The Holocene history of the North American Monsoon: 'known knowns' and 'known unknowns' in understanding its spatial and temporal complexity

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Abstract

Evidence for climatic change across the North American Monsoon (NAM) and adjacent areas is reviewed, drawing on continental and marine records and the application of climate models. Patterns of change at 12,000, 9000, 6000 and 4000 cal yr BP are presented to capture the nature of change from the Younger Dryas (YD) and through the mid-Holocene. At the YD, conditions were cooler overall, wetter in the north and drier in the south, while moving into the Holocene wetter conditions became established in the south and then spread north as the NAM strengthened. Until c. 8,000 cal yr BP, the Laurentide Ice Sheet influenced precipitation in the north by pushing the Bermuda High further south. The peak extent of the NAM seems to have occurred around 6000 cal yr BP. 4000 cal yr BP marks the start of important changes across the NAM region, with drying in the north and the establishment of the clear differences between the summer-rain dominated south and central areas and the north, where winter rain is more important. This differentiation between south and north is crucial to understanding many climate responses across the NAM. This increasing variability is coincident with the declining influence of orbital forcing. 4000 cal yr BP also marks the onset of significant anthropogenic activity in many areas. For the last 2000 years, the focus is on higher temporal resolution change, with strong variations across the region. The Medieval Climate Anomaly (MCA) is characterised by centennial scale 'megadrought' across the southwest USA, associated with cooler tropical Pacific SSTs and persistent La Niña type conditions. Proxy data from southern Mexico, Central America and the Caribbean reveal generally wetter conditions, whereas records from the highlands of central Mexico and much of the Yucatan are typified by long -term drought. The Little Ice Age (LIA), in the north, was characterised by cooler, wetter winter conditions that have been linked with increased frequency of El Niño's. Proxy records in the central and southern regions reveal generally dry LIA conditions, consistent with cooler SSTs in the Caribbean and Gulf of Mexico. This synthesis demonstrates that in some periods, one major forcing can dominate across the whole area (e.g. insolation in the early-mid Holocene), but at other times there is strong variability in patterns of change due to the differential impact of forcings such as the Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO) on precipitation seasonality.

Keywords: North American Monsoon (NAM); Holocene; forcings; climate modelling; seasonality

1. Introduction

In this paper we review evidence for climatic variability across the North American Monsoon (NAM) region over the Holocene and consider the climatic forcing mechanisms that may explain the observed temporal and spatial patterns of change. Given the continuing range of interpretations of what is meant by the NAM region and the complexity of its climatology, however, it is important to start with some definitions. Although 'monsoon' is now largely accepted as an appropriate term to describe the reorganisation of the atmospheric circulation that brings a high proportion of annual precipitation in the Northern Hemisphere (NH) summer (mainly June-September) to large areas of Mexico and the SW USA (Ropelewski et al., 2005), it is a smaller feature than its Asian and African counterparts. It does, however, display the characteristic features of a seasonal reversal of low level temperature gradients associated with continental heating during the summer, the subsequent development of a surface low pressure system and a high level high pressure anticyclone or ridge.

Some papers (e.g. Higgins et al., 1997) have focused attention on the relatively small area of NW Mexico and the SW USA that might be described as the core of the monsoon, where moisture is drawn in at low levels from the eastern tropical Pacific and the Gulf of California. The occurrence of a summer precipitation 'singularity' in Arizona was recognised as early as the mid 1950s (Bryson and Lowry, 1955) and the system was called the Arizona monsoon. Douglas et al. (1993) however pointed out that the source of the monsoon and the maximum summer precipitation actually occurs in Mexico (as identified by Hales, 1974) and proposed the term Mexican Monsoon. The northward migration of the subtropical high-pressure ridge and a displacement of the Bermuda High, however, allows for sustained easterly flow to develop over Mexico and Central America bringing moisture from the Gulf of Mexico and Caribbean (Fig. 1a). There has been a long running debate about the relative importance of the Gulf of Mexico and tropical Pacific/Gulf of California as sources of summer rainfall given the topography of Central America and Mexico. The high mountain ranges of the Sierra Madre Oriental would undoubtedly block any low level flows from the Gulf of Mexico into areas such as Arizona, while the Sierra Madre Occidental effectively channel low level flows on the Pacific side, northwards along the west coast of Mexico and into the SW USA (Fig. 2). It is now generally agreed that the Gulf of Mexico and Caribbean (also known as the Intra-Americas Sea, IAS) is the main source of moisture over much of Central America, eastern and north-central Mexico, probably extending into the SW USA at height (above 850 hPa) (Higgins et al., 1998). There is also a low level flow (the Great Plains Low level jet) that originates over the Gulf of Mexico and brings moisture through eastern Mexico and up into the Great Plains of the USA (Higgins et al., 2003). On the Pacific side, precipitation associated with the Intertropical Convergence Zone (ITCZ) and the warm pool of the eastern tropical Pacific, and the low level flow of the Gulf of California Low level jet (particularly in the form of Gulf surges (Mitchell et al., 2002)) predominate (Figs. 1a and 2a). In this paper we take a wider view of the NAM region to include parts of Central America, the whole of Mexico and the southwestern USA, where a combination of a northward shift in the ITCZ, the onset of deep easterly flows, low level jets originating over the Gulf of Mexico and the Gulf of California and tropical cyclones bring a summer precipitation maximum (Fig. 2).

Across this region there is a broad S – N gradient from the wetter, tropical south, to the dry sub-tropical north (Fig. 3a). In the south, areas such as the Caribbean coast of Central America and the Isthmus of Tehuantepec receive precipitation throughout the year (although with a maximum in the NH summer), while the north is distinguished by the presence of the Chihuahua, Sonora and Mojave deserts. The majority of the NAM region is marked by a strongly unimodal (summer) precipitation regime typical of a monsoon, although in the north the regime is bimodal, with a higher proportion of winter than summer rain in some places (Figs. 1b and 3b). At a finer spatial scale, precipitation distribution is strongly determined by topography, which drives orographic rainfall, helps to determine areas of major convection and constrains flows at low levels (Fig. 2b). The degree of continentality is also important, with the funnel shaped landmass allowing easy exchange of moisture between the Pacific and Gulf of Mexico in the south, but isolating the two moisture sources in the north.

Across the wider NAM region, there is clear variation between the southern portion, where the location of the ITCZ *per se* is probably the dominant feature, and the northern region where the monsoon dominates. The ITCZ occurs primarily over the oceans, where northeast and southeast trade winds converge, giving rise to a belt of clouds and high precipitation. Mechoso et al. (2005) point out that there is a very close relationship between seasonal movements in the ITCZ in the eastern Pacific and the NAM, but Hu and Feng (2002) suggest that the ITCZ has little effect on monsoon rain in NW Mexico/SW USA. This controversy highlights the fact that even within the monsoon dominated region there are important spatial differences that need to be understood in relation to both the present and the past. Broadly speaking, these differences relate to the impact of climatic forcings on precipitation seasonality and the relative importance of winter precipitation. In this regard there are important differences between the northern margins of the NAM and its tropical core. The focus of many studies on the northern margins of the NAM (i.e. the SW USA) has led to an unfortunate neglect of these differential responses when considering long-term variability in the NAM. Detailed analysis of station data has produced a number of suggested regionalisations of the NAM (e.g. Comrie and Glenn, 1998; Englehart and Douglas, 2002; and Gutzler,2004). Castro et al. (2001) have gone so far as to suggest that the NAM in Mexico should be treated separately from the NAM in the SW USA.

In terms of interpreting the Holocene history of the NAM at the broadest scale, it is evident that factors that influence the location of the ITCZ are going to be significant. The link between changes in the global energy balance and the location of the ITCZ (modern average position 6^{o}N) and the Hadley Cells has been discussed by Schneider et al. (2014). Over the long term, this is likely to be determined by insolation, but over shorter time scales, the ITCZ will also respond to SST variability in the Pacific and Atlantic (Magaña et al., 2003; Knight et al., 2006) and changes in solar irradiance (Versteegh, 2005). Over centuries, decades or less, other forcings are likely to dominate, particularly factors such as the Pacific Decadal Oscillation/El Niño Southern Oscillation (PDO/ENSO) and the Atlantic Multidecadal Oscillation/North Atlantic Oscillation (AMO/NAO). As the strength of insolation forcing in the NH declined over the course of the Holocene, it is likely that these other forcings became more important (Faurschou Knudsen et al., 2011). Today, there are significant differences in the impacts of these forcings across the wider NAM region, again separating the northern margin from the tropical core and south. In central and southern Mexico and the Pacific coast of Central America, El Niño events and positive phases of the PDO (Castro et al., 2001; Magaña et al., 2003) reduce NAM summer rainfall when the eastern tropical Pacific warms, the ITCZ stays closer to the equator and the thermal gradient to the continental interior is reduced. In contrast, summer precipitation increases along the Caribbean coast of Central America during El Niño events due to a stronger Caribbean Low Level Jet. During La Niña or negative PDO phases, summer NAM rainfall increases over these southern regions, as an intense ITCZ forms over the eastern Pacific at about 10^{o}N and strong tropical convection and easterly waves develop over the Caribbean and Gulf of Mexico. In contrast in the northern margins of the NAM region, where winter precipitation is more important (Figs. 1b and 3b), El Niño or positive PDO phases give rise to increased winter precipitation (which may lead to wetter conditions overall) while La Niña or negative PDO phases lead to dry winters and drier conditions overall.

It appears that drivers associated with the North Atlantic, specifically the AMO and the NAO (Jury, 2009; Mendez and Magaña, 2010) have their greatest impact on the NAM in the summer season. Positive (warm) phases of the AMO give rise to wetter summers in central and southern Mexico and the wider Caribbean, moving the ITCZ to the north and generating more tropical cyclones (Sutton and Hodson, 2005), while the northern part of the NAM region is drier (Nigam et al., 2011). This antiphase pattern, which today seems to transition around 20^{o}N, warrants more attention in the consideration of palaeoclimatic records. In the context of the Holocene, understanding the mechanisms that might explain severe drought is particularly important, especially given the interest in the possible relationships between climatic and cultural change. Understanding the complexity of response to different forcings across the NAM region is central to this effort (Mendez and Magaña, 2010; Stahle et al., 2012a).

In the rest of this paper, we present and discuss palaeoclimatic data from across the broadly defined NAM region, drawing on both continental and marine records. The emphasis is on terrestrial records that document precipitation change and therefore potentially the strength and geographical extent of NAM. In relation to marine records the emphasis is on those that document Holocene SST changes in monsoon source regions in the Caribbean, Gulf of Mexico, Gulf of California, and eastern tropical Pacific Ocean, as well as those that record the changing influence of ocean-atmosphere teleconnections such as ENSO. We focus mainly on records published since 2000 (see Metcalfe et al., 2000) and draw on more recent reviews which deal with parts of the NAM region (Metcalfe, 2006; Barron et al., 2012) and data model comparisons (e.g. Harrison et al., 2003; Ruter et al., 2004). We have made efforts to highlight recent developments across the region. Our synthesis has been comprehensive rather than selective and draws on the original authors' interpretation of their data in relation to both chronology and the meanings of their climatic proxies. Below, we consider the nature and abundance of the palaeoclimate records. As the nature of the evidence from continental and marine records is rather different, these are first discussed separately. Sites included in this paper (except for tree-ring records) are listed in Tables 1 and 2 and their locations shown on Figure 4; sites with poor dating control across the Holocene have been excluded completely or may not have been considered for specific time periods. We then draw these records together to present maps of conditions across the region for 12,000, 9000, 6000 and 4000 cal yr BP. We review data for the last 2000 years, including the Medieval Climate Anomaly (MCA) and Little Ice Age (LIA) and then reflect on possible forcing mechanisms to explain the observed patterns of change. The paper concludes with some consideration of the use of climate models to explore the NAM region and we reflect on our current levels of understanding and outstanding issues.

2. Palaeoclimatic Evidence

2.1 Continental records

For convenience, the continental records are divided into areas: south, central and north, although we recognise that the boundaries between these are somewhat arbitrary. The southern area includes sites in highland Central America and southern Mexico, the limestone lowlands of the Yucatan peninsula (parts of Guatemala, Mexico and Belize) and the wider Caribbean region. Records from the volcanic highlands of the Trans Mexican Volcanic Belt (TMVB) and the plateau to the north dominate the central area, while the north area includes sites from northern Mexico and from the southwest USA, including the southern Great Basin. We have chosen to include some sites that lie beyond the current northern and western limits of major NAM precipitation. The majority of these continental records are based on lake sediment sequences, but are complemented by speleothem and packrat midden records where these are available. Inevitably, the resolution of these records is highly variable between the different studies and in relation to the different methods of reconstruction applied to the same sequences. Issues involved in the interpretation of some of the proxy records, particularly in relation to precipitation seasonality, are discussed further below. For the latest Holocene, we also make use of tree-ring records. Records from deep-sea cores are also summarised, emphasising their length and temporal resolution.

Lake records from the Yucatan Peninsula are largely confined to the Holocene itself, with many not starting until between 9000 and 8000 cal yr BP when sea level rose sufficiently to support the regional groundwater which today maintains these lake systems (Hodell et al., 1995; Gabriel et al., 2009). Only the deepest lakes in this region (e.g. Peten Itzá, Guatemala) were able to persist through the Last Glacial Maximum and the early Holocene portions of many records reflect sea-level change rather than climate. Early publications on Yucatan lakes were predominantly based on pollen records, most of which show clear evidence of human impact, in some cases being dominated by this (Deevey et al., 1979; Leyden et al., 1998; Wahl et al., 2006). Since the mid 1990s, there has been more focus on stable isotope records ($\delta^{18}O$ and $\delta^{13}C$ in calcite), as these are felt to be less vulnerable to anthropogenic influence (but see Rosenmeier et al., 2002b). Lakes Chichancanab and Punta Laguna (Mexico) have provided benchmark papers for our understanding of both Holocene climatic variability and its possible link to cultural change in relation to the Maya (Hodell et al., 1995, 2001, 2005a; Curtis et al., 1996). Modelling work is beginning to quantify the palaeoclimatic signal in these isotope records (Hodell et al., 2012; Medina Elizalde and Rohling, 2012). Peten Itzá (Guatemala) has yielded the longest continuous record from this southern region (ca. 85,000 years) and has been analysed using a range of methods (Hillesheim et al., 2005; Hodell et al., 2008; Mueller et al., 2009; Bush et al., 2009; Correa-Metrio et al., 2012; Escobar et al., 2012). Following the early pollen work of Barbara Leyden (Leyden, 1984; Leyden et al., 1993, 1994), there have been renewed efforts to develop palynological tools to provide more reliable palaeoclimatic reconstructions in this area, making greater use of modern vegetation data and modelling approaches (Carrillo-Bastos et al., 2012; Correa-Metrio et al., 2013). Dating of Holocene sequences using radiocarbon continues to be problematic where terrestrial macrofossils are not found in core material (e.g. Carrillo-Bastos et al., 2010) and the application of corrections for apparent hardwater error usually still relies on a single value for the whole sequence. The use of other radioisotopes (^{210}Pb, ^{137}Cs, ^{241}Am) offers clear potential for dating the most recent parts of sediment sequences, but to date their application has been limited and has met with variable success (Metcalfe et al., 2009). In this limestone region, speleothems clearly have great potential to yield high-resolution records (using U-series dating), but so far most records cover only the late Holocene and there have been problems in obtaining reliable chronologies for some sites (D. Hodell and M. Brenner, pers. comm.). In some cases, ^{210}Pb dating has been used to ascertain whether the cave has been active in the most recent past (e.g. Webster et al., 2007). Speleothem records of $\delta^{18}O$ in the southern region are dominated by the amount effect, with lower $\delta^{18}O$ values reflecting increased precipitation (Bernal et al., 2011). Quantification of climatic change (precipitation) using the isotope signature in the speleothems from this region is an emerging field (Medina Elizalde et al., 2010).

Purely palaeolimnological records from highland southern Mexico and Central America are rare (but see Habeyran and Horn, 2005; Slate et al., 2013), with much more emphasis being placed on palynology and charcoal analysis. Both pollen and charcoal records from around 4000 cal yr BP often reflect anthropogenic disturbance (Rue, 1987; Clement and Horn, 2001), but may also indicate climatic changes. Slate et al.'s (2013) record from Lake Nicaragua shows similar complexity, with the addition of the effects of tectonic activity. Clearer palaeoclimatic signals have been derived from higher elevation sites (e.g. Islebe et al., 1996). Stable isotope records from lake sediments are also quite rare. $\delta^{13}C_{organic}$ records have been interpreted in both palaeoclimatic (Lane et al., 2011a and b) and anthropogenic (Lane et al., 2004) terms. There are few lacustrine records from across the Caribbean and in limestone areas, radiocarbon dating has been problematic (e.g. Peros et al., 2007), as in the Yucatan. Recently published records from the volcanic islands of the Caribbean (e.g. Fritz et al., 2011), highlight the climatic complexity of this area. As in the Yucatan, speleothems are increasingly providing high-resolution palaeoclimate data from southern Mexico, Central America and the Caribbean (Lachniet et al., 2004a and b; Mangini et al., 2007; Fensterer et al., 2012), although long records are still rare (but see Fensterer et al., 2013; Lachniet et al., 2013).

