range occur outside the calibration period show qualitatively similar results for this version of CFR methodology (von Storch *et al.,* 2006, version without trend removal; Bürger *et al*., 2006, 'MBH98 analogue' version). Bürger *et al*. (2006) show a variant of PC regression that has much less amplitude loss when significant portions of the full climatological range occur outside the calibration period (their version '11110'); note that this variant predicts the mean NH temperature directly rather than the full climate field, using an inverse regression of the pseudo-proxy data on PCs and might be more sensitive to the particular proxy records used than other methods (Juckes *et al*., 2007; their inverse regression method shows such sensitivity and is similar, though not identical, to the Bürger *et al*., 2006, version '11110').

Küttel *et al*. (2007) also find a reduced amplitude reconstruction for the multivariate PC regression CFR method (cf. Luterbacher *et al*., 2004), which has been employed extensively for European regional reconstructions (section 'Regional climate reconstructions'). This loss of amplitude was clearly found to be related to the insufficient predictor network prior to ~1750 and to a much lesser degree to the noise added. Riedwyl *et al*. (2008b) find that

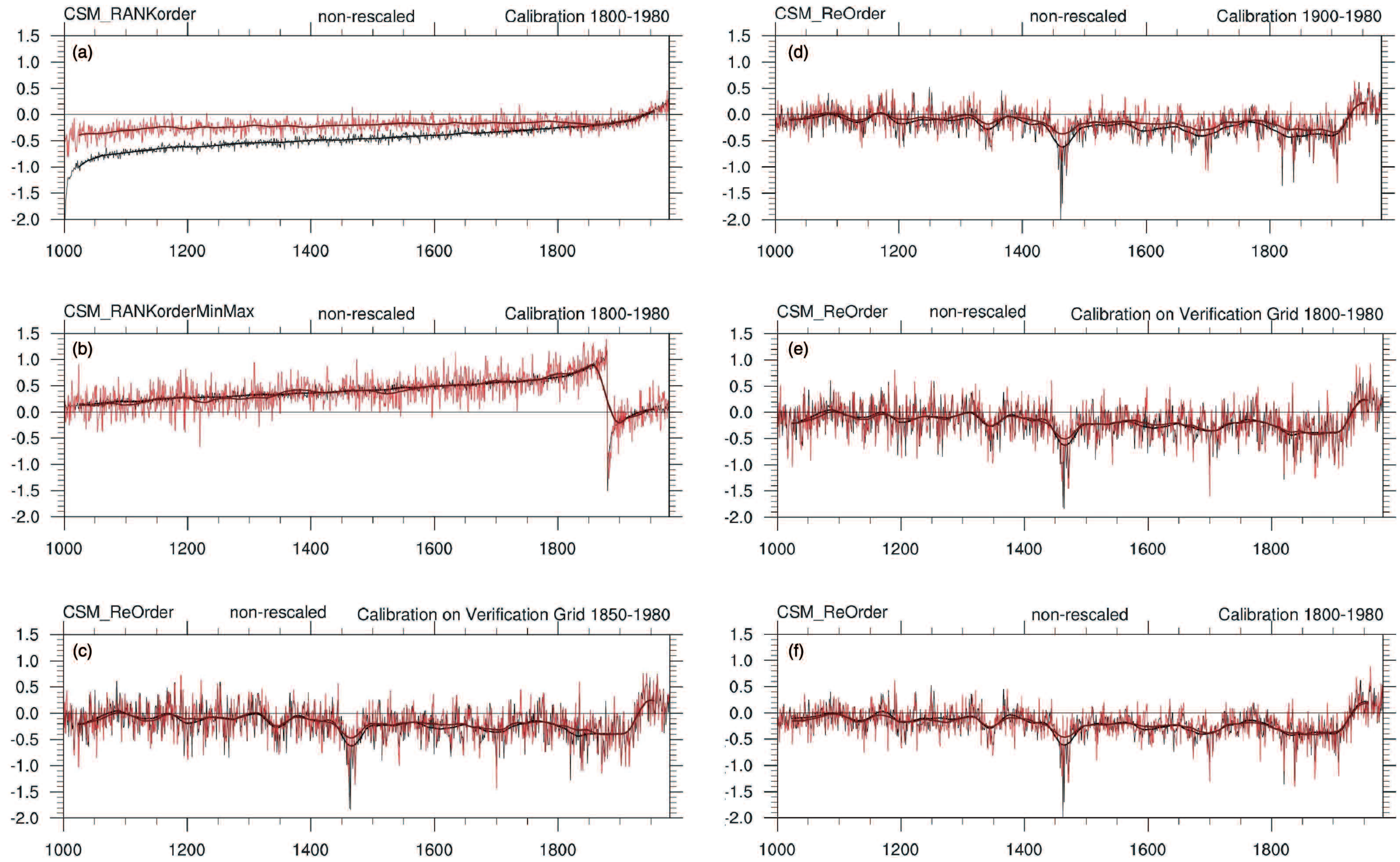

Figure 3 Six examples of single-series pseudo-proxy experiments, illustrating how reconstruction performance depends on the range of temperature variation within the calibration period, calibration period length and spatial coverage. In each example, the black line is the actual model-simulated NH temperature time series of the grid cells used for calibration and the red line is a pseudo-proxy reconstruction using the Wahl and Ammann (2007) emulation of the MBH98 CFR method. Full 1000-yr annual time series and 30-yr smoothed time series are shown: the reference ('zero') line in all graphs is the 1900–1980 mean. In all examples, simulated data are from the NCAR-CSM simulation of Ammann *et al.* (2007b), pseudo-proxies are sampled from the grid boxes and climate variables that match the MBH98 1820–1980 proxy network and have white noise added to give SNR = 0.25 (in terms of relative variances). (a) Simulated temperatures are re-ordered so that NH average surface temperatures progress from coldest to warmest and the 1800–1980 calibration period contains only the warmest temperatures. (b) Simulated temperatures are re-ordered so that the highest and lowest temperatures all occur within the 1800–1980 calibration period. (c)–(f) Simulated temperatures are re-ordered to have the same year-to-year progression as the MBH99 temperature reconstruction, thus emulating a possible real-world situation in terms of the temperature range sampled within the calibration period compared with the range that may have occurred outside the calibration period (cf. Ammann and Wahl, 2007). The differences are that (c) and (d) use calibration periods and spatial coverage that could be implemented using real data (1850–1980 using the sparser MBH98 verification grid and 1900–1980 using the full MBH98 calibration grid, respectively) while (e) and (f) demonstrate the potential improvement that might be achieved if the longer 1800–1980 calibration were possible (using the sparser MBH98 verification grid and the full MBH98 calibration grid, respectively). Note that the rescaling, used by MBH98, so that the variance of the reconstructed PCs matches that of the actual temperature PCs is not employed here and tests indicate that it has little effect on these pseudo-proxy reconstructions

rather high degrees of noise added to the pseudo-proxy signal lead to underestimation of target European average temperature for both this and the RegEM (implemented with TTLS) techniques, though less for RegEM than for multivariate PC regression, particularly for summer temperature. As noted, in this set of experiments, RegEM introduces strong low-frequency artefacts for winter temperature at the highest levels of added white and red noise. Rutherford *et al.* (2005) used the RegEM algorithm to reconstruct global temperature fields. Pseudo-proxy experiments reported by Mann *et al.* (2005) appeared to show that this method did not underestimate the amplitude of the reconstructed NH temperature anomalies. Smerdon and Kaplan (2007), however, show that this may have been a false negative result arising from differences between the implementation of the RegEM algorithm in the pseudo-proxy experiments and in the real-proxy reconstructions (also noted by Lee *et al*., 2008). Mann *et al*. (2007; cf. their figures 3 and 4) demonstrate that a variant of the RegEM method, that uses TTLS rather than ridge regression, produces an NH temperature reconstruction whose amplitude fidelity does not exhibit the calibration interval dependence of the previous implementation by Mann *et al*. (2005), and yields reconstructions that do not suffer from amplitude loss for a wide range of signal-to-noise ratios and noise spectra (though Lee *et al*., 2008, suggest that an appropriately implemented ridge regression can also produce good results). With TTLS as implemented by Mann *et al*. (2007) reconstructing NH temperature, RegEM performs without amplitude loss in model-based tests (versions without trend removal), including using the high-amplitude ECHO-G model output utilized by Bürger *et al*. (2006), von Storch *et al*. (2006) and Küttel *et al*. (2007) to examine truncated-EOF methods. When used to reconstruct the regional-level Niño3 index, however, a longer calibration period improves amplitude fidelity in these experiments (Mann *et al*., 2007: auxiliary material, section 9).

A custom re-ordering of the data to follow the trajectory of the MBH99 reconstruction of NH temperature (Figure 3c, d) can be examined to determine if this theoretical potential for systematic amplitude loss significantly impacts the MBH98/99 result in a model simulating its real-world context (Ammann and Wahl, 2007). The proxy instrumental evidence available to MBH98/99 suggests that the coldest extended period during the 1000–1980 reconstruction time span occurred during the nineteenth century and the warmest temperatures occurred during the latter twentieth century (Ammann and Wahl, 2007). Thus, such a custom re-ordering allows direct examination of whether extending the calibration period from 1902–1980 to ~1850–1980 would make a difference in terms of amplitude fidelity in the real-world instrumental and proxy data setting actually faced by MBH98. Figure 3d demonstrates that there *is* potential for a relatively small absolute amplitude loss in the situation actually faced by MBH98 (using pseudo-proxies with signal-to-noise and climate sensitivity characteristics similar to the real proxies actually employed) when calibration is limited to the twentieth century. Figure 3c suggests that amplitude loss could be much reduced by extending the calibration 50 years earlier (to cover the period 1850–1980) to include much better sampling of both ends of the full climatological range of NH temperatures: systematic amplitude loss is *eliminated* in parallel extension of the calibration period with the multivariate PC regression method (not shown). Such an extension of the calibration period in the real world means that the spatial coverage of the instrumental data used in calibration would be restricted by a factor of nearly five times in the MBH context, to the MBH98/99 verification data set. The question of whether the proxies used by MBH98/99 (or other reconstruction studies) were themselves subject to low-frequency amplitude limitations is not at issue in these experiments, which only focussed on the reconstruction algorithms *per se*.

The outcomes shown in Figure 3c, d are important for further consideration of reconstruction optimization in the truncated-EOF CFR method. Together they suggest that *ensuring maximum inclusion of the full climatological range of a time series being reconstructed during calibration is more critical for amplitude fidelity than the spatial coverage of the instrumental data being used* (cf. Ammann and Wahl, 2007), as long as the instrumental data field is appropriately sampled, which appears to be so in the MBH98/99 case. This conclusion is reinforced by the outcomes shown in Figure 3e, f. The only difference between couplets (c/e) and (d/f) in Figure 3 is that experiments (e) and (f) are calibrated over the longer period used for experiments (a) and (b), 1800–1980. In these cases, the MBH-style truncated-EOF reconstructions show very small (e) and essentially no (f) average amplitude loss (with the exception of the sharp later-fifteenth century cooling). An additional impact of using the spatially reduced verification grid (e) for calibration shows up as a somewhat enhanced overall variation in the high frequency domain, which is expected from sample richness considerations (cf. model annual 'target' data, the black line in Figure 3e). An interesting extension of these experiments would be to include a replication of the period ~1450–1475 within the calibration period, to include the lowest temperatures over the entire model period in calibration. The outcome shown in Figure 3b suggests that in this case, reconstruction of the later-fifteenth century cold period would not be subject to any average amplitude loss. In the real world, extending the length of the calibration data set backwards in time necessarily means limiting the coverage of the instrumental field being reconstructed (eg, to Europe in the late-eighteenth century), so doing this is clearly undesirable when the goal is whole-field reconstruction. For the truncated-EOF method this situation would imply the need to examine possible tradeoffs between the potential for amplitude loss and the spatial coverage of reconstructed fields. The RegEM method appears to avoid this trade-off at the scale of the NH temperature field; and in fact, Mann *et al*. (2007) show that using RegEM with the same real-world proxy data employed by MBH98 results in a nearly identical reconstruction to the MBH98 original, suggesting that the general *potential* for amplitude loss in the truncated-EOF method noted here may have had no effect on the actual MBH outcome.

These considerations represent examples of Levels 1 and 2 in the test protocol (Figure 2), in terms of the ordering of the model output examined and then varying the calibration period. A third level would be to generalize consideration of the noise level(s) and noise model(s) being used to generate the pseudo-proxies (c.f., Mann *et al*., 2007; Lee *et al*., 2008; Riedwyl *et al*., 2008b). Other potential methods for generating pseudo-proxies might eliminate stochastic models altogether, instead using mechanistic approaches to generate tree increment growth from model output (cf. Vaganov, 1996; Fritts *et al*., 1999) for tree-ring proxy sites. Finally, other kinds of analyses might be employed, including frequency decompositions designed to examine whether specific frequency bands are better reconstructed than others.

Pseudo-proxy experiments to test whole-field reconstructions

A similar example protocol for testing whole-field reconstructions is outlined in Figure 4, but with an important difference in Level 1. In the case of whole-field reconstructions, end-member sets and custom orderings of specific large-scale averages are not employed, since the relationship between such artificial cases and field responses is not as well-defined as in the case of examining large-scale averages. Instead, Level 1 in this situation focuses on the fidelity of field reconstructions at times of specific interest in relation to anomalies of climate forcings (see section 'Climate forcing and histories') and in relation to how climate forcings and modes of climate variability relate to each other. It is recognized