The volcanic highlands of central Mexico (TMVB), specifically the lake sediments of both present day and former lakes, provided the initial focus for palaeoclimatic work in the tropical core of the NAM region, with the earliest studies dating back to the 1950s (Sears, 1952; Sears and Clisby, 1955; Hutchinson et al., 1956). Pollen and diatoms have been the dominant methods of environmental reconstruction, complemented by sediment chemistry and mineral magnetic measurements (Metcalfe et al., 2000; Caballero and Ortega Guerrero, 2011). The application of stable isotopes has been less widespread than in the Yucatan, as few of the studied lakes in the TMVB precipitate calcite or preserve continuous sequences of ostracods or gastropods that could provide hosts for an isotopic signature. Palaeolimnological records from this area may be affected by volcanic events and sometimes by tectonism (Israde-Alcantara et al., 2005), although tephrochronology has clear potential for inter-site correlation and improving age models (Newton et al., 2005). Almost all Holocene records show evidence of human impact over about the last 4000 years, with only the highest elevation sites apparently providing a uniquely climatic record (e.g. Lozano-Garcia and Vazquez-Selem, 2005); in some cases human activity apparently masks any climatic signal. Previous syntheses of records from this area have highlighted the complexity of the signals from the TMVB (e.g. Caballero et al., 2002). It is notable that there are still rather few continuous, well-dated records, covering the whole of the Holocene, from the TMVB. Published records to date continue to come from sites in the centre and west of the TMVB, although records are emerging from basins to the east, especially in the Oriental with its deep crater lakes, which have the potential to yield valuable high-resolution sequences (e.g. Caballero et al., 2003; Bhattacharya et al., 2015). There has been an increasing focus on high resolution records of the last millennium (e.g. Lozano-Garcia et al., 2007; Kienel et al., 2009; Sosa-Najera et al., 2010; Cuna et al., 2013) providing insights into climatic variability during the Medieval Climate Anomaly (MCA) and the Little Ice Age (LIA) (see below).

Over the last decade, the potential of tree rings to yield high resolution, precisely dated palaeoclimatic records for tropical Mexico has been realised (Villanueva Diaz et al., 2011). In the TMVB (and extending down into western Guatemala) these records come mainly from *Taxodium mucronatum* (Montezuma baldcypress) (Stahle et al., 2012b). Unlike the long tree ring records from the southwest USA, however, few of the Mexican and Guatemalan *T. mucronatum* records extend back beyond AD 1500. Further north, in the highlands of the Sierra Madres Occidental and Oriental and extending into the USA, records from *Pseudotsuga menziesii* (Douglas Fir) are more common and cover longer time periods (e.g. Stahle et al., 2009). Many records can be accessed through the NOAA NCDC http://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/tree-ring.

For the northern part of the NAM region, Barron et al. (2012) have reviewed evidence for precipitation change over the Holocene in relation to sea surface temperature (SST) changes in the Gulf of California and the subtropical Pacific, but here we consider palaeoclimatic records from a wider geographical area. It is notable, that there is a clear contrast between the volume of information available from either side of the Mexico/USA border. Since the synthesis published by Metcalfe (2006), palaeolimnological work on lake sediments in the modern deserts of northern Mexico has continued, although more often that has involved reworking previously published sites in greater detail, or using new methods (e.g. Roy et al., 2013). There has been work on new sites (e.g. Ortega-Rosas et al., 2008a, b), but Holocene chronologies are often rather poor and there is more focus on longer records extending back into the last glacial (Roy et al., 2014). Unfortunately, there have been very few new packrat midden records published from Mexico in recent years (but see Holmgren et al., 2011). Tree ring records from northern Mexico, have burgeoned and are providing detailed, if rather short, records of precipitation variability, although many of these reflect winter precipitation (see Forcings section below, Diaz et al., 2002; Villanueva-Diaz et al., 2007; 2011).

In the areas of the western USA directly affected by the NAM, one of the major developments of the last 10 years has been the increasing number of speleothem records from New Mexico, Arizona and the Great Basin (e.g. Asmerom et al., 2007, 2010, 2013; Lachniet et al., 2014) with high-resolution U-series chronologies. At these more mid latitude locations, temperature affects the isotopic signature as well as precipitation source area and amount, but the focus has been on source area, particularly variations in the amount of winter precipitation originating in the Pacific which has a lower $\delta^{18}O$ than summer rain from the tropics (Asmerom et al., 2010). The interpretation of these speleothem $\delta^{18}O$ records is, however, complicated as today winter rain is more sustained than summer rain and hence more influential in modern cave hydrology (Asmerom et al., 2007; Cole et al., 2007). In contrast to northern Mexico, packrat midden records continue to be developed in the USA (e.g. Holmgren et al., 2006) extending beyond the core NAM region into areas such as the Colorado plateau (Cole et al., 2013). The ability to identify *Pinus* to species level using the needles in the midden records and hence to discriminate between winter and summer precipitation regimes, remains a key strength of packrat middens in the context of understanding the history of the NAM. Pollen data generally do not have the taxonomic resolution (e.g. summer flowering plants) needed to identify the seasonality of precipitation. In relation to lake records, recent work has focused on high-resolution records from California (e.g. Kirby et al., 2010; 2012). Although lying within the modern winter-dominated rainfall area, these records provide insights into NAM dynamics over Holocene and longer timescales and into possible forcing mechanisms. As in our central region and in northern Mexico, tree ring records contribute high-resolution data for the later Holocene again discriminating between winter (early wood) and summer (late wood) rainfall (e.g. Stahle et al., 2009).

2.2 Marine Records

The links between the amount of summer precipitation in the NAM region and sea surface temperatures (SSTs) in both the Atlantic warm pool region of the Caribbean and Gulf of Mexico (GoM) and in the Gulf of California (GoC) have been well established (Higgins and Shi, 2000; Mitchell et al., 2002; Hu and Feng, 2002; Wang et al., 2008; Arias et al., 2012) and as a result, are our focus here. In both regions, increased SSTs and increased surges of monsoon precipitation toward the NAM region are tied to the northward expansion of the ITCZ, although the drivers vary over different timescales. The primary surface-ocean current in the GoM is the Loop Current that brings warm waters from the Caribbean Sea through the Yucatan Strait into the GoM (Fig. 2a). The Loop Current exits the GoM into the North Atlantic Ocean through the Florida Straits. Surface-water and wind circulation in the Caribbean GoM region show large annual changes linked to seasonal migration of the ITCZ. Poore et al. (2005; 2011) argued that century- to decadal-scale variability in *Globorotalia sacculifer* (planktonic foraminifer) abundance was also related to changes in the average position of the ITCZ. Warmer intervals would result in a more northerly average position of the ITCZ and higher *G. sacculifer* abundances in gulf sediments, whereas cooler intervals would result in a more southerly average position of the ITCZ and lower *G. sacculifer* abundances in GoM sediments.

The Gulf of California (GoC) provides a critical link between the tropical eastern Pacific and the southwest US, as it extends for over 1000 km from ~23°N to ~30°N and is separated from the cool waters of the mid latitude North Pacific by the Baja California Peninsula. During the late autumn to early spring, northwest winds blow down the axis of the GoC causing upwelling of nutrients, which leads to extensive blooms of diatoms and other phytoplankton. As these winds wane or reverse in the late spring, there is a steady northward advance of tropical waters (>26°C) up the axis of the GoC that is associated with the northward advance of the ITCZ. Mitchell et al. (2002) argue that the timing of monsoonal rainfall in Arizona is closely linked with the arrival of SSTs >26°C in the northern GoC. The southern half of the GoC contains 4 major deep (>1000 m) basins (the Guaymas, Carmen, Farallon, and Pescadero Basins) that are the locus of biogenically rich (calcium carbonate and biosilica) sediments characterized by high (>50 cm/ka) depositional rates and by the widespread occurrence of laminated sediments due to an expanded oxygen minimum zone. Consequently, sediments collected from the GoC have been the subject of numerous paleoceanographic studies, especially during the past 10 years (Barron et al., 2005, 2012; Pichevin et al., 2012; McClymont et al., 2012).

The two main proxies used for Holocene SSTs in the monsoon source areas within Gulf of Mexico and the Caribbean in the east and the tropical Pacific in the west are alkenones, highly resistant organic compounds (ketones) produced by Coccolithophorids, and Mg/Ca measurements made on the calcium carbonate shells of surface dwelling planktonic foraminifers. As reviewed by Leduc et al. (2010), alkenone production occurs in the thermocline and has a strong seasonal bias, whereas differing hydrological conditions and ecological behaviour of planktonic foraminifers can affect Mg/Ca SST estimates. The only detailed Holocene alkenone SST record from the eastern tropical Pacific source area for NAM is that of McClymont et al. (2012) from the Gulf of California; however, it does not extend younger than ~6800 cal yr BP. Microfossil assemblage changes (e.g., planktonic foraminifers, calcareous nannofossil, diatoms, and silicoflagellates) have also been used to infer Holocene SST changes within Atlantic and Pacific source areas for NAM precipitation (Barron et al., 2005; Poore et al., 2005). While transfer functions relating microfossil assemblage changes to SSTs have been established in other regions of the world's oceans, quantitative relationships between microfossil assemblage changes and SST within the tropical water source areas of NAM precipitation are not well developed. Key marine records and their SST reconstructions are shown in Table 2.

These marine records are divided into those from the GoM and Caribbean (the IAS), the eastern tropical Pacific, the GoC, and the eastern Pacific off Baja California, which is north of monsoon source area. Records of El Niño frequency from Ecuador (Moy et al., 2002) and the Galapagos Islands in the eastern equatorial Pacific (Conroy et al., 2008) have been taken as important points of reference, as they provide indirect evidence about the summer position of the ITCZ in the eastern tropical Pacific (a more northerly ITCZ is associated with La Niña conditions – a more southerly ITCZ occurs during El Niño conditions). Similarly, the east-west SST gradient in the equatorial Pacific is a measure of the prevalence of El Niño *vs.* La Niña conditions (Koutavas and Joanides, 2012), with reduced east-west gradients during El Niños and increased gradients during La Niñas.

Considered as a whole, the IAS sites record their warmest SSTs between ~ 9000 and 4000 cal yr BP and during the Medieval Climate Anomaly (MCA), intervals that have been linked with strong NAM (Harrison et al., 2003; Poore et al., 2005; Richey et al., 2007; Feng et al., 2008; Zhao and Harrison, 2012; Barron et al., 2012). In this paper we shall also make reference to the record from the Cariaco Basin (off northern Venezuela), which has been widely interpreted as a record of change in the location of the ITCZ (e.g. Haug et al., 2001). Although it is arguable how much the record from this low latitude site (at 10°N) is directly related to the strength of the NAM in regions north of ~20°N, Cariaco is a global benchmark site for the tropics and ITCZ variability (Schneider et al., 2014).

Tropical Pacific SST records are variable during the Holocene, but most show increasing SSTs during the past 5-6000 cal yr BP (Leduc et al., 2010), presumably due to a gradual increasing expression of El Niño events. According to Koutavas and Joanides (2012) and numerous studies in the mid to high latitude North Pacific (Barron and Anderson, 2011), the middle part of the Holocene between ~8000 and 4000 cal yr BP was characterised by increased La Niña-like conditions which would favour a more northerly ITCZ and increased monsoonal rainfall in northwest Mexico and Arizona (Barron et al., 2012). The detailed record of McClymont et al., (2012) from the Gulf of California (see above) reveals a ~4°C warming of GoC SSTs at ~8200 cal yr BP that was immediately preceded by relatively cool SSTs between ~10,600 and 8200 cal yr BP. Barron et al. (2012) cited McClymont et al.'s (2012) SST record in arguing that modern NAM circulation in the GoC did not exist prior to the middle Holocene. Barron et al. (2005) use calcium carbonate and tropical diatom proxies of SST in suggesting that GoC SSTs were warmest between ~6200 and ~3500 cal yr BP. We suggest that monsoon moisture surges up the GoC probably diminished after ~3000 cal yr BP due to the increased expression of El Niño events which would have caused the Pacific ITCZ to have positioned further to the south (Perez-Cruz, 2013).

3. Patterns of change in the NAM

Figures 5 – 8 present a series of snapshots of conditions across the NAM region for 12,000, 9000, 6000, and 4000 cal yr BP based on the records listed in Tables 1 and 2. Details of the coding assigned to the continental sites are given in the expanded version of Table 1 in the Supplementary Material. Similar maps are presented for the MCA and LIA (Figs. 10 and 11) to give a broad indication of patterns for these time periods, even though there may be considerable variation within them (see below). To give some indication of the robustness of a signal for any of the chosen time periods, entries in Table 1 in the Supplementary Material are in bold where there are at least three data points within ± 200 years of the chosen date and the symbols on the maps have solid fill. Otherwise, entries in the detailed Table 1 are in italics text and the map symbols are open. Continental sites are coded based on effective moisture (water balance) relative to the full period of record. It should be noted that the effect of this on sites in the north (dry today) is that they may be coded low/dry when conditions were actually still significantly wetter than today. Marine records have been used to reconstruct sea surface temperature and ENSO state. Where possible, for both continental and marine records, we have used the authors' own interpretations, but in some cases have applied our own to their published data.

3.1 12,000 cal yr BP

12,000 cal yr BP (Fig. 5) represents conditions during the Younger Dryas chronozone (12,900-11,600 cal yr BP). Pollen-based temperature estimates from continental sites (e.g. Lozano-García and Vázquez-Selem, 2005; Ortega-Rosas et al., 2008a) and most sea surface temperature reconstructions (Table 2) show widespread cooling, although relatively warm SSTs are reported by McClymont et al. (2012) from the GoC. Terrestrial records (Table 1 Supplementary Material) indicate generally wet conditions in the inland parts of the north (see also Antinao et al., 2013), while records from the central and southern NAM regions are limited. In the Yucatan peninsula the lack of information reflects the impact of lower sea level on the regional water table (see above). The limited data available point to dry conditions in the south and central region, only the record from the Pátzcuaro Basin in the TMVB is an exception to this (Caballero et al., 2010). In contrast, records from the northern area of NAM show generally wet conditions. It is worth noting that Babicora in the interior of northern Mexico still had more effective moisture than present, although drier in relation to previous conditions, while in the highlands of the Sierra Madre Occidental (Ortega-Rosas et al., 2008a) conditions both drier and colder than present at the YD (ie a south/central pattern) have been recorded. Tufas from the Salton Basin show a complex pattern with generally dry conditions interrupted by an extremely wet episode around 12,300 cal yr BP. The south/central pattern observed is consistent with the southward displacement of the ITCZ and weakened monsoon, widely reported from other tropical locations at this time (e.g. Wang et al., 2001; Tierney et al., 2011). Wet conditions in parts of the north reflect the variation in seasonality of precipitation across the NAM region. Southward migration of the ITCZ and the continuing presence of the Laurentide Ice Sheet allowed the mid-latitude westerlies originating from the Pacific, to penetrate further south as the jet stream was displaced equatorward (e.g. Asmerom et al., 2010). Similar conditions apparently occurred during the full glacial (Benson et al., 1990; Bradbury et al., 2001; Polyak et al., 2004;

Lachniet et al., 2014). This southward displacement of the jet stream has been supported by some climate modelling (see below). The increase of winter precipitation in the northern part of the NAM under glacial-type background conditions is a well established idea, but how far south this influence reached and the actual balance between winter and summer precipitation is difficult to establish using the proxies available, most of which have limited seasonal discrimination. There have been suggestions that wetter conditions at this time in the northern NAM area were associated with an early re-establishment of the NAM (Roy et al., 2014 and references therein). Midden records from San Pedro Martir in northern Baja California (Holmgren et al., 2011) show increasing proportions of C4 grasses and summer flowering plants from 13,600 cal yr BP, which are interpreted as reflecting the onset of summer, monsoonal rainfall. SST reconstructions do show warm water in the GoC (Table 2). Summer rainfall is also invoked to explain wetter conditions at Lake Tulane in Florida. The authors suggested that cooling in the North Atlantic effectively trapped warm water in the GoM leading to the persistence of summer rain (Grimm et al., 2006).