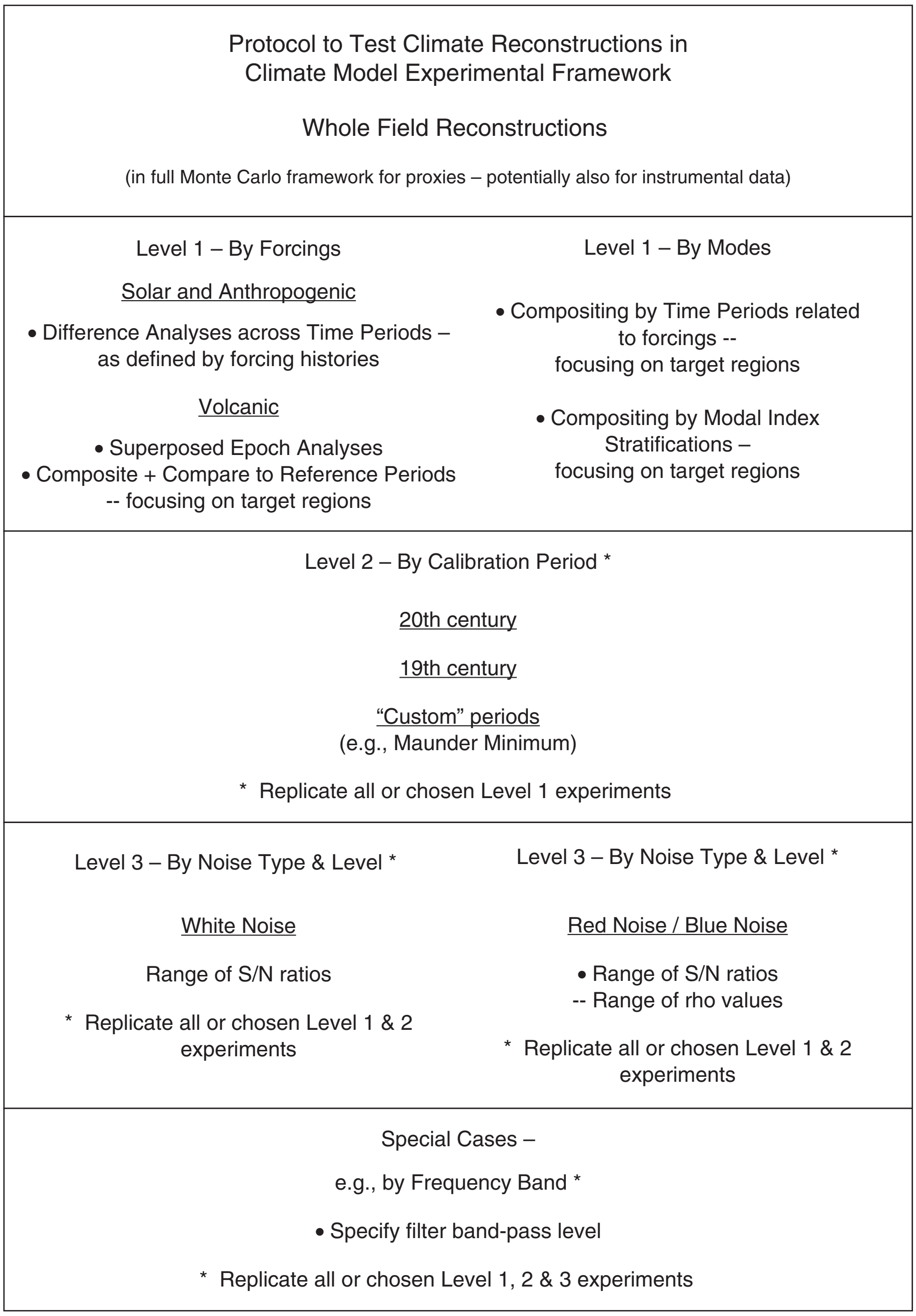

Figure 4 Protocol for testing whole field climate reconstructions in model-based pseudo-proxy experiments

that organizing Level 1 examinations in such a context is potentially problematic, as a given model may not respond in dynamically/spatially realistic ways to climate forcings (and with differing degrees of realism to different forcings), or it may not exhibit realistic modes of internal variability (eg, ENSO). However, an unrealistic response to climate forcing changes by a model does not mean, *ipso facto*, that good fidelity between a pseudo-proxy reconstructed field and the corresponding model field is of little meaning (cf. section 'Introduction to the pseudo-proxy approach').

In the context of these caveats, the example protocol suggests using various kinds of analyses that either target composite differences across time periods of interest in relation to forcing histories, or that target the effects of specific forcings (eg, superposed epoch analyses of volcanic forcing events; cf. Adams *et al*., 2003; Fischer *et al*., 2007). Then, the further dimensions of altering calibration periods, noise type and level, and other examinations of specific interest are introduced, as in the protocol for testing reconstructions of individual time series in Figure 2. As in the individual time series case, the protocol for whole-field reconstructions is not intended to be paradigmatic.

An example of testing model pattern fidelity is shown in Figure 5, for whole-field reconstructions of winter temperature anomalies over Europe induced by large tropical volcanic eruptions. In this examination, the climate model itself (NCAR-CSM 1.4; Ammann *et al*., 2007b) exhibits a spatial pattern quite similar to that reported by Fischer *et al*. (2007), based on a combination of instrumental, documentary and other proxy information (Luterbacher *et al.,* 2004). The pseudo-proxy performance of the MBH98-style inverse regression truncated-EOF CFR in the model context is quite good in terms of capturing the overall spatial pattern of the forcing effect, but it significantly underestimates the amplitude of response. This amplitude loss is, at least in part, related to the amount of noise added to create the pseudo-proxies (not shown). Notably, in this example, the loss of amplitude for