3.2 9000 cal yr BP

By 9000 cal yr BP (Fig. 6), wetter conditions are well established in the southern part of the NAM region, as well as in areas more clearly under the direct influence of the ITCZ. Most sites in the Yucatan are still dry at this time, but with rising sea levels, many lakes are filling and rising rapidly. Wet conditions also persist in the western part of the northern region, most clearly in California and Nevada. Sea surface temperatures were generally warm (Table 2), with the GoC being an exception. It seems that the NAM was becoming established by this time, although it was still not fully developed. A number of sites in the TMVB display a short period of drier conditions around 9000 cal BP, followed by lake level rise at most sites. In the deserts of northern Mexico there is a return to wetter conditions after a relatively dry YD that has been interpreted as reflecting increased summer precipitation (Metcalfe et al., 2002). Midden data from Chihuahua (Van Devender, 1990) and pollen data from the Sierra Madre Occidental (Ortega-Rosas et al., 2008a, b), however, suggest that increased winter precipitation probably persisted until this time. Wetter conditions in California and northern Baja California are generally interpreted as reflecting the persistence of increased winter precipitation, but with signs that the belt of increased winter rain was moving northwards as the LIS retreated (Barron et al., 2012; Kirby et al., 2012). Antinao and MacDonald (2013) suggest that enhanced winter frontal storm activity originating in the tropical Pacific was responsible for alluvial fan aggradation between 11,500 and 9,000 cal yr BP in the northern NAM region. This is consistent with Kirby et al's (2012) suggestion that atmospheric river storms have played an important role in delivering substantial amounts of moisture to southern California through the Holocene.

3.3 6000 cal yr BP

At 6000 cal yr BP there are many more records for the NAM region (Fig. 7). Although individual sites vary, there is a broad pattern of wetter conditions across the NAM area, extending from Central America up into Arizona and New Mexico. This is a time of warm SSTs across the Intra Americas Sea and the GoC (Table 2), although there is some evidence of cooling in the eastern equatorial Pacific (Koutavas and Joanides, 2012). This period seems to mark the peak of the NAM, with its influence extending north of its current area (Harrison et al., 2003; Barron et al., 2012). Barron et al. (2012) argue that prior to ~8000 cal yr BP, lower GoC SSTs would not have fuelled northward surges of tropical moisture, arguing that modern NAM climatology began at ~7500 cal yr BP.

At the northern margin of the NAM there are indications, particularly from the lake records, that peak available moisture occurred prior to 6000 cal yr BP, as increasing temperatures outweighed any increase in summer rain (E > P). As the monsoon extended, some areas dominated by winter precipitation dried out as the westerlies tracked still further north. There are a few sites in the central and southern regions of the NAM region that show drying around 6000 cal yr BP, although in most cases, this is just a brief interruption of increasing moisture availability. There may be more variability from the mid-Holocene on at sites more directly affected by the ITCZ (see Fritz et al., 2011).

3.4 4000 cal yr BP

The map for 4000 cal yr BP shows a complex pattern (Fig. 8). Wet conditions persist across much of the southern part of the NAM region, although at many sites this marks the final stage of the mid Holocene wetter period. In contrast, most sites in the TMVB are already relatively dry by this date. Exceptions include the high elevation site of Agua El Marrano (3860 m) where an increase in moisture after a dry mid Holocene (around 5000 cal yr BP), is associated with cooling and the spread of pine forest (Lozano-Garcia and Vazquez-Selem, 2005). This pattern would be consistent with a reduction in monsoon strength associated with the gradual southward migration of the ITCZ as NH insolation declined, although SSTs remain generally warm (Table 2). At Zirahuen in the TMVB a dry event around 4000 cal yr BP has been linked to increased ENSO forcing (Lozano-García et al., 2013). In the northern part of the NAM region, there is an interesting pattern of clearly drier conditions along the eastern and western margins, with a belt of wetter conditions between. It has been suggested that wetter conditions in some of the lake basins (e.g. Palomas) were being driven by increased winter precipitation associated with more El Niño type conditions as a result of the onset of neoglacial conditions (Castiglia and Fawcett, 2006). An increase in winter precipitation is also reported from sites in the NW Sierra Madre Occidental by Ortega Rosas et al., (2008b) based on pollen biomisation. It is however hard to reconcile this interpretation with drier conditions in southwest California (where winter precipitation dominates) (Kirby et al., 2012) and eastern Arizona, but dating control may play a part.

An interesting feature of this 4000 cal yr BP map is the emergence of antiphase conditions between the northern Yucatan Peninsula and the TMVB even though they both lie within the core of the summer rainfall regime. Together with the varied responses in the north of the NAM region, this tends to suggest that by 4000 cal yr BP more complex forcings were coming in to play as the direct role of insolation declined. This issue is discussed further below.

3.5 Last 2000 years

By ca. 2000 cal yr BP, as the influence of orbital forcing continues to decline through the Late Holocene, there appears to be an increasingly complex pattern of moisture availability across the NAM region, although this may just reflect higher sampling resolution. Consequently, we have not attempted to construct a map for 2000 cal yr BP. Many lacustrine records have been complicated by the increasing intensity of human impact across the region and teasing apart climatic and anthropogenic signals can be problematic (see section 2.1). Despite these challenges, investigating the links between cultural change and climatic variability over the last 2000 years has been a key focus of research, notably in relation to the Late-Classic Maya civilisation of the Yucatan Peninsula and Guatemala (e.g. Hodell et al., 1995; Curtis et al., 1996; Wahl et al., 2013) and the pre-Columbian Native American populations in SW USA, such as the Anasazi (Benson et al., 2003; 2007). The examination of spatial patterns and mechanisms of change through the last 2000 years has focused on the nature and timing of hydrological extremes, with a particular emphasis on the Medieval Climate Anomaly (c. AD 900 – 1300), the Little Ice Age (c. AD 1400 - 1850), and the transition between the two (e.g. Graham et al., 2007; 2011; Metcalfe and Davies, 2007; Feng et al., 2008). The dates assigned to both the MCA and the LIA are quite variable, especially in relation to the former, with published dates ranging between AD 800 and 1400. Here we look generally at these periods as well as focussing in on particular intervals. A dense network of tree-ring records in the SW USA, which now extends into northern Mexico with some sites in the TMVB, coupled with a growing number of speleothem records (e.g. Bernal et al., 2011, Lachniet et al., 2012; 2013), provide insights into interannual and decadal scale change which is superimposed on the broader trends. Tree ring records have been crucial in identifying the timing and duration of 'megadroughts', which are implicated in societal upheaval across Mesoamerica and the SW USA (e.g. Therrell et al., 2004; Stahle et al., 2007).

In general terms, the last 2000 years is marked by highly variable conditions, associated with the increased influence of ENSO/PDO and NAO/AMO. This occurs against a backdrop of a drying trend in the central and much of the southern region as the ITCZ continues to move southwards (although some Caribbean sites stay wet). Figure 9 compares selected high-resolution records with robust chronological control over the period along a N-S transect through the NAM region. Long-term changes in the July precipitation reconstruction from tree rings at El Malpais are more difficult to identify as the record is detrended. However, as a record of July rainfall, it is more directly comparable with other proxy records. Warmer SSTs in both the GoM and the GoC are observed during the early part of the MCA. Southern records indicate wetter conditions (e.g. Yok Balum, Lago el Gancho) associated with a more northern position of the ITCZ reflected in the Cariaco Basin titanium record (Haug et al., 2001). The Terminal Classic Drought is evident at Yok Balum. The TMVB is clearly a transition between the northern and southern monsoon regions. The Ti record from Juanacatlan closely corresponds to the Yok Balum record from Belize after about AD 1100 when wetter conditions are evident at both sites. However, the records diverge during the LIA, with Juanacatlan reflecting a clear two-part signal of drought between c. AD 1400-1600 and generally wetter conditions during the 17th and 18th centuries, which continues through the 19th century (Metcalfe et al., 2010). Asmerom et al., (2013) report a similar two-part signal. Historical climatology is providing greater insights into the degree of climatic variability within the LIA (see below) and the potential for integrating conventional palaeoclimatic approaches with the use of historical records should be exploited more often (Endfield and Marks, 2012). In addition to their value to palaeoclimatology, such historical records are invaluable for assessing the impacts of climate change and the resilience of different societies, which is of such interest today.

During the MCA a marked difference between the northern and southern NAM regions is observed (Fig. 10). Low lake levels are recorded in the Great Basin and California (Benson et al., 2002; Kirby et al., 2010), extending down as far as the TMVB. Tree ring records provide evidence of severe, centennial scale ‘megadrought’ across SW USA (Cook et al., 2004; Herweijer et al., 2007; Stahle et al., 2009) during the MCA, whilst drought-resistant taxa dominate pollen from laminated sediments in the Santa Barbara Basin between c. AD 800 and 1090 and again between c. AD 1200 and 1270 (Heusser et al., 2015). Megadroughts in the northern region have been associated with cooler SSTs in the tropical Pacific typical of La Niña type conditions (Kennett and Kennett, 2000; Cobb et al., 2003; Graham et al., 2007). This hypothesis is supported by recent high-resolution data south of the TMVB, where generally wetter conditions during the MCA are observed (e.g. at Juxtlahauca Cave in the Sierra Madre del Sur: Lachniet et al., 2012), throughout the circum-Caribbean region (e.g. Fritz et al., 2011; Fensterer et al., 2012) and Central America (e.g. Lane et al., 2009; Stansell et al., 2011). It appears that this dipole was clearest between about AD 950 and 1050 with La Niña like conditions in the Pacific and warm SSTs in the IAS (as in a positive AMO), although the picture in the central region is rather more complicated, where drier conditions prevail (Fig. 10). As mentioned above, there is considerable variability during the MCA, some of which is captured in the modelling study of Feng et al. (2008).

The well-documented Terminal Classic Drought (TCD) linked to the collapse of the Late-Classic Maya civilisation (Hodell et al., 1995; 2005a; Curtis et al., 1996) between AD 850 and AD 910 (ie within the timeframe for the early MCA) was originally suggested to be a drought spanning c. 150 years. High-resolution records now indicate a series of relatively short, multi-year droughts was responsible (Haug et al., 2003; Medina-Elizalde et al., 2010), offering an explanation for the gradual and spatially uneven societal response. Recent evidence from Frappier et al. (2014), however, suggests that more sustained drought may have been masked by more frequent and intense hurricane seasons in the tropical Atlantic, consistent with warmer SSTs recorded in the GoM and Atlantic during this time. Evidence for the TCD is found across the circum-Caribbean (Lane et al., 2014) and north into the TMVB (Metcalfe et al., 2007; Vazquez et al., 2010; Stahle et al., 2011) as far west as Jalisco (Metcalfe et al., 2010), although the lower resolution of some lake records makes correlation problematic (Metcalfe and Davies, 2007). Wahl et al.'s (2013) record from Puerto Arturo suggests that the TCD represents the earlier phase of a two-part MCA, with a transition to wetter conditions occurring c. AD 1100 (Wahl, written communication, 2011). This pattern is also evident in the oxygen isotope record from Laguna de Felipe in the Dominican Republic (Lane et al., 2011b).

The LIA (c. AD 1400-1850) was characterised by drier conditions across much of the tropical NAM region (Fig. 11). Many records lack sufficient chronological control or resolution to examine this period in detail, but there are an increasing number of high-resolution lacustrine (Metcalfe et al., 2010; Stansell et al., 2012) and speleothem (Medina-Elizalde et al., 2010; Fensterer et al., 2012; Kennett et al., 2012; Lachniet et al., 2012) records. Cooler SSTs in the GoM (Table 2) were associated with a southward shift in the ITCZ, weakening the summer monsoon (Hodell et al., 2005b). From AD 1521, Colonial documentary sources from Mexico provide detailed records of drought, with particularly severe episodes reported during the late 1500s and late 1700s (1785-86 being the Año del Hambre – Year of Hunger). In the most northern NAM region, generally wetter conditions prevailed (Benson et al., 2003; Cook et al., 2004), although Stahle et al. (2007) identify three megadroughts in the 14th, 15th and 16th centuries from tree ring data. In comparison with the MCA map, it appears that during the LIA, the transition between the northern and southern regions of the NAM has migrated northwards. It has been suggested that variations in sunspot activity may have played a part in driving the changes in climate observed during the LIA (Metcalfe et al., 2010; Asmerom et al., 2013) with dry conditions over much of the NAM region except the extreme north and west. Such coherency of behaviour probably reflects the combined effects of different forcings all having similar overall impacts on precipitation (Stahle et al., 2012a) and a return to a climatology more driven by the location of the ITCZ (in response to NH cooling) rather than changes in the tropical Pacific and Atlantic Oceans.

4. Forcings

The broad trends across the NAM region over the Holocene show clear differences between the southern and northern regions. Overall, the southern region starts dry and gets wetter (peaking in the mid Holocene) (Figs. 5 – 7) and is still wet today, dominated by summer rainfall. The northern part of the NAM, however, starts wetter than today (maintaining a pattern generally thought to have been established around the LGM) and then gets drier, evolving towards the modern conditions of the Chihuahuan, Sonoran and Mojave Deserts. In this northern region the earliest Holocene (Fig. 6) apparently sees the persistence of increased winter precipitation, a period in the mid Holocene when summer rain increased, probably in both amount and regional extent, and a late Holocene interval when winter rain became more important and the modern bimodal precipitation regime of the extreme NW of the NAM region became established. The TMVB (central region) is intermediate between these, but generally following the more southerly pattern. Clearly, patterns of available moisture (water balance) across the land areas, reflect a balance of precipitation and evaporation (temperature) and there is clear evidence for the cooler conditions of the last glacial persisting into the earliest Holocene (Lozano-Garcia and Vazquez-Selem, 2005; Ortega-Rosas et al., 2008a, b). Over most of the Holocene, however, it is changes in precipitation amount and seasonality that predominate. In monsoon areas, the dominant forcing of longer-term climatic changes is insolation, predominantly driven by precession (Kutzbach, 1981). Overall NH summer (June) insolation peaked around 12,000 cal yr BP and then declined. The peak increase in autumn (September) insolation occurred around 6000 cal yr BP (Fig. 12) and from around 5000 cal yr BP winter (December) insolation showed its largest relative increase. Records from the NAM region can be seen to follow this classical monsoonal forcing in general terms (Fig. 7), but there are clear differences from other NH monsoonal areas (e.g. African/Asian monsoon) in relation to the timing of the early Holocene monsoon onset, the magnitude of the monsoon peak and the greater influence of winter precipitation in both the early and late Holocene.

In the late glacial and early Holocene, the presence of the Laurentide Ice sheet (LIS) and the impact of pulses of glacial meltwater entering the Gulf of Mexico (and its wider effect on the Atlantic Meridional Overturning Circulation) have to be taken into account. The impact of the LIS on the position of the jet stream and the mid-latitude Westerlies has been discussed above and its continued, if diminishing presence through to around 7000 - 6000 cal yr BP would have affected the rate and magnitude of warming over the North American land mass (see Modelling section), the warming of SSTs in the GoM and the northward progression of the ITCZ (Poore et al., 2005;

Montero-Serrano et al., 2011). Feng et al. (2011) suggest that the presence of the LIS through the early Holocene had a marked impact on the sensitivity of North American climate to a range of forcings. Although Table 2 and Figures 5 and 6 reflect the generally cooler SSTs in the GoM in the latest glacial and in some cases into the early Holocene, Aharon (2003) reports a series of meltwater pulses (termed MWF's) entering the GoM via the Mississippi between 16,000 and 9000 cal yr BP. As the GoM is one of the major source regions of monsoonal moisture, the impact of this must be considered. His MWF-4 (11,900 cal yr BP) was apparently the largest of these and resulted in a large isotopic excursion in the record. Between 9900 and 8900 cal yr BP they record four separate pulses, although not as large as the earlier events. Rasmussen and Thomsen (2012) also report meltwater events, including one in the early Holocene (11,700 – 10,500 cal yr BP) which they suggest reduced SSTs and salinity in the Gulf Stream west of Florida. Although short lived, it seems highly probable that these events would have had some impact on monsoon onset over the eastern part of the NAM.

In spite of the limiting effects of both the LIS itself and the resulting meltwater, the evidence suggests that insolation driven change in the ITCZ and the strength of the monsoon dominated the climatic signal across the NAM region through the early Holocene, with the NAM region reaching its greatest spatial extent around 6000 cal yr BP (even if the peak of effective moisture occurred a little earlier than this when P-E was greater with slightly lower temperatures). This pattern is summarised in Figure 13 which compares the Cariaco Basin %Ti record of Haug et al. (2001), with the GoM proxy SST records of LoDico et al. (2006) and Poore et al. (2001), and the cave oxygen isotope records from Pink Panther Cave in New Mexico (Asmerom et al., 2007) and Buckeye Creek Cave in West Virginia (see Fig. 4) (Hardt et al., 2010). These are referenced against June insolation at 30°N and Shuman et al.'s (2005) estimate of the size of the Laurentide Ice Sheet. These records support Hardt et al.'s (2010) and Montero-Serrano et al.'s (2011) hypothesis that the Bermuda High was shifted to the south during the early part of the Holocene, diminishing the supply of monsoonal rainfall to New Mexico and the southwest USA. Liu et al. (2004) suggest that northward displacement of the ITCZ over the tropical eastern Pacific at this time was also driven by a strong temperature gradient in the central eastern Pacific (cold equator, warm midlatitudes) consistent with observations of modern NAM climatology. Over this period, the climate of the NAM region and the Caribbean was effectively dominated by the location of the ITCZ and conditions over the IAS and the wider Atlantic. The influence of the Pacific on the NAM region was relatively weak.