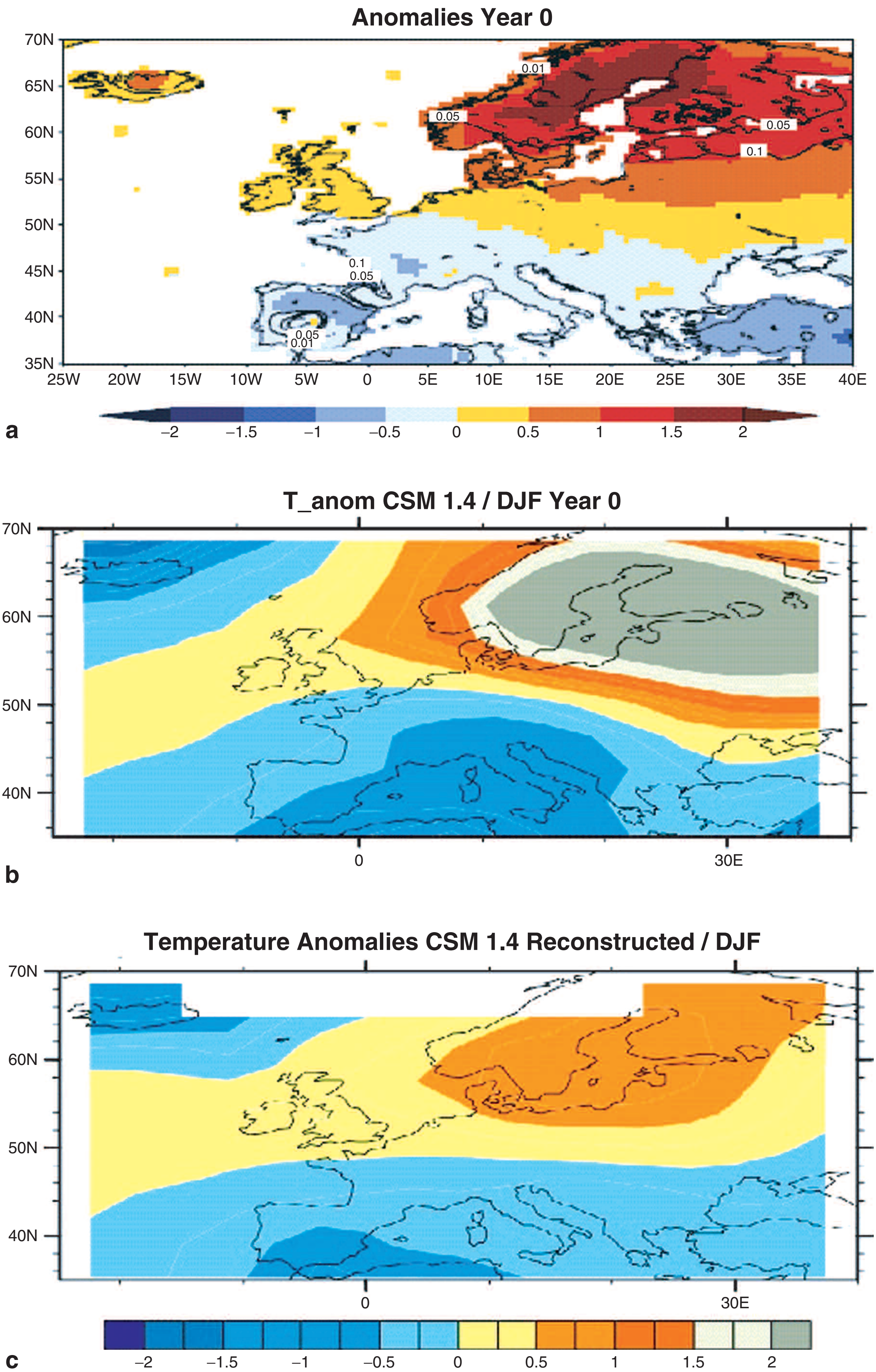

Figure 5 Example of field reconstruction performance: composite winter temperature anomalies for Europe after 15 tropical volcanic events over the past 500 years (cf. Fischer *et al*., 2007). (a) Anomalies reconstructed from instrumental, documentary and proxy records using a truncated-EOF method (multivariate PC regression, Luterbacher *et al*., 2004). (b) Anomalies from NCAR-CSM model (Ammann *et al*., 2007) over the same time period. (c) Pseudo-proxy reconstructed anomalies in same model context as (b), using the Wahl and Ammann (2007) emulation of the MBH98 truncated-EOF inverse regression method

the positive anomaly focused on Fennoscandia and northwestern Russia might be sufficient to reverse the sign of the European-mean composite response to volcanic forcing in reasonably conceivable reconstruction situations (eg, low-enough proxy SNR and/or short-enough calibration period). This indicates that for some situations (eg, where a spatial-mean represents the average of two similar-magnitude but opposite sign anomalies) reconstructing climate field anomaly *patterns* may be a more robust overall capacity of the truncated-EOF CFR methods than reconstructing spatially averaged climate. The level of noise does not have a direct relationship with loss of amplitude fidelity in a whole-field reconstruction simulation undertaken with the RegEM method for the post-Tambora global temperature field (Mann *et al*., 2007) (if anything the relationship may vary inversely), although additional noise does have a detrimental effect on the ability to resolve fine details of the field.

Finally, we note that quantitatively measuring the significance of differences between model and reconstructed fields (and this is relevant both to the evaluation of pseudo-proxy experiments and to the comparison of models with real-world reconstructions) raises a number of issues that apply more generally to measuring (and determining the significance of) differences in ≥2 dimensions. Statistics such as Reduction of Error (RE) and Coefficient of Error (CE) (eg, Cook *et al*., 1994; MBH98; Mann *et al*., 2007) and the Mann-Whitney Rank Sum Test (eg, Fischer *et al*., 2007) have been used for this purpose, both at the level of the entire field and on a grid-box by grid-box basis. However, such tests are not able to capture the potential differential importance of parts of a reconstructed field in terms of sensitivity to the direct effects of forcings and the indirect effects of forcing-induced dynamical changes. It is beyond the scope of this paper to address this issue in detail, but it is an important area of analysis that needs significant attention in order to more fully operationalize the quantitative examination of field reconstruction fidelity.

Other considerations for use of model-based experiments

Model-based examinations make possible the creation of 'perfect-pseudo-proxy' cases, ie, situations in which the simulated proxy values for a given grid box have no noise added and thus by definition have SNRs of ∞. In this kind of examination, once successful performance is demonstrated for a reconstruction methodology by an examination similar to the one shown in Figure 3b (ie, when the full range of reconstructed variables is present in the calibration period), the question being addressed reduces to fidelity losses driven solely by the spatial coverage of the proxies. A combination of noisy-proxy and perfect-proxy analyses with the custom temporal ordering of model output in Figures 3e, f could be used to separate the extent to which reductions of proxy richness are related to the generation of amplitude loss *and* the extent to which noise hastens the onset of this problem (ie, the extent to which noise acts as a *de facto* reduction of proxy richness). Examinations of this kind have already been started with the RegEM CFR technique, showing that even a very reduced proxy set leads to no low frequency amplitude loss for reconstruction of the NH temperature series, although higher frequency variation is significantly increased in this case (cf. Mann *et al*., 2007).

In a related vein, pseudo-proxy experiments can be used to explore how the addition of one or more specific proxies can affect reconstruction quality. Küttel *et al*. (2007), for example, evaluated the impact of adding a single pseudo-proxy in northern Finland on the amplitude fidelity for reconstruction of European average winter surface temperatures. Their results show a strong enhancement of reconstruction amplitude by addition of the single predictor in this previously poorly reconstructed subregion, suggesting that it would be an excellent investment of resources to attempt to develop real-world 'winter' proxy data from this area. Historical archives from southern Sweden (eg, Leijonhufvud *et al.,* 2008) or from the eastern Baltic (Tarand and Nordli, 2001) might provide some of the documentary records being extended back into the early sixteenth century, allowing the future development of a southern Scandinavian or eastern Baltic winter temperature reconstruction for the last approximately 500 years.

It is likely that many more applications may benefit from testing within potentially realistic surrogate climates provided by GCM simulations of recent centuries and millennia. Gonzalez-Rouco *et al.* (2006), for example, used the pseudo-proxy method to evaluate the recovery of surface temperature variations from vertical profiles of ground temperature anomalies (ie, pseudo-borehole temperatures) that were themselves generated using models of vertical heat diffusion driven by GCM-simulated surface temperatures. Their results indicate that the current distribution of borehole temperature records may be sufficient to provide useful estimates of surface temperature change over recent centuries, though with expected attenuation at the shorter timescales during earlier times (e.g. Gonzalez-Rouco *et al.*, 2008 and references therein). The impact of different sources and magnitudes of noise, the possible underestimate of Mediaeval warmth in situations with strong cooling between Mediaeval and 'Little Ice Age' periods (see Gonzalez-Rouco *et al.*, 2006: figure 3c), and the precision with which the timing of maximum 'Little Ice Age' cooling can be estimated (again, see their figure 3c), all deserve further investigation.

In addition to providing opportunities for testing statistical reconstruction methods, information from climate model simulations can be combined, via data assimilation techniques, with information derived from proxy records, to yield improved estimates of past climate change. This approach has proved to be very successful for 'reanalysis' of atmosphere and ocean states over the last 60 years. The few studies performed so far for the pre-instrumental period have been mainly devoted to the testing of particular techniques (Goosse *et al.*, 2006a) or to specific short periods (van der Schrier and Barkmeijer, 2005). Nevertheless, these initial studies have shown the potential advantages of combining information from forcings, models and data in this way, to yield physically consistent reconstructions of variables that are difficult to estimate on the basis of proxy records alone, such as the large-scale oceanic and atmospheric circulation. Data assimilation exercises should be pursued further, with a focus on the last millennium, to adapt and test assimilation techniques to the specific problems that arise in the palaeoclimate context, such as coarse time resolution and large uncertainties in proxy and forcing data. This approach should thus be considered as a valuable complement to statistical reconstruction of past climate.

Climate forcing and histories

Forcing and climate models

An important reason for improving climate reconstructions of the past few millennia is that these reconstructions can help both evaluate climate model responses and sharpen understanding of important mechanisms and feedbacks. Therefore, a parallel task to improving the reliability of proxy climate data and climate reconstructions is to assess and independently constrain forcings of the climate system over that period.