As the strength of summer and autumn insolation forcing in the NH declined and winter insolation became increasingly important, the ITCZ began to track back south and the NAM weakened (Hodell et al., 1991; Haug et al., 2001; Poore et al., 2005; Peterson and Haug, 2006; Mueller et al., 2009). This weakening of the monsoon was matched by increasingly complex patterns of climatic change across the NAM region (see the 4000 cal yr BP reconstruction, Fig. 8) that included evidence for an increase in winter precipitation (originating in the Pacific) in the north. It has been widely proposed that this change reflected the increase in frequency and strength in ENSO beginning around 5000 years ago (Conroy et al., 2008; Donders et al., 2008) and with it more ENSO type variability (including the PDO) operating over longer timescales. Today, during El Niño events, the ITCZ is displaced south in the eastern tropical Pacific resulting in reduced (summer) precipitation across most of Mexico (Magaña et al., 2003) hence magnifying the broader scale impact the southward migration of the ITCZ. The increasing importance of ENSO (sensu lato) is clearly shown in marine records from the eastern Pacific and GoC in Figure 14, which compares Holocene El Niño frequency and ENSO conditions in the equatorial Pacific with SST reconstructions in the eastern tropical Pacific, the GoC, and Pacific waters off Baja California (Mexico). Between ~8000 and 4000 cal yr BP, cooler equatorial Pacific SSTs during extended La Niña-like conditions occur at the same time as warmer GoC SSTs, an inferred increased NAM strength in the southwest US (Barron et al., 2012) and increased coastal upwelling (cooler SSTs) off the Pacific coast of Baja California. Enhanced ENSO cycles after 4000-3000 cal yr BP coincide with a decrease in GoC oceanographic/atmospheric conditions conducive to strong NAM circulation.

Although less widely reported, there is also some evidence for increased definition in and influence of the AMO, as the ITCZ moved south (Faurschou Knudsen et al., 2011). The period between 4000 and 3000 cal yr BP appears to be one of significant reorganisation in the climatology of the NAM region (and elsewhere), as the impact of ENSO (and the PDO) increased (Donders et al., 2008; Barron and Anderson, 2011). In the GoC we see the onset of modern conditions (Barron et al., 2005; Douglas et al., 2007) at about this time. Perez-Cruz (2013) suggests that after ~2400 cal yr BP an increase in ENSO dominance and frequency diminished monsoon moisture surges up the GoC, causing a further southward displacement of the ITCZ in the eastern tropical Pacific. As drivers of change originating in the Pacific became more important, so did the present day pattern of a unimodal rainfall regime in the south and bimodal regime in the north, and differential patterns of response to ENSO/PDO and NAO/AMO became established, with their clear differences in seasonality. Bernal et al. (2011) point out that after 4300 cal yr BP the correlation in behaviour noted in the early and mid-Holocene between their Cueva de Diablo speleothem record (in our southern region) and the record from Pink Panther cave (in our northern region) breaks down. In the north, an increase in winter precipitation associated with increasing ENSO dominance seems to have brought a return to slightly wetter conditions in some locations in the driest parts of the NAM (e.g. Menking and Anderson, 2003 (Estancia); Castiglia and Fawcett, 2008 (Palomas) and a return to periods of sustained wet climate in the winter rain dominated areas (e.g. Kirby et al., 2010 (Elsinore)). The possible impact of neoglaciation is also referred to by some authors (e.g. Enzel et al., 1992; Lozano Garcia and Vazquez Selem, 2005 and Castiglia and Fawcett, 2008). In the southern NAM region, the pattern is generally one of gradual drying (Hodell et al., 2001; Mueller et al., 2009; Bernal et al., 2011). In addition to the increasing variability within the NAM region, there is evidence for fluctuating conditions across the Caribbean after the humid, ITCZ/Atlantic dominated mid-Holocene (Mangini et al., 2007; Fritz et al., 2011).

Records, which can reliably capture short-term forcings such as ENSO/PDO and NAO/AMO, are necessarily of very high resolution and typically relatively short in duration. Tree ring records that often provide the clearest evidence of this scale of forcing are increasing in number across the NAM region. This has been stimulated by the more frequent separation of early wood and late wood records, allowing the season of precipitation to be isolated. Papers reporting ENSO include Stahle and Cleaveland (1993, northern, winter-spring precipitation), Diaz et al., (2001, Baja California Sur, northern; 2002 Chihuahua, winter-spring), Cleaveland et al., (2003, Durango, northern, winter) and Stahle et al. (2011, Barranca de Amealco TMVB, June PDSI). Speleothem records can also have the resolution to capture ENSO signals, although the seasonal interpretation may be regarded as more subjective. Lachniet et al. (2004a) in a short record from Chilibrillo Cave in Panama report a strong ENSO response (El Niño being drier). ENSO forcing is also suggested by Lachniet et al. (2012) in their record from Juxtlahuaca cave in south-central Mexico. A number of the ENSO records also refer to the PDO, as the expressions of these forcings are similar (see the Introduction). Asmerom et al. (2007) suggest a combination of both solar forcing and PDO/ENSO to explain their record from New Mexico. More marine records have the resolution needed to capture this decadal scale variability, especially those from the GoC (e.g. Staines Urias et al., 2008). Barron and Anderson (2011) report changes in ENSO/PDO frequency in a time transgressive manner along the eastern North Pacific margin over the late Holocene.

As mentioned above, AMO records are scarcer and more temporally limited than ENSO/PDO records. In part, this reflects the focus of activity and high-resolution records in the northern part of the NAM region where the AMO is less important. AMO forcing has been reported from the Cariaco Basin (Black et al., 2007), the Gulf of Mexico (Poore et al., 2009) and central Mexico (Stahle et al., 2012a). Winter et al. (2011) also record AMO forcing in their 800-year speleothem record from Puerto Rico. In these cases, a positive AMO (warmer SSTs) corresponds with wetter conditions as occurs today (Mendez and Magaña, 2010). The dominance of the tropical Pacific in driving conditions during the MCA in North America has also been challenged by modelling (Oglesby et al., 2012) which places more emphasis on the AMO.

The significance of these higher resolution records, which can capture at least decadal scale variability, is that they cast more light on the issue of antiphase behaviour between the truly tropical south and the north of the NAM region which is so evident in the modern climatology (see Introduction). Given the seasonal resolution of the tree ring records, this opposite behaviour is perhaps most convincingly captured by this type of record. Therrell et al. (2002) summarised 18 chronologies from different species and pointed out that northern and southern Mexico showed different patterns. Stahle et al. (2012a) emphasise the important differences in response to forcings between the SW USA (our northern region) and Mesoamerica (effectively our southern and central regions), with warm Atlantic SSTs (positive AMO), La Niña or cold (negative) PDO in the Pacific giving rise to a wet Mesoamerica due to increased summer precipitation and dry SW (but see below regarding the MCA), while cold Atlantic SSTs (negative AMO), El Niño or warm (positive PDO) resulting in dry conditions in Mesoamerica and wet conditions in the SW US (due to increased winter precipitation). Understanding the patterns of behaviour during periods such as the MCA and LIA can only be achieved with this understanding of the forcings (e.g. Winter et al., 2011) and their spatial expression (Graham et al., 2011).

In addition to the forcings referred to above, many records from the NAM region contain periodicities consistent with short-term solar variability (which can itself be tracked by looking at changes in the production of cosmogenic isotopes such as ^{14}C and ^{10}Be). As well as the 11-year sunspot cycle, there are longer periodicities (80 – 90 yr, 208 yr and ~2300 yr) that may be climatically significant (Bond et al., 2001). Even though the exact mechanisms by which these variations are translated into changes in the climate system remain obscure, the most likely effect is through an impact on the Hadley Cell circulation, ITCZ location and monsoon strength (Fleitmann et al., 2003). Asmerom et al. (2007) suggest solar variability could affect the PDO – ENSO system, and Barron and Bukry (2007) link decreased sunspot numbers (low solar irradiance) to the enhanced development of NW winds over the GoC during the winter, which they argue decreased the likelihood of strong summer monsoons. Apparent cycles of around 200 years (consistent with the 208 year Suess or

DeVries cycle (Suess, 1980) have been particularly widely reported including marine records from the Cariaco Basin in the south (Peterson et al., 1991; Black et al., 2004), the GoM (Poore et al., 2003; Richey et al., 2007) and the GoC (Barron et al., 2005; Douglas et al., 2007). Amongst continental records, the 200-year cycle has also been recorded in Yucatan lakes (Hodell et al., 2001) and speleothems (e.g. Medina-Elizalde et al., 2010) in the south of the NAM region and from a bog record (Jimenez-Moreno et al., 2008) in the north. Schimmelmann et al. (2003) argue for a ca. 200-year periodicity in drought and flood events across the whole NAM region and extending into northern South America driven by solar variability, although the dating control in many sites is rather poor and the climatological case strongest for their sites in and off California. Across the NAM a range of other cycles have been reported, some of which may be solar (e.g. Jimenez-Moreno et al., 2008; Bernal et al., 2011). The 11-year sunspot cycle can only be captured by the highest resolution, well dated, records, being recorded in Cariaco (Black et al., 2004) and the GoC (Pike and Kemp, 1997). In contrast, Lachniet et al. (2004a) in their Panamanian speleothem suggest that ENSO forcing has been much more significant than solar forcing and do not find the 200 year cycle widely reported elsewhere. A reanalysis of the Chichancanab record (Carelton et al., 2014) has also disputed its occurrence there.

During the late Holocene changes in solar output have been related to changes in upwelling in the GoC (Barron and Bukry, 2007, see above), changes in SSTs in the GoM (Poore, 2008) and to changes in runoff. The correspondence of periods of solar minima with episodes of drought, has been reported, especially in the central and southern portions of the NAM (e.g. Hodell et al., 2001; Metcalfe et al., 2010), but they are also suggested as driving summer drought in the north of the area (Asmerom et al., 2013). It is also worth noting that human activity (deforestation) has itself been identified as a possible amplifier of drought, particularly for southern Mexico and the Yucatan (e.g. Oglesby et al, 2010; Cook BI et al., 2012), but given the limitations of climate models for this area (see below) some caution needs to be exercised in making these attributions.

The separation of shorter solar cycles from other forcings e.g. PDO, AMO across the wider NAM region is problematic given the continuing lack of high-resolution records and the need for seasonal resolution of the climatic signatures. Given the known interactions between the PDO and AMO and the uncertainties about the real impact of solar variability, these are all areas for future research.

Modelling

Climate models can provide important insights into the roles of different forcings of past climatic change at both global and continental scales, with much of the focus being on insolation (orbital forcing), the role of ice sheets, greenhouse gas concentrations and ocean feedbacks. Hydrological and vegetation feedbacks are topics of increasing study using the latest generations of coupled ocean-atmosphere models (Liu and Braconnot, 2012). The application of climate models to the NAM region could, therefore, help to address many of the questions about forcings discussed above. Modelling change in the monsoon was an early focus through the pioneering work of John Kutzbach (Kutzbach, 1981; Kutzbach and Guetter, 1986), which established that the early to mid Holocene NH summer insolation peak strengthened the NH summer monsoon. A systematic approach to data-model comparison was first developed by the Cooperative Holocene Mapping Project (COHMAP, 1988) and more recently by PMIP (Palaeoclimate Modelling Intercomparison Project) (Braconnot et al. 2007a and b). In both cases, a time slice approach was adopted often with a focus on the LGM and the mid Holocene (represented by 6000 cal yr BP). Although there is no doubt of the contribution of climate modelling to our understanding, their application to the NAM region has been problematic, as the grid scale of most models has been inadequate to capture the region's complex geography, topography and atmosphere – ocean interactions (Castro et al., 2012). Early attempts to model North America with what was basically an atmosphere only version of the NCAR CCM (e.g. Kutzbach, 1987) could not even capture the land connection between North and South America. The early Kutzbach papers concentrated on modelling July and January conditions for

18,000 and 9,000 years ago (relative to present) driven by changes in insolation, some variation in atmospheric composition and estimates of ice sheets, sea ice extent and SST. For the area defined as the SW, the modelling indicated increased winter rain and reduced evaporation at the LGM due a split jet stream coming south of the LIS. The jet apparently merged by about 12,000 BP but remained south of its modern location until around sometime between 9000 and 6000 BP. For the period 9000 – 6000 BP the model indicated an intensified summer monsoon, leading to conditions slightly wetter than present. Although broadly consistent with the rather limited palaeoclimatic data available at that time, it was clear that this scale of modelling could only offer limited insights into the behaviour of the NAM. It should be noted that later modelling of the LGM was more equivocal about the split jet idea, but Bartlein et al. (2014) state that it is 'one of the robust features of paleoclimatic simulations over North America (p.22) and that the general pattern persists into the early Holocene.

Using two models incorporating later versions of the Community Climate Model (CCM), Harrison et al. (2003) explored changes at 6000 cal yr BP across the monsoonal Americas using two coupled ocean-atmosphere models, FOAM and CSM. CSM had a higher resolution and hence a better representation of topography, but FOAM had a higher resolution representation of the ocean and because of its more efficient coding, fully coupled simulations could be run for longer. Both models were used to look at the climatic response to insolation forcing, while FOAM was used to look at the impact of SST feedbacks. Both models showed an increase in summer (JJA) precipitation in northern South America, Central America and the SW USA in response to increased NH insolation due to both a northward displacement of the ITCZ and a stronger monsoon. This area was surrounded to the north and east by a decrease in precipitation associated with increased subsidence (extending from the Pacific NW down to the eastern USA and Gulf of Mexico). The exploration of the impact of SST feedbacks showed the importance of these in enhancing onshore summer moisture transport over Central America and the west of North America, but with some drying over northern South America. The authors note some decoupling of changes over Central American and the ITCZ. Changes in the winter season were rather small in response to reduced insolation (compared to today), but with a small increase in precipitation over Central America and western coast of North America. Precipitation decreased over the continental interior. At the broad scale, these findings are not dissimilar from those of earlier studies, but with their finer spatial resolution showed an improved ability to reproduce observed climatologies. Comparison with our map for 6000 cal yr BP (Fig. 7) also indicates broad agreement, although in the north of the NAM region, the palaeodata indicate a rather mixed picture even in the area dominated by summer precipitation. The balance of effects between increased monsoonal precipitation and increased summer temperatures was clearly a fine one.

It is evident that finer spatial resolutions and fully coupled ocean-atmosphere models have improved our ability to model the climatology of the NAM region, although it is still far from perfect. Cook KH et al. (2012) report a comparison of the fully coupled Community Climate System version 4 (CCSM4) model with the earlier CCSM3 and an atmosphere only version 4 (CAM4). Both CAM4 and CCSM4 showed better representations of summer rainfall than earlier models, capturing low level westerly and south westerly flows from the Pacific, south easterlies from the GoM and with better representation of orographic precipitation driven by the highlands of western Mexico (the Sierra Madre Occidental). In spite of this, all the models used in the comparison overestimated precipitation across the NAM area over the whole year, for CCSM4 this overestimation was particularly pronounced in the autumn and winter (although the overestimation was less than in CCSM3). This overestimation affects the models' ability to reproduce the decay of the monsoon in September and October.

Given the geographical complexity of the NAM region, it would seem that application of regional climate models could be particularly valuable. Castro et al. (2012) have applied a regional climate model (35 km grid) to see whether this approach can improve forecasting of the NAM summer season compared with coarser resolution (200 km grid) Climate Forecast System model. The regional model used two forms of downscaling from the standard model. It should be noted that here the NAM region is as defined by the North American Monsoon Experiment (NAME) study being limited to the south west USA and northwest Mexico. The study shows that the regional model does provide an improved simulation of precipitation because of the better representation of mesoscale processes. However, features such as the GoC low level jet and moisture surges up the Gulf are not captured, nor are convective precipitation. The authors also report that the regional scale model was better at capturing monsoon response to changes in Pacific SSTs (El Nino-like SSTs = dry, La Nina-like SSTs = wet). Another application of a regional climate model Reg-CM4) to Central America and Mexico (Fuentes-Franco et al., 2014) also showed the improved capabilities of this type of model, reproducing the annual cycle of precipitation in different areas of Mexico, including the mid summer drought in the south of Mexico and Yucatan areas. Even at this scale however, there were areas when the model could not reproduce observed (daily) precipitation, most notably in the summer, with over estimation in southern Mexico (including parts of the TMVB) and underestimation over the Yucatan Peninsula. To date, there has been no attempt to model palaeoclimates of the NAM region at anything approaching the resolution of these regional climate models, but it seems that models of this spatial resolution will be required to capture the complexity indicated by the palaeodata.