Forcings can generically be described as external effects (also called exogenous effects in other modelling communities, such as economics) on a specific system. Responses within that system that also have an impact on its internal state are described as feedbacks. For the atmosphere, sea surface temperature changes could therefore be considered a forcing, but in a coupled ocean–atmosphere model they could be a feedback to another external factor or be intrinsic to the coupled system. Thus the distinction between forcings and feedbacks is not defined *a priori*, but is a function of the scope of the modelled system. This becomes especially important when dealing with the biogeochemical processes in climate that affect trace gas concentrations (CO_2 and CH_4) or aerosols. For example, if a model contains a carbon cycle, then the CO_2 variations as a function of climate will be a feedback, but for a simpler physical model, CO_2 is often imposed directly as a forcing from observations, regardless of whether in the real world it was a feedback to another change, or a result of human industrial activity.

It is useful to consider the pre-industrial period (pre-1850) separately from the more recent past, since the human influence on many aspects of atmospheric composition has increased dramatically in the twentieth century. In particular, aerosol and land-use changes are poorly constrained prior to the late-twentieth century and have large uncertainties. Note, however, there may be a role for human activities even prior to the nineteenth century owing to early agricultural activity (Ruddiman, 2003; Goosse *et al*., 2006b).

In pre-industrial periods, climate forcings can be separated into purely external changes (variations of solar activity, volcanic eruptions, orbital variations) and those that are intrinsic to the Earth system (greenhouse gases, aerosols, vegetation, etc.). Such changes in Earth system elements will occur predominantly as feedbacks to other changes (whether externally forced or simply as a function of internal climate 'noise'). In the more recent past, the human role in affecting atmospheric composition (trace gases and aerosols) and land use have dominated over natural processes and so these changes can, to large extent, be considered external forcings as well.

Traditionally, the 'system' that is most usually implied when talking about forcings and feedbacks are the 'fast' components of the atmosphere–land, surface–upper ocean system that, not coincidentally, correspond to the physics contained within atmospheric general circulation models (AGCMs) coupled to a slab ocean. What is not included (and therefore considered as a forcing according to the previous definition) are 'slow' changes in vegetation, ice sheets or the carbon cycle. In the real world these features will change as a function of other climate changes, and in fact may do so on relatively 'fast' (ie, multidecadal) timescales. Our choice then of the appropriate 'climate system' is slightly arbitrary and does not give a complete picture of the long-term Earth system sensitivity.

These distinctions become important because the records available for atmospheric composition do not record the distinction between feedback and/or forcing. They simply give, for instance, the history of CO_2 and CH_4 (though for the industrial era, good estimates of CO_2 emissions are available, based on fossil fuel consumption). Depending on the modelled system, those records will either be a modelling input, or a modelling target.

While there are good records for some factors (particularly the well-mixed greenhouse gases such as CO_2 and CH_4), records for others are either relatively uncertain in magnitude (tropospheric and volcanic aerosols, and solar activity), incomplete (dust, vegetation) because of poor spatial or temporal resolution or non-existent (eg, ozone). Estimates of the magnitude of these latter forcings can only be made using a model-based approach. This can be undertaken using GCMs that include more Earth system components (interactive aerosols, chemistry, dynamic vegetation, carbon cycles, etc.), but these models are still very much a work in progress and have not been used extensively for palaeoclimatic purposes. Some initial attempts have been made for selected feedbacks and forcings (Gerber *et al*., 2003; Goosse *et al*. 2006b) but a comprehensive assessment over the millennia prior to the pre-industrial does not yet exist.

Even for those forcings for which good records exist, there is a question of how well they are represented within the models. This is not so much of an issue for the well-mixed greenhouse gases (CO_2, N_2O, CH_4) since there is a sophisticated literature and history of including them within models (though some aspects, such as minor short-wave absorption effects for CH_4 and N_2O are still not universally included; Collins *et al*., 2006).

Individual climate forcings

In this section, we discuss the major external forcing factors that are important for the climate system and its palaeoclimatic modelling. Human-caused changes to the atmosphere (well known for greenhouse gases, but markedly less so for aerosols) are fully discussed in detail by Forster *et al*. (2007). We later illustrate how the various forcings are prescribed and/or determined by one particular A/OGCM.

Solar irradiance

The most straightforward way of including solar irradiance effects on climate is to change the solar 'constant' (more accurately described as total solar irradiance – TSI). However, observations show that solar variability is highly dependent on wavelength with UV bands having about ten times as large amplitude changes than TSI over a solar cycle (Lean, 2000). Thus including this spectral variation for all solar changes allows for a slightly different behaviour (eg, larger solar-induced changes in the stratosphere where the UV is mostly absorbed). Additionally, the changes in UV affect ozone production in both the stratosphere and troposphere, and this mechanism has been shown to affect both the total radiative forcing and dynamical responses (Haigh, 1996; Shindell *et al*., 2001, 2006). Within a chemistry–climate model this effect would potentially modify the radiative impact of the original solar forcing, but could also be included as an additional (parameterized) forcing in standard GCMs.

Reconstructions of solar variability in past millennia are usually based on cosmogenic isotopes whose production is modulated by solar magnetic activity and whose concentrations can be found in ice cores (ie, ^{10}Be) or tree rings (^{14}C) (eg, Bard *et al*., 2000; Solanki *et al*., 2004; Muscheler *et al*., 2005). Calibration of these archives is based on the ~30 year series of satellite observations of solar irradiance over three solar cycles and more recent observations of the spectral character of that variability. The main uncertainty is the magnitude of any long-term secular trend in solar forcing that is unconnected with the direct effects of sunspots and faculae (Foukal *et al*., 2006) and the potential for climatic contamination of the proxy archives (Field, C.V. *et al*., 2006).

It has also been argued that an indirect effect of solar magnetic variability on the shielding of cosmic rays may affect the production of cloud condensation nuclei (Dickinson, 1975). There have been no quantitative calculations of the magnitude of this effect (which would require a full study of the relevant aerosol and cloud microphysics), and so its impact on climate has yet to be included.

Explosive volcanicity

Large volcanic eruptions produce significant amounts of sulphur dioxide (SO_2). If this is injected into the tropical stratosphere during a particularly explosive eruption, the resulting sulphate aerosols can persist in the atmosphere for a number of years (eg, Pinatubo in 1991). Less explosive, but more persistent eruptions (eg, Laki in 1783) can still affect climate though in a more regional way and for a shorter period (Oman *et al*., 2005). These aerosols have both a shortwave (reflective) and longwave (absorbing) impact on the radiation and their local impact on stratospheric heating can have important dynamical effects. It is therefore better to include the aerosol absorber directly in the radiative transfer code. However, in less sophisticated models (eg, Crowley, 2000) or experiments (eg, von Storch *et al*., 2004), the impact of the aerosols has been parameterized as an equivalent decrease in TSI.

Reconstructions of the volcanic forcing history are based on well-dated sulphate layers in ice cores combined with calibration from historically observed eruptions. Magnitudes of effects, hemispheric distribution and details of the aerosol microphysics that might be unique to each eruption are the main sources of uncertainty (Naveau and Ammann, 2005). Combining information from multiple ice-core records, especially from both polar regions, is important to distinguish between local and large-scale events. Recent work (Gao *et al.*, 2008) has attempted to further improve both the accuracy and detail of the estimated forcing history by utilizing sulphate records from 54 ice cores (more than double that used by previous studies). Not only do the additional records reduce the overall uncertainty, but they also allow information from the patterns of aerosol deposition to be used in conjunction with analysis of seasonally dependent atmospheric transport and deposition to estimate stratospheric sulphate loading from ice-core deposition. The resulting data set provides estimates of sulphate loading as a function of latitude, altitude and month. This level of detail, though subject to a number of caveats and uncertainties, is nevertheless useful for forcing the next generation of GCM simulations.