The more detailed records included in this synthesis (and any not included) indicate that we are now well placed for a renewed effort on data-model comparisons across the NAM region.

5. Conclusions

Using information from both continental and marine records, we have reviewed evidence for climatic change in the wider NAM region over the Holocene and considered the forcing mechanisms that might give rise to observed patterns of variability. Over the decade or so since the last major reviews of the palaeoclimate of this area, there has been a significant increase in the number of published records, particularly in certain areas such as the Yucatan Peninsula, and the exploitation of a wider range of archives, specifically speleothems and tree rings. In continental records we have yet to see much use of organic geochemical methods such as analysis of plant leaf waxes (δD_{wax}) or lacustrine algae (δD), which could yield more insights into precipitation and evapotranspiration (but see recent examples in Kirby et al., 2013 and Lane et al., 2014). Temperature reconstructions using GDGTs (glycerol dialkyl glycerol tetraether) and other markers have been assessed, but not widely applied. Sadly, problems of security have restricted recent research in significant areas of central and northern Mexico. The spatial coverage of records covering the whole of the Holocene is, however, still very uneven and radiocarbon chronologies are often not well constrained. In the GoC and along the Pacific margins, high-resolution late Holocene marine records typically lack detailed radiocarbon chronologies. The problems posed by hardwater error and catchment disturbance and reworking remain, even with the advantages of AMS dating. It is probably also fair to say that in many cases the resolution of the chronologies has not kept pace with the resolution of the palaeoenvironmental data sets that are being generated. Speleothem records suitable for U-series dating and tree ring records are obvious exceptions to this. In keeping with the general trend, there has been a move to applying a wider range of methods to individual sequences, but there is no doubt that the rate at which geochemical and isotopic records can be generated, compared with palaeoecological methods, means that for lake sediment cores in particular, the former are increasingly dominating over the latter. It is evident from reviewing the literature that one of greatest challenges for reconstructing climate over the NAM region is the need for methods that discriminate seasonally. In the vast majority of cases, seasonality is inferred rather than reconstructed directly. Many of the key questions about change in the NAM and its forcings require high resolution, well dated and seasonally resolved records.

The nature of the transition from the LGM into the Holocene remains rather unclear (see also Metcalfe et al., 2000), with particular complexity in the northern region of the NAM where the relative importance of winter and summer (monsoonal) rainfall, remains uncertain with different authors making cases for both. The lingering presence of the LIS undoubtedly had an impact probably both delaying the response to the NH insolation maximum and around the last glacial – interglacial transition affecting atmospheric circulation directly. The possible impact of glacial meltwater entering the GoM is discussed in the Forcings section. The absence of such events from the GoC and whether this was reflected in an earlier reestablishment of the summer monsoon in the Pacific (western) part of the NAM is worthy of consideration. A number of papers have linked their records to events originating in the North Atlantic, but to date clear evidence for things such as the 8.2 k 'event' has been relatively limited, with changes spanning 8200 cal yr BP or using indirect evidence (e.g. Lachniet et al., 2004b; Lozano-Garcia and Vazquez-Selem, 2005). Lozano-Garcia et al., 2013 report a change around 8200 cal yr BP in their pollen record, but not at the scale of change that they reported at 7500 cal yr BP. This lack of clarity may be related to dating control over this period or sampling resolution, or may simply reflect that this event was not strongly expressed uniformly over the NAM region. Over the Holocene as a whole, there is evidence for the classical insolation forcing of the NAM, also associated with migration of the ITCZ. It seems that the summer rainfall dominated monsoon system reached its greatest geographical extent around 6000 cal yr BP (Fig. 7), although this was not the peak of moisture availability in the central and northern regions, where increased precipitation was counteracted by higher temperatures and greater evapotranspiration. The period around 4000 cal yr BP (Fig. 8) marks a significant change in the climatology of the NAM region as NH summer and autumn insolation declined and the ITCZ tracked south. This seems to be the time when the modern antiphase pattern between north and south emerged, with summer rain continuing to dominate in the south, but with winter rain or a bimodal regime, dominating in the north. It seems that this pattern was associated with the onset of stronger ENSO-type variability. These differences in rainfall seasonality explain the variable responses to forcings such as PDO/ENSO and AMO/NAO across the NAM region and hence, some of the patterns seen at the MCA and the LIA (Figs. 10 and 11). The central region (roughly central Mexico) lies at the boundary between these two regions, probably more often following the southern pattern, but this does not seem to be stationary in time (i.e., the MCA and LIA). Strong insolation forcing led to more coherent behaviour across the whole NAM region and this became less common as other forcings became more important. As Stahle et al. (2012a) point out, however, there are still periods in the recent past when the north (their Southwest) and south/central (their Mesoamerican) converge as the interactions of different forcings give rise to similar responses.

Questions about the relationship between climate and people over the last 2000 years have become of particular interest in the NAM area. The periodic apparent 'collapses' of a number of Mesoamerican cultures (most famously the Maya) have prompted great interest in the possible role of climate, specifically drought, in these social and cultural events. Similar themes have been explored in relation to Native American cultures in the SW USA (including the Anasazi) during the period of the MCA (Benson et al., 2007). Whilst there is no doubt that our views on this have become more nuanced (e.g. Aimers and Hodell's News & Views pieces in Nature, 2011), there can still be a temptation to overreach the limits of our palaeoclimatic data and ignore the complexity and individuality of particular people at particular times (Butzer, 2012).

In our earlier review paper (Metcalfe et al., 2000) we suggested that the time was ripe for using palaeoclimate models to help us understand the palaeo-data from the NAM region. It still is. Unfortunately, although modelling the modern NAM has made considerable advances (see above), the application of high-resolution, fully coupled models to the past climates of the region remains to be done. Although the difficulties of such modelling should not be underestimated, further dialogue between climatologists, modellers and the palaeoclimate community would be fruitful for setting the agenda for future research.

Perhaps unsurprisingly, there remains a great deal about the Holocene history of the NAM that we do not understand. We have more records, from more places and we are getting better at integrating what we find on the continents and in the oceans, although in these two contexts we are often reconstructing different aspects of the environment (e.g. water balance in the former and temperature in the latter). In the range of palaeoenvironmental archives, differences in inherent and analytical resolution remain, as does the challenge of dating. The quantification of change also remains in its infancy. At an individual site level, we often have a much more detailed understanding of its climatic and wider environmental history now than a decade or so ago, but that added complexity can make the task of broad syntheses such as this even more challenging. Patterns at the broad scale will subsume a great deal of local variability, which reflects the sensitivity and resilience of both the natural and human environment. Much of the NAM region is under severe pressure from both human impact (particularly related to water abstraction) and current climate change. Whilst extreme care is needed in claiming that the past can help us cope with the future, better understanding about the range of natural variability and the sensitivity and resilience of different elements of the environment to climate change over different timescales must be valuable.

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Tables

Table 1 Continental records referred to in the text and figures with key references. See Fig. 4 for locations.

Table 2 Sea surface temperature and ENSO conditions for marine records referred to in the text and figures with key references. X = no record. See Fig. 4 for locations.

Supplementary Table 1. Continental records referred to in the text and figures with key references. Sites with 3 or more data points within ± 200 years shown in bold, others in standard font. Key for colour coding and other symbols given on the table.

Figures

Fig. 1. Climatology of the North American Monsoon region (a) summer, (b) winter.

Fig. 2. a). North American Monsoon (NAM) climatology showing low level moisture flow (LLF) into the southwest US from the Gulf of California and Gulf of Mexico that is constrained by the Sierra Madre Occidental and Sierra Madre Oriental mountains in Mexico. Northward movement of the ITCZ fuels tropical moisture surges (especially in the Pacific). The North Pacific (NP) and Bermuda (BH) high-pressure systems constrain this tropical moisture. Mid level moisture from the Gulf of Mexico supplies much of the summer precipitation in Mexico and the northern NAM region. b). Cross section across northern Mexico showing moisture transport into the NAM. LLJ = Low Level Jet. Officially "Schematic vertical section for the corresponding summer season east-west at about 30°N across the NAM (left panel) and southwest-northeast across SAM region. Regions of deep convection and low-level jets are indicated (Panel for NAMS adapted from W. Higgins)" – Introduction to Tropical Meteorology (http://www.goes-r.gov/users/comet/tropical/textbook_2nd_edition/)

Fig. 3. Total annual precipitation (a) and % precipitation July – September (b) across the North American Monsoon area and wider region (after Ropelewski et al. 2005).

Fig. 4. Location of continental records referred to in this paper (see key in Table 1). Tree ring records not included here. Shading represents areas mainly > 2000 m a.s.l.

Fig. 5. Terrestrial proxy response at 12,000 cal yr BP (records +/- 200 yrs). Different symbols for different types of archive and different numbers of data points per time period (see Table 1 Supplementary Material). Solid fill symbols used for sites with 3 or more data points within ± 200 years of chosen date.

Fig. 6. Terrestrial proxy response at 9000 cal yr BP (records +/- 200 yrs). Key to symbols on Fig. 5.

Fig. 7. Terrestrial proxy response at 6000 cal yr BP (records +/- 200 yrs). Key to symbols on Fig. 5.

Fig. 8. Terrestrial proxy response at 4000 cal yr BP (records +/- 200 yrs).   Key to symbols on Fig. 5.

Fig. 9. Compilation of selected proxy precipitation and SST records for the past 2000 years. (Cariaco = Haug et al., 2001; Lago El Gancho = Stansell et al., 2012; Yok Balum = Kennett et al., 2012; Juanacatlan = Metcalfe et al., 2010; GoM MD2553 = Poore et al., 2011; GoM PBBC-1 = Richey et al., 2007; BAM80 E17 = Barron & Bukry, 2007; Bat Cave = Asmerom et al., 2013; El Malpais = Stahle et al., 2012b). Pink shading = Medieval Climate Anomaly; blue shading = Little Ice Age; grey shading = solar minima.

Figs 10. Terrestrial proxy response during the Medieval Climate Anomaly (AD 900—1300). Key to symbols on Fig. 5.

Fig. 11. Terrestrial proxy response during the Little ice Age (AD 1400-1850). Key to symbols on Fig. 5.

Fig .12. Changes in seasonal insolation (December, March, June, September) as a percentage of average over the last 12,000 years for 30°N. (insolation data after Berger and Loutre, 1991)

Fig. 13. Comparison of the Holocene Cariacco Basin %Ti record of Haug et al. (2001), the GoM proxy SST records of LoDico et al. (2006) and Poore et al. (2001), and the cave oxygen isotope records from Pink Panther Cave in New Mexico (Asmerom et al., 2007) and Buckeye Creek Cave in West Virginia (Hardt et al., 2010). These are referenced against June insolation at 30°N (Berger and Loutre, 1991) and Shuman et al.'s (2005) estimate of the size of the Laurentide Ice Sheet. (Pink Panther isotope data is reinterpreted –see Barron et al., 2012). These suggest that maximum SSTs between ~8000 and 5000 cal yr BP coincided with maximum transport of monsoon moisture from the GoM to New Mexico. Prior to ~8000 cal yr BP, the LIS likely forced the Bermuda High to the south. The NAM region was dominated by the location of the ITCZ.

Fig. 14. Comparison of Holocene El Niño frequency and ENSO conditions in the equatorial Pacific with SST reconstructions in the eastern tropical Pacific, the GoC, and Pacific waters off Baja California (Mexico). MV99-GC31 (after Barron et al., 2012). Pink shading = warm SST; blue shading = cool SST; yellow shading = ENSO dominated SSTA after 4000 to 3000 cal yr BP; green shading = peak Monsoon in GoC region (Barron et al., 2012) or Pacific Forcing - Between ~8000 and 4000 cal yr BP, enhanced La Niña-like SSTs in the equatorial Pacific and widespread winter drought in the southwest US corresponded with maximum flow of Pacific-based NAM moisture into northwest Mexico and the southwest US (Barron et al., 2012, Paleoceanogr). After 4000 cal yr BP, increasing ENSO variability (more frequent El Niño's) corresponded with a more southerly mean position of the Pacific ITCZ, decreased flow of Pacific-based monsoonal precipitation into the southwest US, and a condensed geographic pattern of NAM precipitation. Modern regional differences in climatology associated with changing ENSO and AMO conditions were established between 4000 and 3000 cal yr BP.

Acronyms used in this paper

AMO	Atlantic Multidecadal Oscillation
AMS	Accelerator Mass Spectrometry
CAM	Community Atmosphere Model
CCM	Community Climate Model
CCS	Community Climate System
COHMAP	Cooperative Holocene Mapping Project
CSM	Climate System Model
ENSO	El Nino Southern Oscillation
FOAM	Fast Ocean Atmosphere Model
GoC	Gulf of California
GoM	Gulf of Mexico
IAS	Intra Americas Sea
ITCZ	Intertropical Convergence Zone
JJA	June July August
LIA	Little Ice Age
LIS	Laurentide Ice Sheet
MCA	Medieval Climate Anomaly
NAM	North American Monsoon
NAME	North American Monsoon Experiment
NAO	North Atlantic Oscillation
NH	Northern Hemisphere
PDO	Pacific Decadal Oscillation
PMIP	Palaeoclimate Modelling Intercomparison Project
Reg-CM	Regional Climate Model
SSTs	Sea surface temperatures
TCD	Terminal Classic Drought
TMVB	Trans Mexican Volcanic Belt

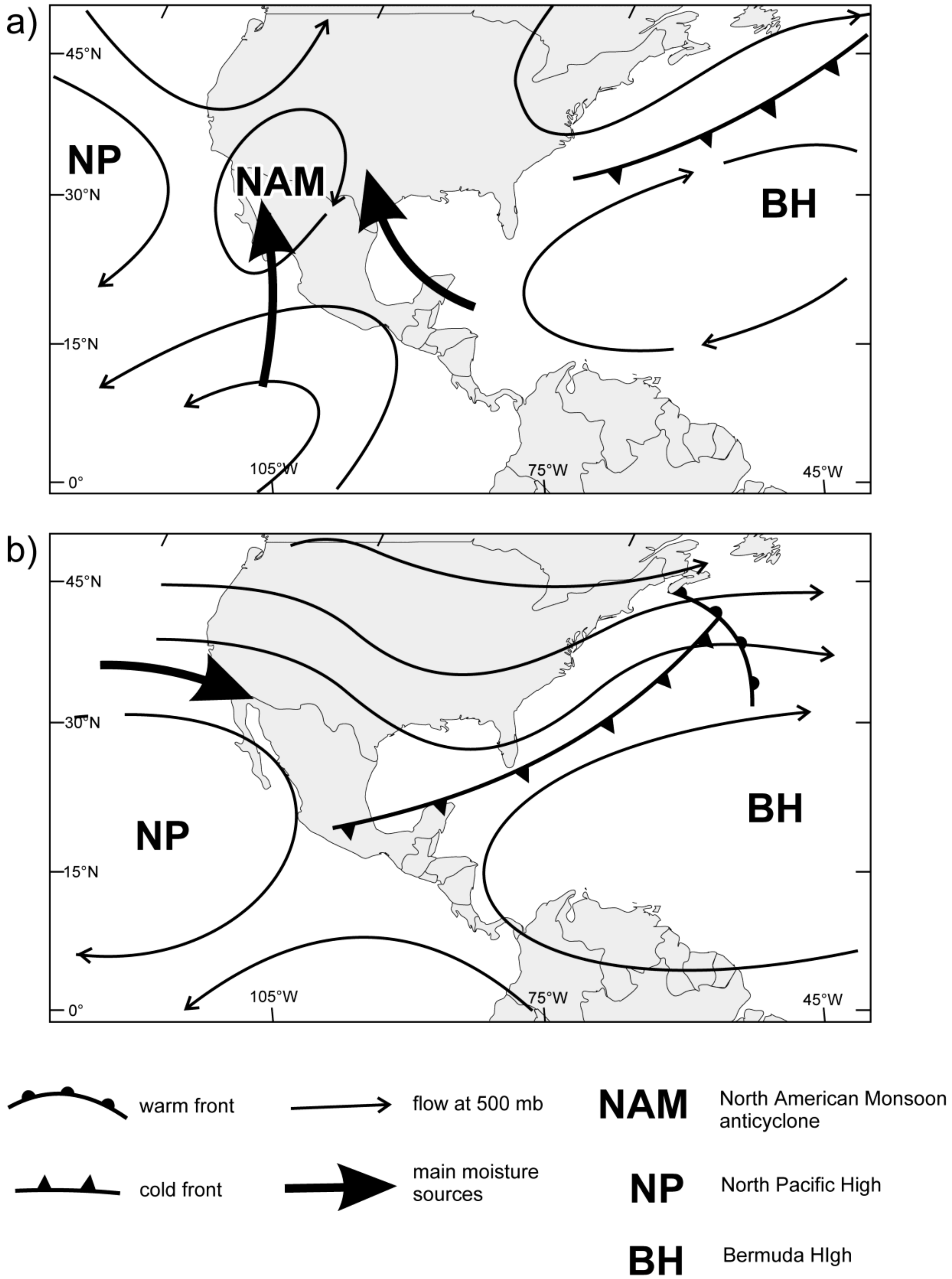

a)
45°N
30°N
15°N
0°
105°W
75°W
45°W
NP
NAM
BH
b)
45°N
30°N
15°N
0°
105°W
75°W
45°W
NP
BH
warm front
flow at 500 mb
NAM
North American Monsoon anticyclone
cold front
main moisture sources
NP
North Pacific High
BH
Bermuda HIgh

Fig. 1

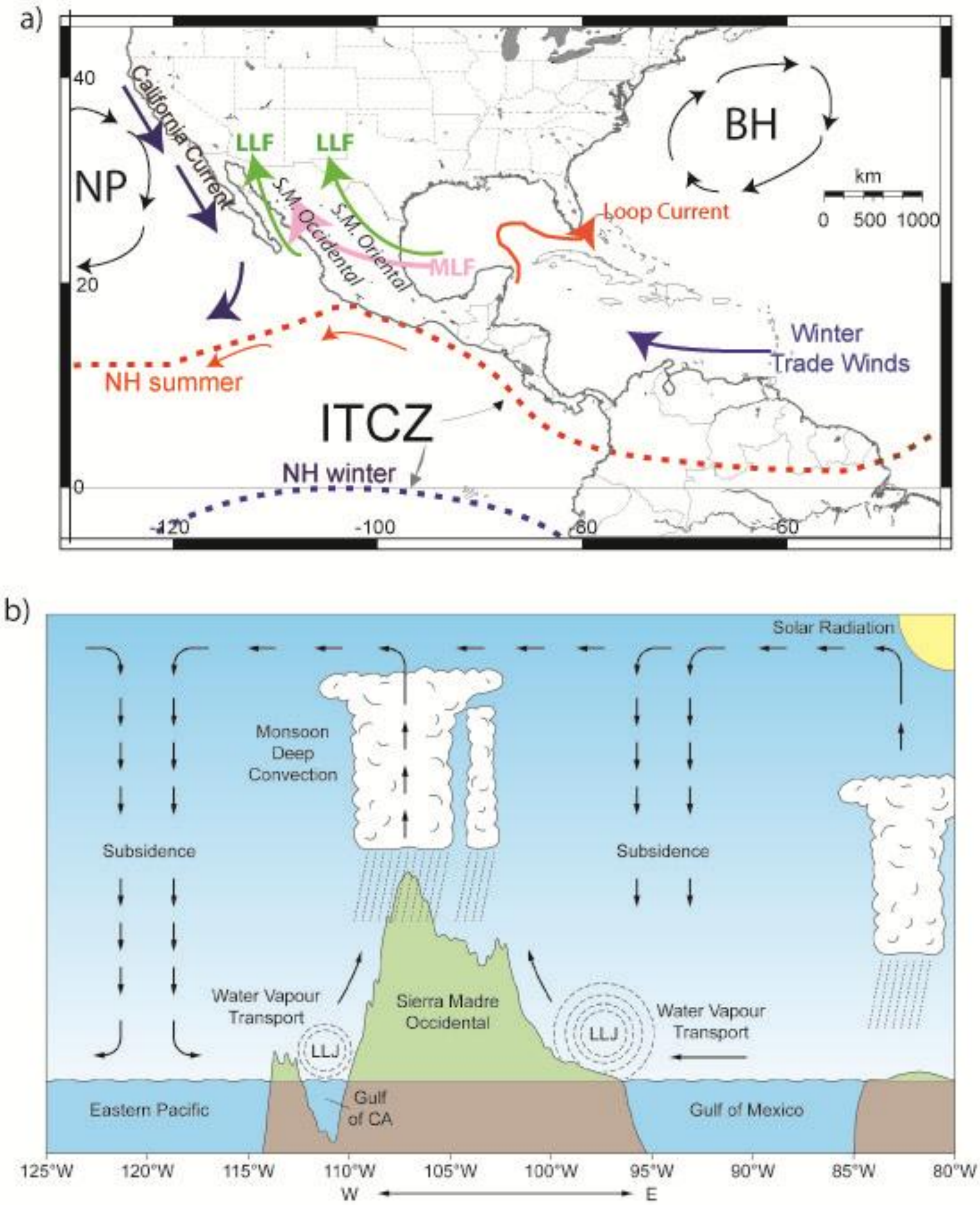
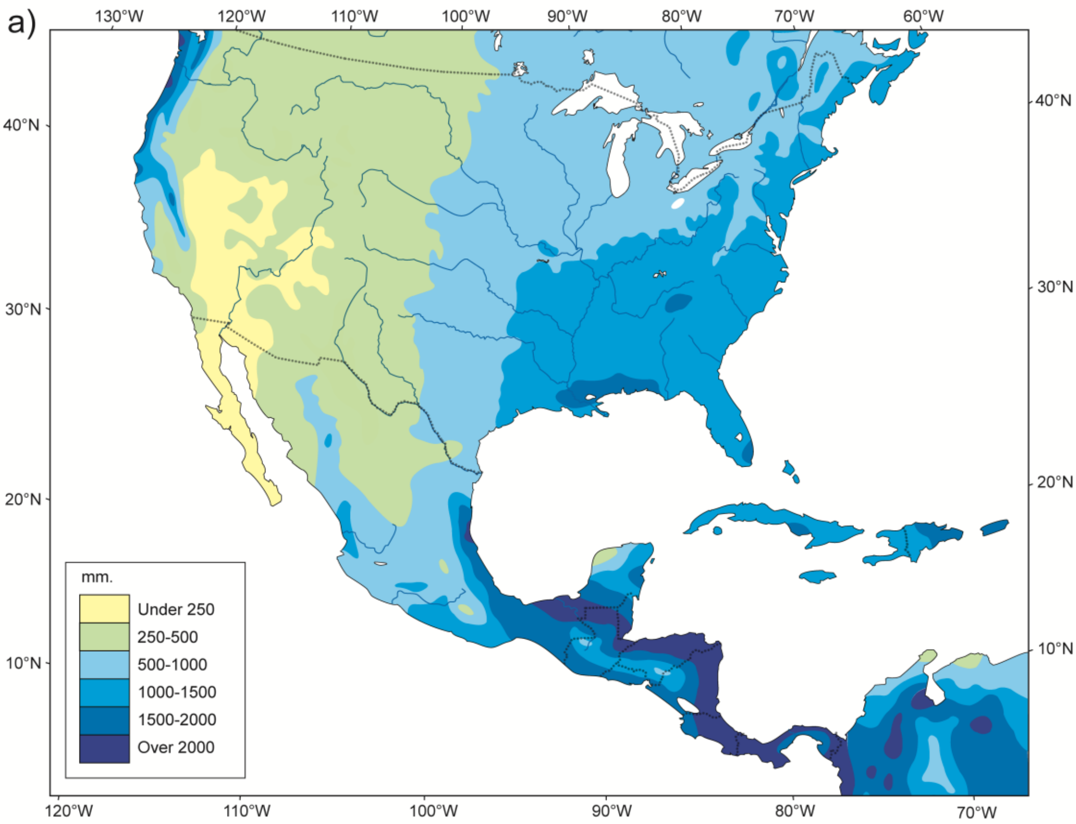

a)
NP
California Current
LLF
LLF
S.M. Occidental
S.M. Oriental
MLF
BH
Loop Current
km
0 500 1000
Winter
Trade Winds
NH summer
ITCZ
NH winter
40
20
0
-120
-100
-80
-60
b)
Solar Radiation
Monsoon
Deep
Convection
Subsidence
Subsidence
Water Vapour
Transport
Sierra Madre
Occidental
LLJ
LLJ
Water Vapour
Transport
Eastern Pacific
Gulf
of CA
Gulf of Mexico
125°W
120°W
115°W
110°W
105°W
100°W
95°W
90°W
85°W
80°W
W
E

Fig. 2


a)
130°W
120°W
110°W
100°W
90°W
80°W
70°W
60°W
40°N
30°N
20°N
10°N
mm.
Under 250
250-500
500-1000
1000-1500
1500-2000
Over 2000

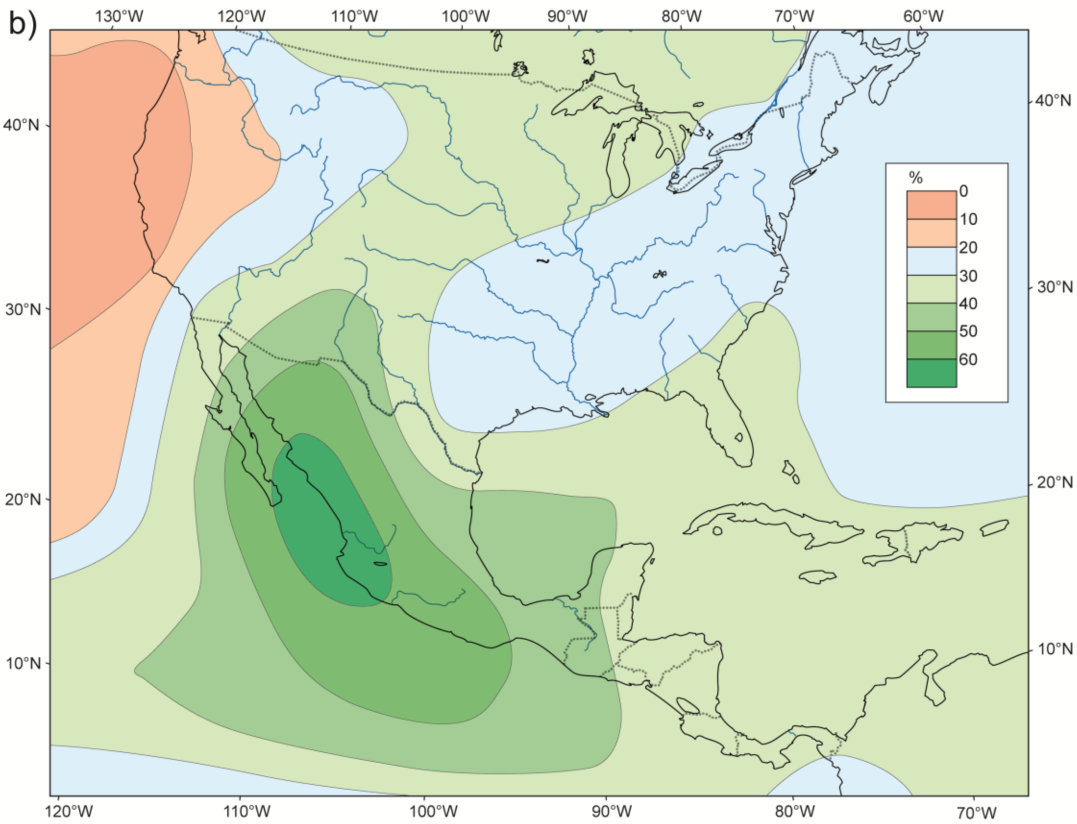

b)
130°W
120°W
110°W
100°W
90°W
80°W
70°W
60°W
40°N
30°N
20°N
10°N
%
0
10
20
30
40
50
60

Fig. 3

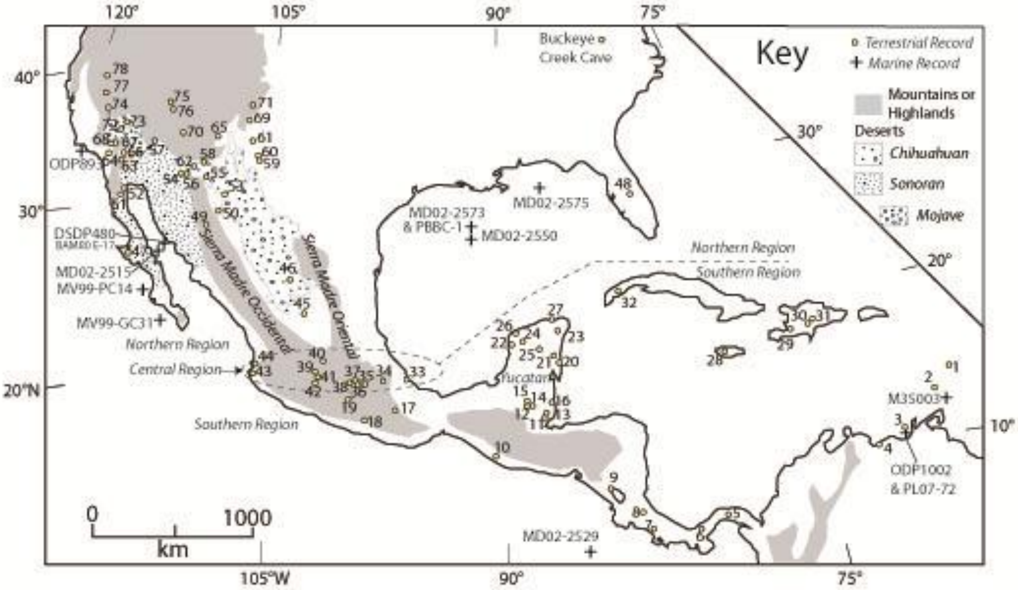

Key
Terrestrial Record
Marine Record
Mountains or Highlands
Deserts
Chihuahuan
Sonoran
Mojave
Buckeye Creek Cave
ODP893
DSDP480
BAM80 E-17
MD02-2515
MV99-PC14
MV99-GC31
Northern Region
Central Region
Southern Region
Sierra Madre Occidental
Sierra Madre Oriental
MD02-2573 & PBBC-1
MD02-2550
MD02-2575
MD02-2529
Northern Region
Southern Region
Yucatan
M35003
ODP1002 & PL07-72
0
1000
km
20°N
105°W

Fig. 4

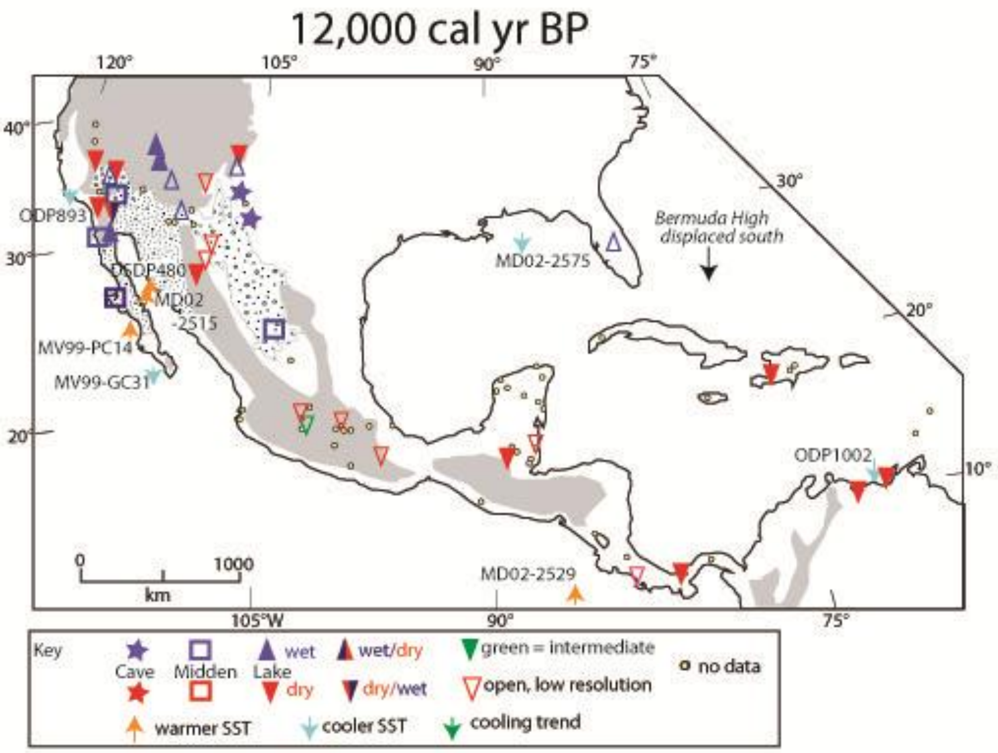

12,000 cal yr BP
120°
105°
90°
75°
40°
30°
20°
10°
ODP893
DSDP480
MD02
-2515
MV99-PC14
MV99-GC31
MD02-2575
Bermuda High
displaced south
ODP1002
MD02-2529
0
1000
km
105°W
Key
Cave
Midden
Lake
wet
wet/dry
green = intermediate
no data
dry
dry/wet
open, low resolution
warmer SST
cooler SST
cooling trend