Land surface characteristics and aerosols

Land-cover and land-use changes have occurred both due to deliberate modification by humans (deforestation, imposed fire regimes, agriculture) as well as a feedback to climate change (the desertification of the Sahara *c*. 5500 yr ago). Changing vegetation in a standard model affects the seasonal cycle of albedo, the surface roughness, the impact of snow, evapotranspiration (through different plant rooting depths), etc. However, modelling of the yearly cycle of crops, or incorporating the effects of large scale irrigation, are both still very much work in progress.

Anthropogenic aerosol effects (through industrial emissions or biomass burning) are a critical feature of modern climate change. Increases in reflective aerosols (sulphates and nitrates), as well as absorbing aerosols (black carbon) have direct radiative effects as well as indirect impacts on cloud formation and lifetime as well as snow albedo. There are natural components to these aerosols as well (sulphates via planktonic emission of dimethyl sulphide, black carbon from natural fires, etc.) but aerosol changes over the few millennia prior to industrialization are very poorly constrained. Changes might have arisen from climatically or human-driven changes in dust emissions, ocean biology feedbacks on circulation change or climate impacts on the emission of volatile organics from plants (which also have an impact on ozone chemistry). Some work on modelling a subset of those effects has been undertaken for the last glacial maximum or the 8.2 kyr event (LeGrande *et al.*, 2006), but there have been no quantitative estimates for the late Holocene.

Combining natural and anthropogenic forcings in GCM simulations

Owing to the relative expense of doing millennial simulations with state-of-the-art GCMs, existing simulations have generally only included the minimum required to incorporate relevant solar, GHG and volcanic forcings (Jansen *et al.*, 2007: table 6.2 identifies which forcings were used in most of the recent modelling studies). Progress can be expected soon on more sophisticated treatments of those forcings and the first quantitative estimates of additional effects.

The combined impact of the individual natural and anthropogenic forcings described in the previous sections – together with other important forcings that we have not considered in detail here, such as orbital changes and anthropogenic tropospheric sulphate aerosols – can be estimated either directly or indirectly. Direct estimates rely on calculations of the impact of each mechanism on the radiation balance of the Earth (eg, functions relating greenhouse gas concentrations to radiative forcing), and these are then combined and provided as input to some model simulations – especially for simpler models (eg, Hegerl *et al.*, 2003), because their formulation requires this. The combined forcing can be estimated indirectly by forcing GCMs with, for example, concentrations or emissions of the relevant forcing agent, and then diagnosing the modification to the radiation balance that the radiative transfer component of the GCM predicts. Figure 6 shows one such example, for the HadCM3 simulation of the last 250 years (Tett *et al.*, 2007). In this case, for example, rather than prescribing the volcanic forcing (eg, as an equivalent reduction in solar irradiance as assumed by von Storch *et al.*, 2004), time series of volcanic aerosol loading were applied (with variations amongst four latitudinal bands to allow for the different effects of high-latitude and tropical eruptions). This approach is recommended for future

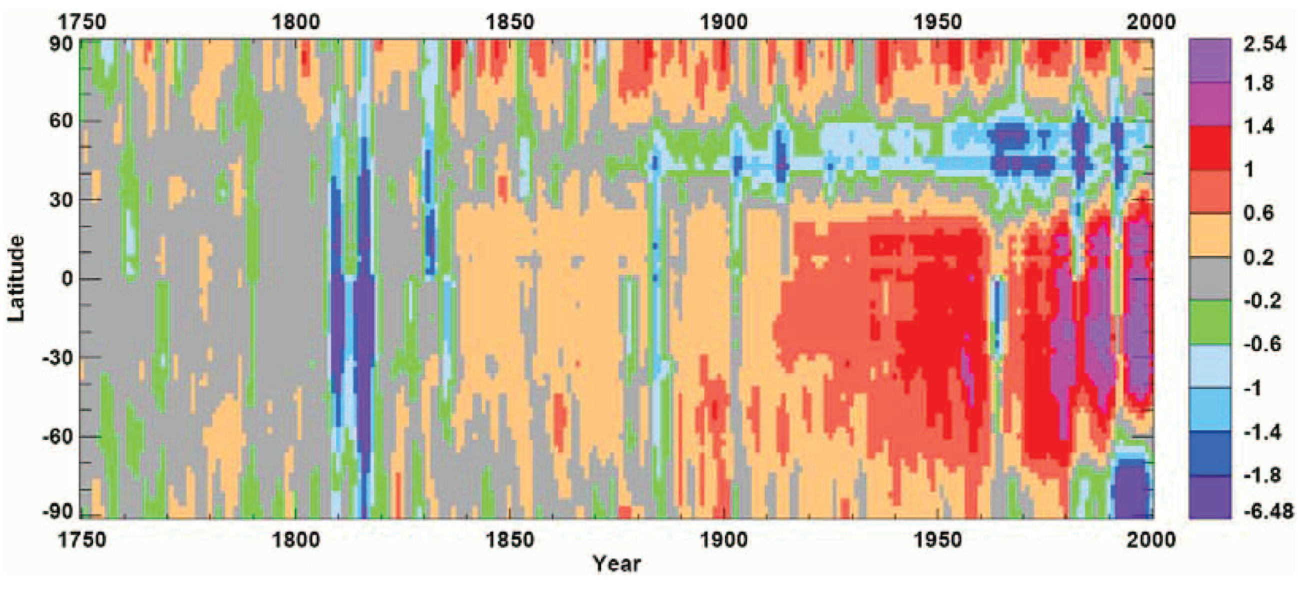

Figure 6 The latitude–time evolution of zonal-mean forcing (W/m^2, relative to the mean of the first 50 years, 1750–1799) diagnosed from the ALL250 simulation with HadCM3 described by Tett *et al.* (2007). This simulation was forced by variations in solar irradiance, volcanic aerosol loading, orbital forcing, land use, sulphate aerosol emissions and concentrations of well-mixed greenhouse gases and ozone. Values are smoothed in the time dimension (by a 5-yr low-pass filter) but not in the latitude dimension