Fig. 5

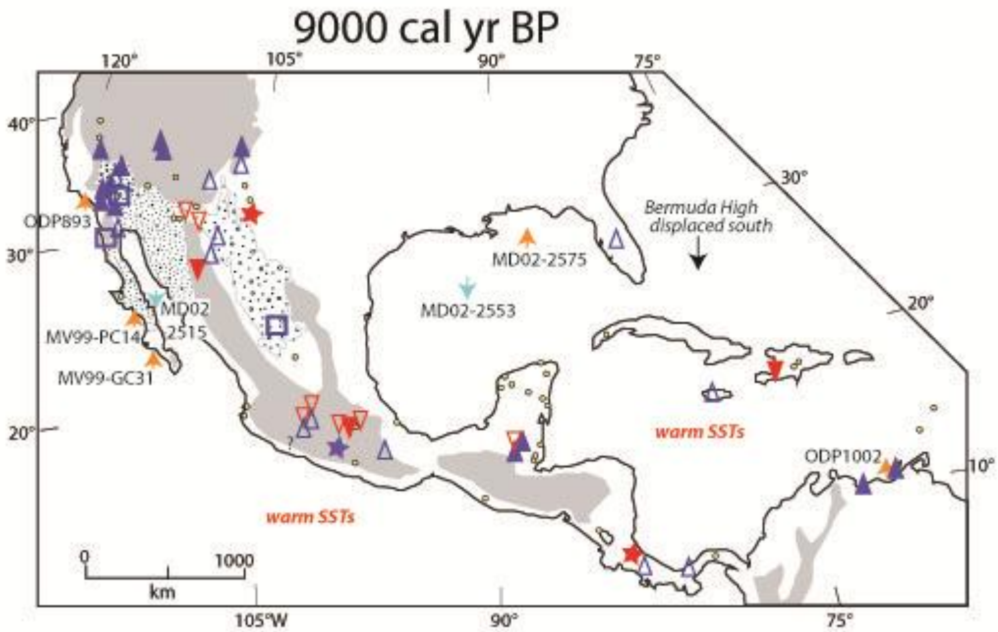

9000 cal yr BP
120°
105°
90°
75°
40°
30°
20°
30°
20°
10°
ODP893
MD02
2515
MV99-PC14
MV99-GC31
MD02-2575
MD02-2553
Bermuda High
displaced south
warm SSTs
warm SSTs
ODP1002
0
1000
km
105°W
90°
75°

Fig. 6

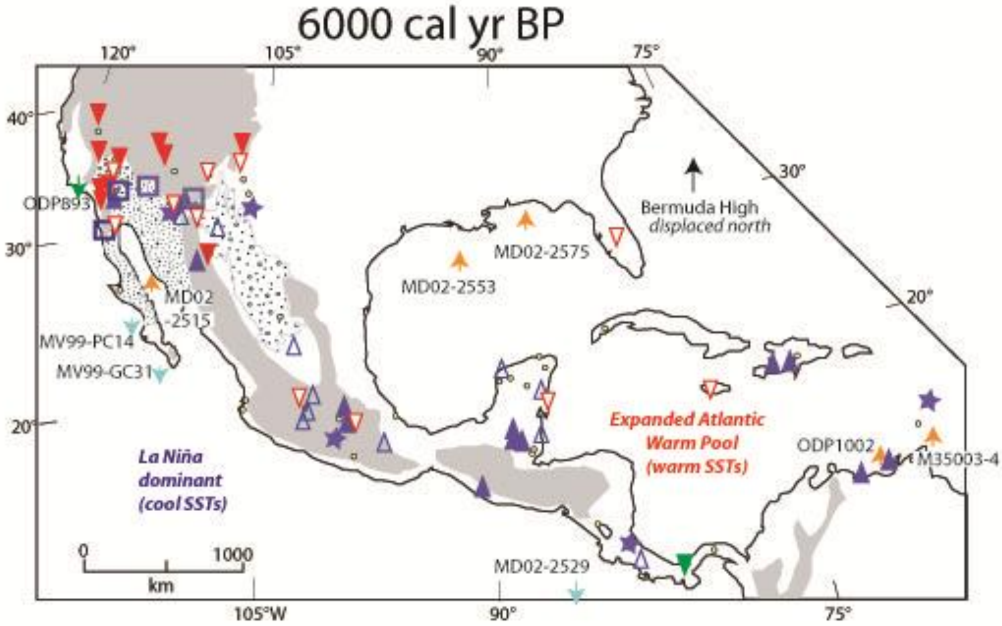

6000 cal yr BP
120°
105°
90°
75°
40°
30°
20°
30°
20°
ODP893
MD02
-2515
MV99-PC14
MV99-GC31
MD02-2553
MD02-2575
Bermuda High
displaced north
Expanded Atlantic
Warm Pool
(warm SSTs)
La Niña
dominant
(cool SSTs)
ODP1002
M35003-4
MD02-2529
0
1000
km
105°W
90°
75°

Fig. 7

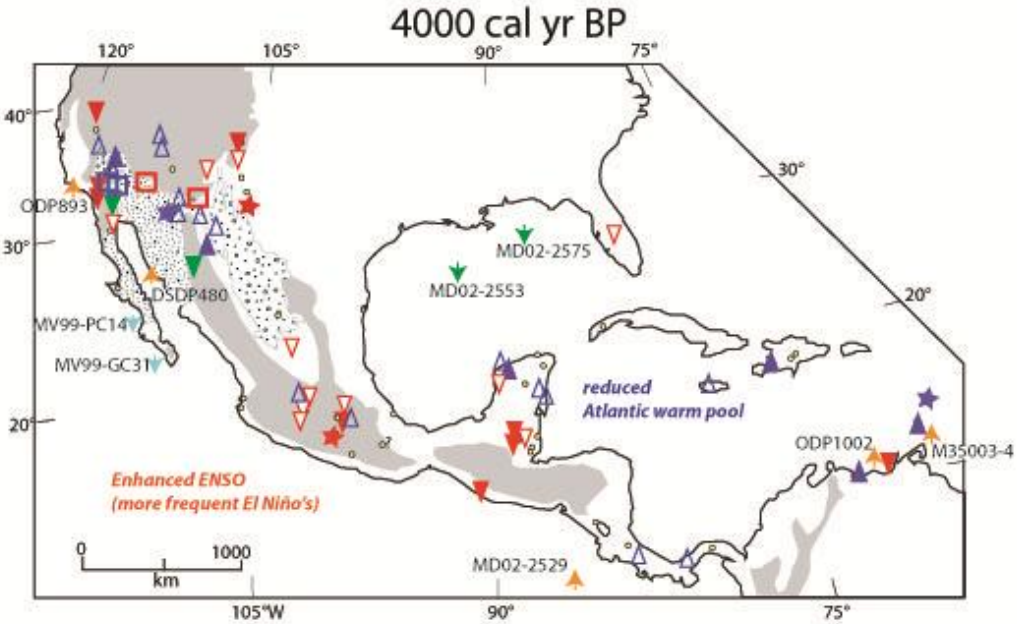

4000 cal yr BP
120°
105°
90°
75°
40°
30°
20°
30°
20°
ODP893
DSDP480
MV99-PC14
MV99-GC31
MD02-2575
MD02-2553
reduced
Atlantic warm pool
ODP1002
M35003-4
Enhanced ENSO
(more frequent El Niño's)
0
1000
km
MD02-2529
105°W
90°
75°

Fig. 8

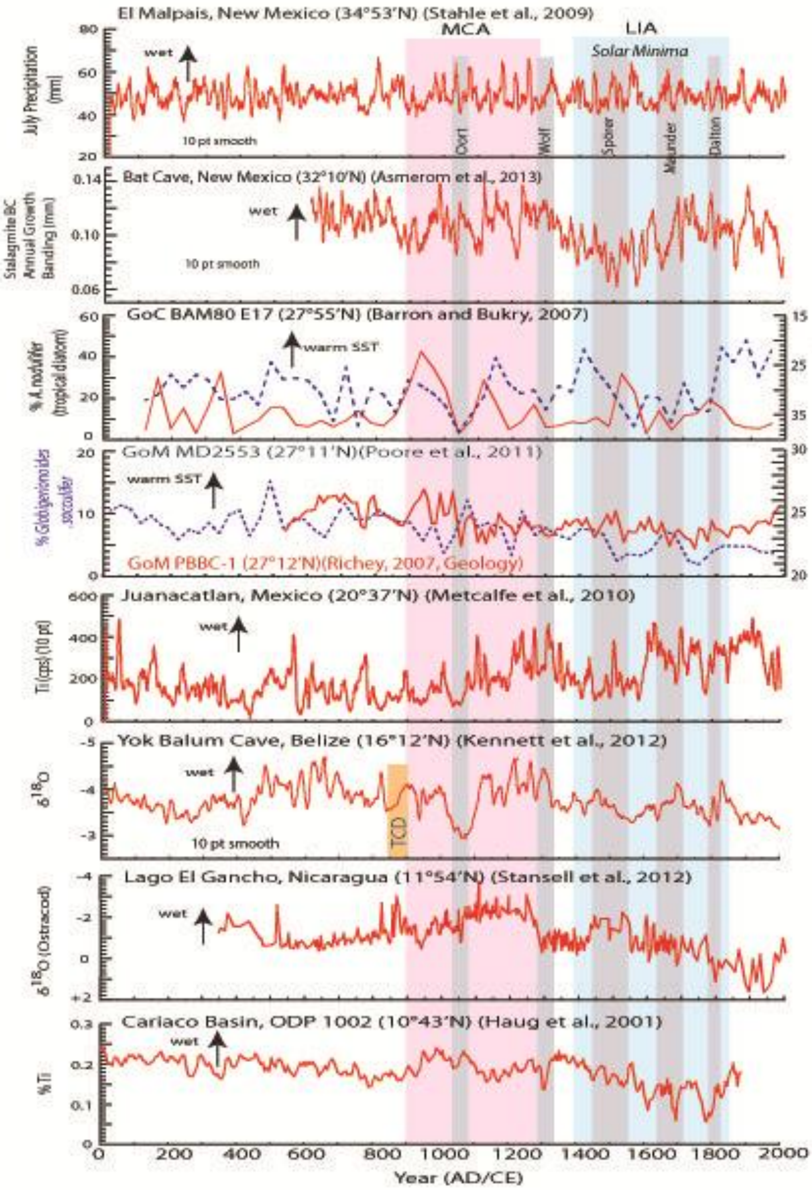

El Malpais, New Mexico (34°53'N) (Stahle et al., 2009)
MCA
LIA
Solar Minima
July Precipitation (mm)
wet
10 pt smooth
Oort
Wolf
Sporer
Maunder
Dalton
Bat Cave, New Mexico (32°10'N) (Asmerom et al., 2013)
Stalagmite BC Annual Growth Banding (mm)
GoC BAM80 E17 (27°55'N) (Barron and Bukry, 2007)
% A. nodulifer (tropical diatom)
warm SST
% Biogenic Silica
GoM MD2553 (27°11'N)(Poore et al., 2011)
% Globigerinoides sacculifer
Mg/Ca SST (°C)
GoM PBBC-1 (27°12'N)(Richey, 2007, Geology)
Juanacatlan, Mexico (20°37'N) (Metcalfe et al., 2010)
Ti (cps) (10 pt)
Yok Balum Cave, Belize (16°12'N) (Kennett et al., 2012)
$\delta^{18}O$
TCD
Lago El Gancho, Nicaragua (11°54'N) (Stansell et al., 2012)
$\delta^{18}O$ (Ostracod)
Cariaco Basin, ODP 1002 (10°43'N) (Haug et al., 2001)
%Ti
Year (AD/CE)

Fig. 9

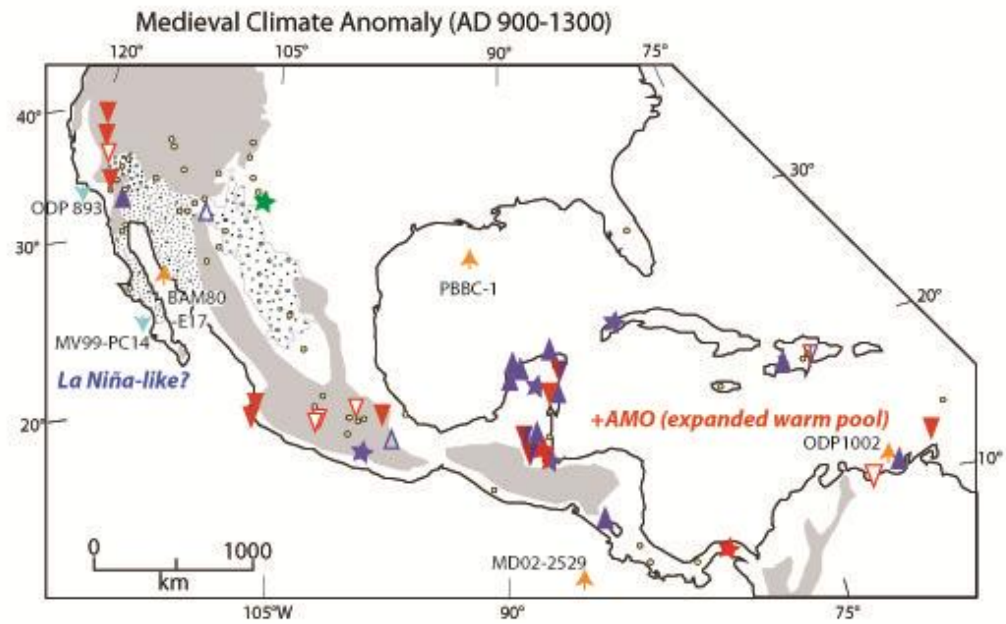

Medieval Climate Anomaly (AD 900-1300)
120°
105°
90°
75°
40°
30°
20°
30°
20°
10°
ODP 893
BAM80
-E17
MV99-PC14
La Niña-like?
PBBC-1
+AMO (expanded warm pool)
ODP1002
MD02-2529
0
1000
km
105°W
90°
75°