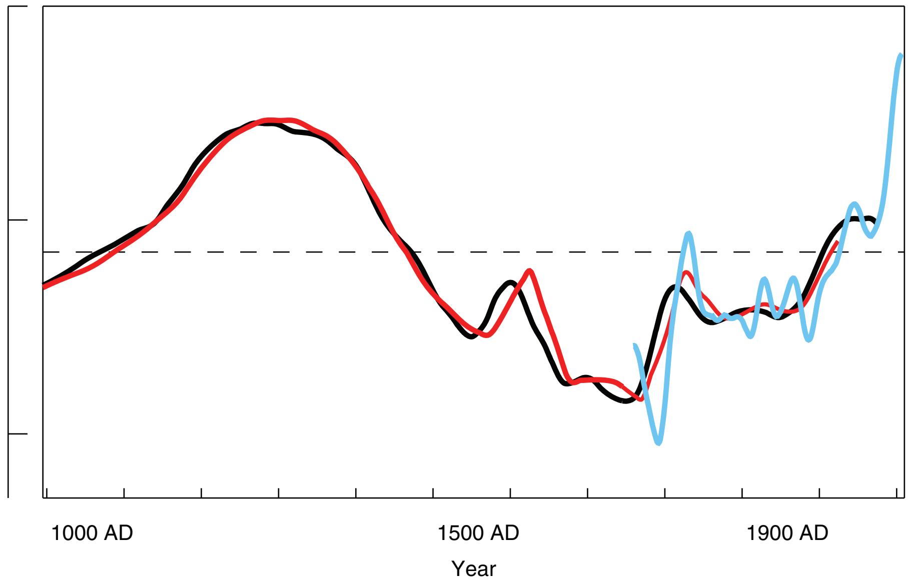

Figure 7 The black curve and the x- and y-axes are a redrawn version of figure 7.1c from the First Working Group of IPCC Report (Folland *et al*., 1990). The y-axis originally had only unnumbered tick markings and was labelled ‘temperature scale’. The red curve is from Lamb (1982: figure 30, the upper (annual) curve). The amplitude of this curve has been scaled to correspond to that of the black curve. The Lamb (1982) time series does have an explicit temperature scale, and the best-fit scaling between this curve and the IPCC curve indicates that one tick-mark interval on the IPCC figure corresponds almost exactly with 1°C. The degree of smoothing for both these curves is unknown, but Lamb (1982) states that the red curve is based on 50-yr means (supported by earlier publications). The blue curve is a smoothed version of the annual instrumental Central England Temperature record from Manley (1974, updated) including the last complete year of 2007. This has been smoothed with a 50-yr Gaussian weighted filter with padding (Mann, 2004). The blue curve is plotted with the same scaling as used for the red curve, further supporting the conclusions that the red curve is based on the same data after the start of the instrumental record in 1659. The red and blue curves illustrate the differences that can occur between a filtered curve and one composed of non-overlapping 50-yr averages, and also that recent measured warming may be comparable with presumed earlier warmth

experiments, because more complex interactions with atmospheric radiative transfer can be simulated, such as the stratospheric warming caused by absorption of both short- and long-wave radiation by the sulphate aerosol which might have dynamical responses in the atmosphere.

This particular example (Figure 6, from HadCM3) also highlights the potential importance of strong geographic variations in forcing. The short-term impact of volcanic eruptions is clear, influencing most latitudes, and there is a long-term trend towards positive forcing arising mostly from greenhouse gas forcing (with a small contribution from solar irradiance changes). The forcing from sulphate aerosol emissions into the mid-latitudes of the Northern Hemisphere, while rather uncertain in its magnitude, is strong enough in this implementation to dominate the greenhouse gas forcing throughout the simulation (even in the late-nineteenth century, as pointed out by Tett *et al.*, 2007), though the negative forcing begins to reduce in recent decades. The negative forcing from Antarctic stratospheric ozone depletion is also clear in the last decade of this simulation. Forcings may have important seasonal as well as geographic structure; Goosse *et al.* (2006b), for example, suggest that land-use changes (with a minor influence from orbital changes) might have contributed to warming during the Mediaeval period (relative to the present day) but only in summer and only for some regions in the mid- and high-latitudes of the Northern Hemisphere.

Conclusions

This article has reviewed the characteristics and current research status of documentary and high-resolution proxy climatic sources and addressed the various approaches by which they may be combined to provide field reconstructions or large-area averages. We have also extensively discussed the use of climate model simulations and how they can inform the debate and take forward research into the optimal combination and interpretation of the various data. Finally, we summarize our principal findings and recommendations for the continued exploitation of palaeoclimatological data for reconstructing large-scale averages and spatial patterns of climate variability over recent millennia. These recommendations are ordered according to the section order in the review:

(1) In the area of tree-ring research the number of available long chronologies is expanding but remains small, and although potential sites with known subfossil data are limited, the effort in developing these is fully justified: there remain large areas of the terrestrial world where chronology network development is in its infancy (much of the lower-latitudes and virtually all of the SH).

(2) It was widely believed 20 years ago that cross-dating trees in tropical regions was not possible. Recent work has led to a growing number of cross-dated chronologies being developed, principally in southeast Asia in regions of marked seasonality in rainfall. It is important this work continues.

(3) There have been recent and continuing improvements in statistical methods for producing long chronologies, which can be shown to retain low-frequency climatic variability more realistically than was previously the norm. Work is needed to further assess the applicability of these methods in a wider range of situations than have been explored to date.

(4) Significant continued effort is required to provide the high degree of intrasite and importantly intersite (ie, regional scale) sample replication needed to demonstrate reliable long-timescale chronology expression. Local and regional-average chronologies and regression-based estimates of climate variability should be routinely presented with explicit indications of their separate timescale and time-dependent confidence limits.

(5) There is pressing need for further study of the likely precedence and causes of the apparent ‘divergence’ between instrumentally recorded and some dendroclimatically estimated temperature trends (typically some high-latitude NH regions) in

recent decades. This emphasizes the priority requirement for systematic updating of many existing tree-ring data, to continue in parallel with efforts to expand the representation of data into new areas.

(6) In coral proxy climate research, replication of important single coral $\delta^{18}O$ and Sr/Ca records would allow quantification of signal *versus* noise in coral reconstructions. Owing to the rarity of long coral cores, however, this may only be possible over the late-twentieth century.

(7) There should be greater effort to increase the number of long coral climate reconstructions using both modern and subfossil corals, taking full advantage of existing resources and by collecting new materials.

(8) *In situ* monitoring of environmental variables (eg, SST, salinity, seawater $\delta^{18}O$ etc.) is required to improve the interpretation of coral $\delta^{18}O$ and Sr/Ca records with respect to regional climate patterns.

(9) Standard calibration and verification procedures (as commonly used in dendroclimatology) should be developed for coral records, with a focus on the interannual-to-decadal timescale, rather than seasonal, calibrations.

(10) Coral Sr/Ca ratios should be routinely measured with $\delta^{18}O$ to improve the climatic interpretation of both geochemical tracers.

(11) All corals used for climate reconstruction should be carefully screened for diagenesis, and this information, along with metadata related to the sampling location, methods and environment should be made available through an established data centre.

(12) Cross-dating of the many Greenland ice cores is reducing dating uncertainties and improving understanding of the factors that cause variations at interannual timescales. Volcanic horizons are particularly important in this regard, especially known events in the historic past such as Icelandic eruptions and Vesuvius in AD 79. This cross-dating of ice cores in Greenland needs to extend to the Antarctic, even though it will be much harder to achieve.

(13) Both analyses of time series of ice-core isotope data and modelling approaches have shown that the traditional spatial calibration of the isotopic thermometer may be unsuitable in many cases. It needs to be supplemented by an improved, quantitative understanding of processes.

(14) Calibration of ice-core isotopic series and of coral records should be undertaken at the appropriate timescale for the problem being studied whenever possible. Attempts to reconstruct the annual cycle may give a false sense of calibration skill.

(15) Further intercomparison and process studies using model and meteorological data are required to improve understanding of the calibration of the ice-core thermometer, its seasonal biases, temporal stability and geographical applicability. The processes controlling isotopic content in non-polar ice cores require particular attention.

(16) Changes in ice sheet elevation and changes in climatic conditions upstream of an ice-core drill site can introduce non-climatic biases in isotopic series. Hence such effects should be considered when interpreting isotopic records from ice cores.

(17) The full range of proxy information available from ice cores has not been exploited. Approaches which use a wider range should be pursued.

(18) Documentary data are limited to regions with long-written histories, but archives from Turkey, Venice, the Vatican and the Middle East have barely been exploited.

(19) In Europe, documentary information decreases significantly once instrumental records commence. This is a severe impediment to their use in CFR and CPS reconstructions, as the use of degraded instrumental data to extend the series to the present may give a false sense of their reliability.

(20) Extending early instrumental data is vital, particularly for the calibration and verification of variability on decadal and longer timescales. Longer instrumental records (than those readily available in climatic data bases) can be found in many regions, with some extending for more than 100 years before the founding of NMSs.

(21) In Europe, there is a potential warm bias in pre-1860 summer temperatures (related to thermometer exposure), particularly in central Europe and Scandinavia.

(22) Wind information from ship logbooks is a reliable source for the reconstruction of past large-scale atmospheric circulation, but has hardly been exploited. There are also numerous (particularly British) logbooks yet to be digitized, which have the potential to improve and extend re-analyses of air or sea temperature as well as sea-level pressure further back in time.

(23) Varved sediments and speleothem records are finding increased palaeoclimatic utility, but a greater focus on quantitative documentation of chronological accuracy and climate sensitivity of both types of records is needed.

(24) Externally forced GCM simulations of the last millennium provide useful test beds for assessing the characteristics of the range of CPS and CFR techniques now available. A range of standard experiments should be developed to test these techniques, perhaps using common sets of pseudo-proxy networks, which could be made widely available for extensive testing of current and new methods.

(25) CPS-based reconstructions of NH temperature averages at the decadal timescale are relatively independent of the reconstruction approach. CFR-based reconstructions of internally consistent climate fields can offer key additional insights into spatial climate processes, but their reliability must be carefully tested (eg, the individual reliability of specific patterns and regions) and linked to the underlying proxy records. Issues regarding the potential underestimation of long-term variability in both CPS and CFR reconstructions have not yet been fully resolved: however, initial work in examining field fidelity for CFRs suggests that pattern reproduction may be robustly resolvable even in the face of significant amplitude loss.

(26) More realistic assessments of reconstruction uncertainty are needed that consider all sources of error and which can powerfully assess field fidelity. Methods need to be developed that incorporate uncertainties in individual proxies, the effect of proxy selection and uncertainties in the regression/scaling models. The ability of simple residuals between reconstructions and observations during calibration or verification to represent the total error needs to be further assessed.

(27) Further simulations with a hierarchy of climate models (GCMs, EMICs and EBMs) are needed to quantify the uncertainties associated with past forcings (eg, using a range of possible past forcings, perhaps with separate simulations for some individual forcings), with internal variability (eg, using ensembles) and varying climate processes (eg, using multiple models and by perturbing physical parameters with a model).

(28) Comparisons between simulations and climate reconstructions must take these forcing/model-related uncertainties into consideration, in addition to errors in the climate reconstructions. Only then can robust conclusions be drawn from such model–data comparisons.

We noted in the Introduction the dramatic improvements in late-Holocene palaeoclimatology made since the early 1990s. Interest in the subject and the large-scale reconstructions has multiplied over the same period. This development has, however, not kept pace with the needs for a more reliable picture for the late-Holocene climates. The questions that are now being asked are also different and can only be adequately addressed with realistic consideration of reconstruction uncertainty ranges. More realistic climate models are providing multiple simulations at higher-spatial resolution for assessing reconstructions and combination (CPS and CFR) approaches, but improvements in reconstructions and reductions in uncertainties in our understanding of late-Holocene climate change will only come with better and more widespread proxy climatic information from more diverse sources.

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Appendix A

Figure 7.1c of IPCC (1990)

In the first report of the Intergovernmental Panel on Climate Change (IPCC, 1990) a ‘schematic’ diagram representing temperature variations over the last millennium was used (Folland *et al*., 1990: figure 7.1c, p. 202). The caption of part (c) of the figure reads: ‘Schematic diagram of global temperature variations for the last thousand years. The dotted line represents conditions near the beginning of the twentieth century’. In the Supplementary IPCC Report in 1992 (Folland *et al*., 1992), the diagram had been dropped and the need for more data that would allow for the spatial aspects of past changes acknowledged. Subsequent IPCC reports included some of the first hemispheric reconstructions based on the burgeoning proxy archives (Bradley and Jones, 1993, in Nicholls *et al*., 1996 (Second IPCC Assessment Report, SAR) and MBH98, 1999; Jones *et al*., 1998; Briffa, 2000 and Crowley and Lowery, 2000 in Folland *et al*., 2001 (Third IPCC Assessment Report, TAR)). Hence the original ‘schematic’ 1990 diagram appeared to have been confined to history by subsequent IPCC reports, although this was never specifically stated. It has continued to reappear in a number of guises – web pages, reports (eg, Wegman *et al*., 2006), school teaching literature, sometimes with phrases evoking reminders of warmer/colder periods in the past (eg, vineyards in southern Britain, Vikings in Greenland in Mediaeval times, Frost Fairs on the Thames and icebergs off Norway in later centuries) – but as far as palaeoclimatologists were concerned the diagram was nothing more than how it was originally described in the caption: a schematic.

So where did the schematic diagram come from and who drew it? It can be traced back to a UK Department of the Environment publication entitled *Global climate change* published in 1989 (UKDoE, 1989), but no source for the record was given. Using various published diagrams from the 1970s and 1980s, the source can be isolated to a series used by H.H. Lamb, representative of central England, last published (as figure 30 on p. 84) by Lamb (1982). Figure 7 shows the IPCC diagram with the Lamb curve superimposed – clearly they are the same curve. The ‘Central England’ curve also appeared in Lamb (1965: figure 3 and 1977: figure 13.4), on both occasions shown as an ‘annual’ curve together with the extreme seasons: winter (December to February) and high summer (July and August). The IPCC diagram comes from the 1982 publication as the vertical resolution of the annual plot is greater. The data behind the 1977 version are given in table app. V.3 in Lamb (1977), but these are essentially the same as previously given in Lamb (1965). All three versions of the plot have error ranges (which are clearest in the 1982 version and indicate the range of apparent uncertainty of derived versions). The 1982 version dispenses with the three possible curves evident in Lamb (1965, 1977) and instead uses a version which accounts for the ‘probable under-reporting of mild winters in Medieval times’ and increased summer temperatures to meet ‘certain botanical considerations’. Lamb (1965) discusses the latter point at length and raised summer temperatures in his Mediaeval reconstructions to take account of the documentary evidence of vineyards in southern and eastern England. The amount of extra warmth added during 1100–1350 was 0.3–0.4°C, or about 30% of the range in the black curve in Figure 7. At no place in any of the Lamb publications is there any discussion of an explicit calibration against instrumental data, just Lamb’s qualitative judgement and interpretation of what he refers to as the ‘evidence’. Variants of the curves also appear in other Lamb publications (see, eg, Lamb, 1969).

Many in the palaeoclimatic community have known that the IPCC (1990) graph was not representative of global conditions (even when it first appeared) and hence the reference to it as a schematic. Lamb’s (1965, 1977, 1982) series has been used as one of the series comprising the NH composite developed by Crowley and Lowery (2000), representative of Central England. Various authors (eg, Farmer and Wigley, 1984; Wigley *et al*., 1986; Ogilvie and Farmer, 1997) have shown that such representativeness is only really the case for the instrumental part of the record from 1659 which is based on the well-known Manley (1974) series. Greater amounts of documentary data (than available to Lamb in the early 1970s) were collected and used in the Climatic Research Unit in the 1980s. These studies suggest that the sources used and the techniques employed by Lamb were not very robust (see, eg, Ogilvie and Farmer, 1997).

In summary, we show that the curve used by IPCC (1990) was locally representative (nominally of Central England) and not global, and was referred to at the time with the word ‘schematic’.

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