Fig. 10

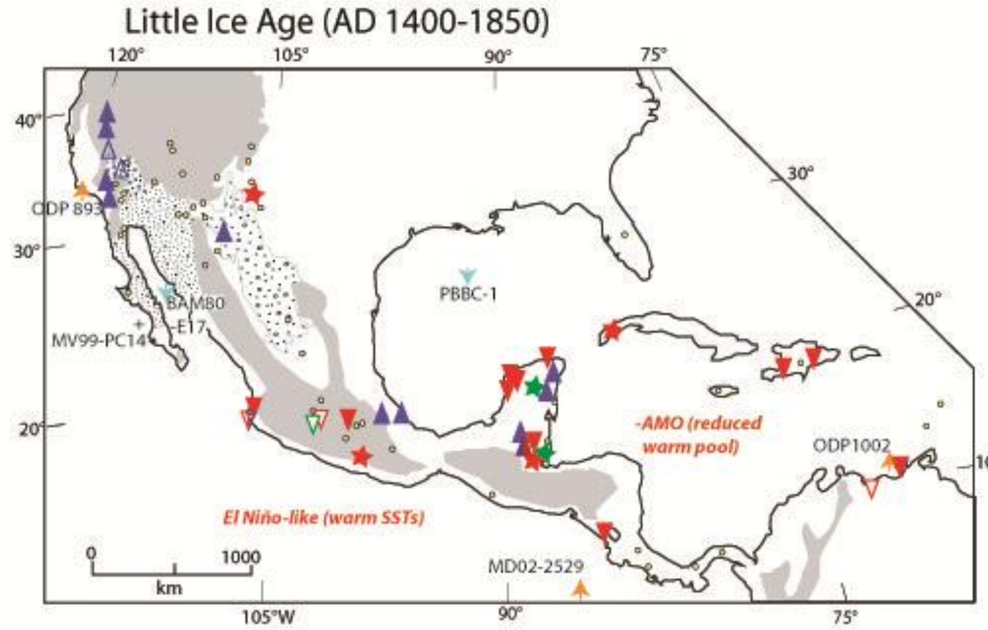
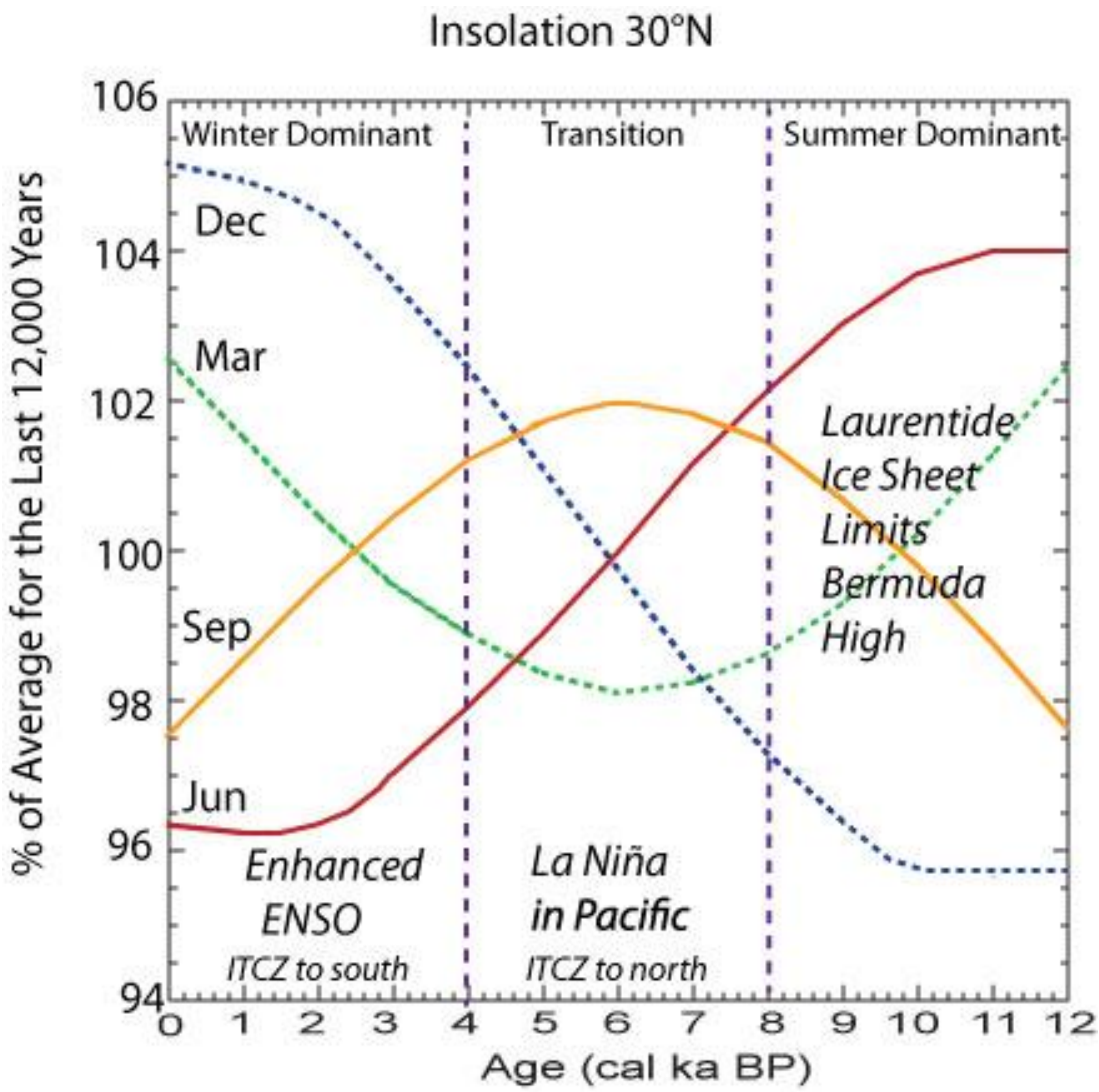
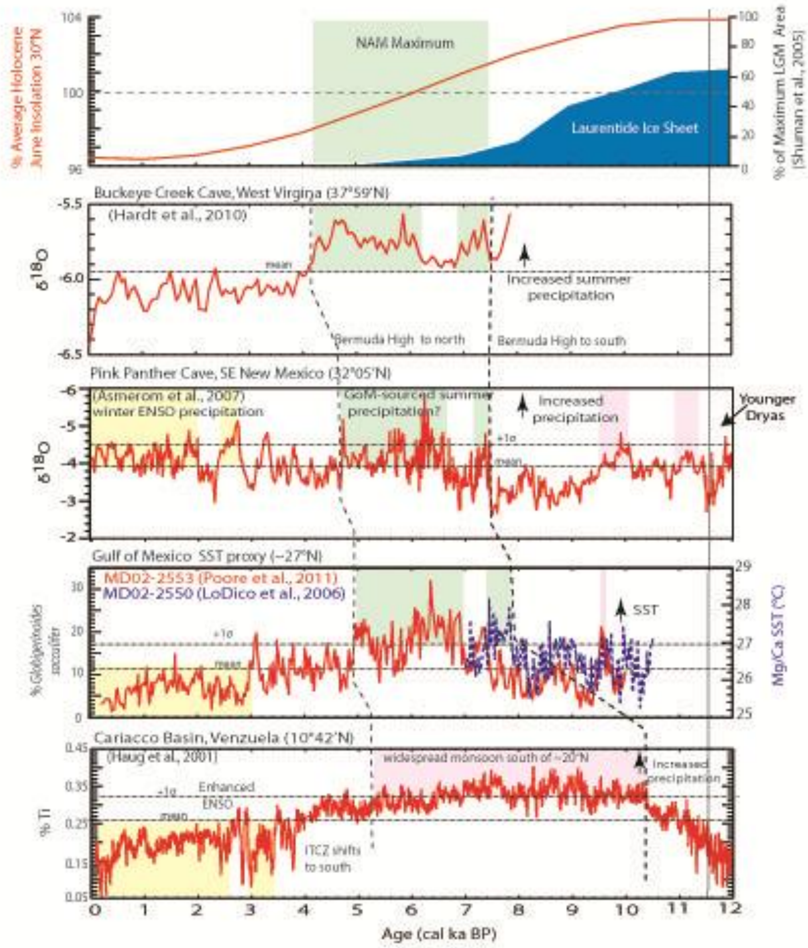

Little Ice Age (AD 1400-1850)
120°
105°
90°
75°
40°
30°
20°
30°
20°
10°
ODP 893
BAM80
E17
MV99-PC14
PBBC-1
-AMO (reduced
warm pool)
ODP1002
El Niño-like (warm SSTs)
0
1000
km
MD02-2529
105°W
90°
75°

Fig. 11



Insolation 30°N
% of Average for the Last 12,000 Years
Age (cal ka BP)
Winter Dominant
Transition
Summer Dominant
Dec
Mar
Sep
Jun
Enhanced
ENSO
ITCZ to south
La Niña
in Pacific
ITCZ to north
Laurentide
Ice Sheet
Limits
Bermuda
High
106
104
102
100
98
96
94
0
1
2
3
4
5
6
7
8
9
10
11
12

Fig. 12



% Average Holocene June Insolation 30°N
NAM Maximum
Laurentide Ice Sheet
% of Maximum LGM Area [Shuman et al., 2005]
Buckeye Creek Cave, West Virgina (37°59'N)
(Hardt et al., 2010)
Increased summer precipitation
Bermuda High to north
Bermuda High to south
Pink Panther Cave, SE New Mexico (32°05'N)
(Asmerom et al., 2007)
winter ENSO precipitation
GoM-sourced summer precipitation?
Increased precipitation
Younger Dryas
Gulf of Mexico SST proxy (~27°N)
MD02-2553 (Poore et al., 2011)
MD02-2550 (LoDico et al., 2006)
SST
Mg/Ca SST (°C)
Cariacco Basin, Venzuela (10°42'N)
(Haug et al., 2001)
widespread monsoon south of ~20°N
Increased precipitation
Enhanced ENSO
ITCZ shifts to south
% Ti
Age (cal ka BP)

Fig. 13

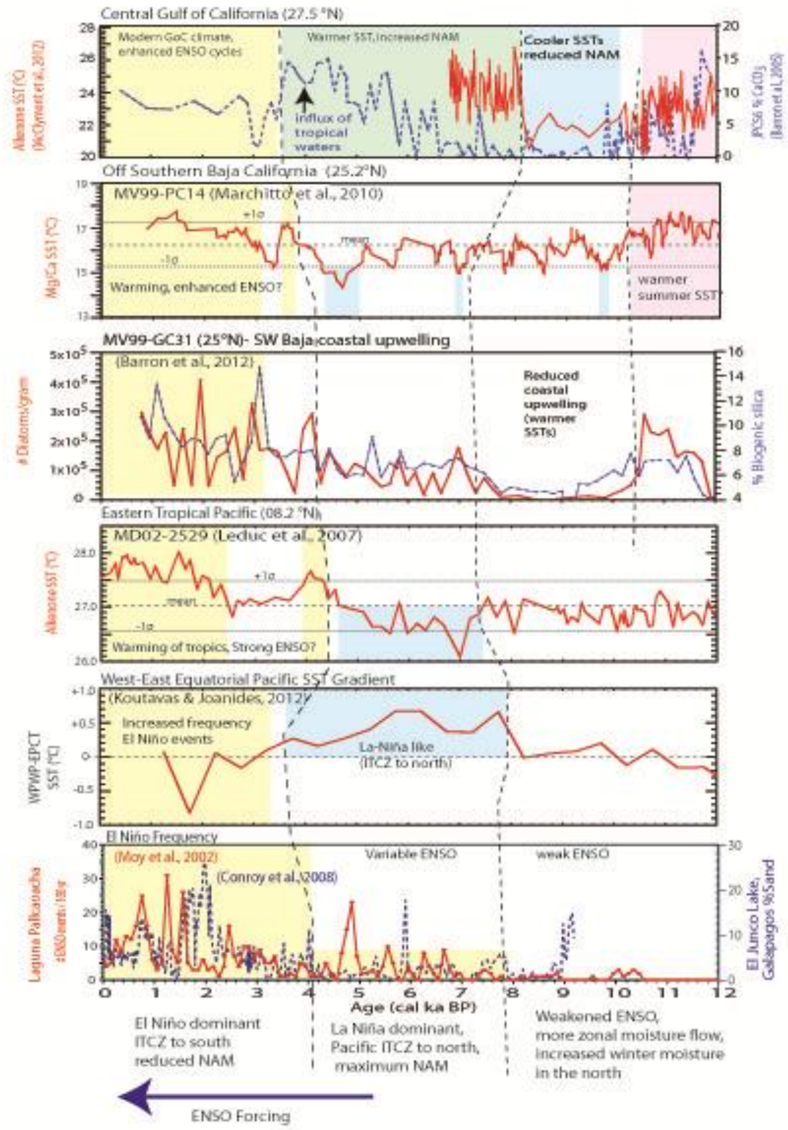

Central Gulf of California (27.5 °N)
Modern GoC climate, enhanced ENSO cycles
Warmer SST, Increased NAM
Cooler SSTs reduced NAM
influx of tropical waters
Off Southern Baja California (25.2°N)
MV99-PC14 (Marchitto et al., 2010)
Warming, enhanced ENSO?
warmer summer SST
MV99-GC31 (25°N)- SW Baja coastal upwelling
(Barron et al., 2012)
Reduced coastal upwelling (warmer SSTs)
Diatoms/gram
% Biogenic silica
Eastern Tropical Pacific (08.2 °N)
MD02-2529 (Leduc et al., 2007)
Warming of tropics, Strong ENSO?
Alkenone SST (°C)
West-East Equatorial Pacific SST Gradient
(Koutavas & Joanides, 2012)
Increased frequency El Niño events
La-Niña like (ITCZ to north)
WPWP-EPCT SST (°C)
El Niño Frequency
(Moy et al., 2002)
(Conroy et al., 2008)
Variable ENSO
weak ENSO
Laguna Pallcacocha
El Junco Lake, Galapagos %Sand
Age (cal ka BP)
El Niño dominant
ITCZ to south
reduced NAM
La Niña dominant,
Pacific ITCZ to north,
maximum NAM
Weakened ENSO,
more zonal moisture flow,
increased winter moisture
in the north
ENSO Forcing

Fig. 14

Table 1

Table 1. Continental Sites

Record Name	Map Key	Proxy Type	Latitude	Longitude	Elev (m)	Reference
Southern Region						
Harrison's Cave, Barbados	1	Speleothem	13°N	59°W	300	Mangini et al, 2007
Antoine, Grenada	2	Lake -diatoms, geochemistryistry	12°11'N	61°36'W	16	Fritz et al., 2011
Cariaco Basin, off Venezuela	3	Marine - geochemistry	10°42'N	65°10'W	-893	Haug et al., 2001; Peterson and Haug, 2006
Valencia, Venezuela	4	Lake -diatoms, isotopes	10°16'N	67°45'W	402	Bradbury et al.,1981; Curtis et al., 1999
Chilibrillo Cave, Panama	5	Speleothem	9°12'N	79°42'W	60	Lachniet et al., 2004a
La Yeguada, Panama	6	Lake -pollen	8°27'N	80°51'W	650	Bush et al., 1992
Lago de las Morrenas, Costa Rica	7	Lake -isotopes	9°29'N	83°29'W	3477	Lane et al., 2011a
Venado Cave, Costa Rica	8	Speleothem	10°6'N	84°8'W	380	Lachniet et al.2004b
El Gancho, Nicaragua	9	Lake -isotopes	11°54'N	85°55'W	44	Stansell et al.2012
Manchon Swamp, Guatemala	10	Mangrove -pollen, geochemistry	14°25'N	92°1'W	180-240	Neff et al., 2006
Yok Balum, Belize	11	Speleothem	16°12'N	89°4'W	366	Kennett et al., 2012
Peten Itza, Guatemala	12	Lake- pollen, isotopes, geochemistry, magnetic susceptibility	16°55'N	89°50'W	110	Mueller et al.,2009; Hillesheim et al., 2005: Bush et al., 2009; Escobar et al., 2012
Macal Chasm, Belize	13	Speleothem	17°N	89°W	520	Webster et al., 2007

Salpeten, Guatemala	14	Lake -isotopes, geochemistry	17°10'N	89°40'W	104	Rosenmeier et al., 2002a; 2002b
Puerto Arturo, Guatemala	15	Lake -pollen, isotopes	17°32'N	90°11'W	100-300	Wahl et al, 2006; 2014
New River Lagoon, Belize	16	Lake -diatom, pollen, isotopes	17°40'N	88°40'W	40	Metcalfe et al., 2009; Rushton et al., 2013
Nochixtlan Valley, Mexico	17	Alluvial	17°30'N	97°20'W	2000-2900	Mueller et al., 2012
Juxtlahuaca Cave, Mexico	18	Speleothem	17°04'N	99°2'W	934	Lachniet et al., 2012; 2013
Cueva del Diablo, Mexico	19	Speleothem	18°11'N	99°55'W	1030	Bernal et al., 2011
Tzib, Mexico	20	Lake-pollen, isotopes	19°17'N	88°12'W	SL	Carrillo-Bastos et al., 2010
Chichancanab, Mexico	21	Lake-isotopes, bulk density	19°52'N	88°46'W	4	Hodell et al., 1995; 2001; 2005a; Medina Elizade and Rohling, 2012
Los Petenes, Mexico	22	coastal wetland - pollen, geochemistry	20°07'N	90°27'W	< 50	Guttierez-Ayala et al., 2012
Punta Laguna, Mexico	23	Lake -isotopes	20°38'N	87°37'W	14	Curtis et al., 1996
Aguada X'Caamal, Mexico	24	Lake -isotopes	20°36'N	89°43'W	~1?	Hodell et al., 2005b
Chaac (Tzabnah Cave), Mexico	25	Speleothem	20°43'N	89°28'W	20	Medina Elizade et al., 2010
San Jose Chulchaca, Mexico	26	Lake -isotopes, diatoms	20°52'N	90°14'W	1	Leyden et al., 1996; Whitmore et al., 1996; Brenner et al, 2000; Hodell et al., 2005b
Rio Lagartos, Mexico	27	Mangrove -pollen	21°34'N	88°04'W	1	Aragon Moreno et al., 2012
Wallywash Great Pond, Jamaica	28	Lake -isotopes, geochemistry	17°57'N	77°48'W	7	Holmes et al., 1995; Street-Perrott et al., 1993
Miragoane, Haiti	29	Lake -isotopes	18°27'N	73°05'W	20	Hodell et al., 1991
Enriquillo Valley, Dominican Republic	30	Coral/lagoon - isotopes	18°N	71°W	-27	Greer and Swart, 2006
Castilla/Salvador/Felipe, Dominican Republic	31	Lake -pollen, isotopes, geochemistry	18°47'N	70°52'/53'W	976, 990	Lane et al., 2009; 2011b; 2014

Dos Anas Cave, Cuba	32	Speleothem	22°24'N	83°59'W	100	Fensterer et al., 2012
Central Region						
Lago Verde, Mexico	33	Lake -pollen, diatom	18°03'N	95°20'W	100	Lozano-Garcia et al.,2010
Lake Aljojuca	34	Lake -pollen, isotopes, geochemistry	19°05'N	97°32'W	2376	Bhattacharya et al., 2015
Agua El Marrano, Mexico	35	Depression -pollen	19°12'N	98°34'W	3860	Lozano-Garica and Vazquez-Selem, 2005
Lakes Quila/Zempoala, Mexico	36	Lake -pollen	19°04'/03'N	99°19'/18'W	3010/2800	Almeida-Lenero et al., 2005
Upper Lerma, Mexico	37	Lake -pollen, diatoms	19°N	99°30'W	2570	Lozano Garcia et al., 2005; Caballero et al., 2001; 2002; Metcalfe et al., 1991
Lago La Luna, Mexico	38	Lake -pollen, diatoms, magnetic susceptibility	19°06'N	99°45'W	4200	Cuna et L., 2013
Zacapu Basin, Mexico	39	Lake -diatoms, pollen, sediments	19°55'N	101°40'W	1970	Metcalfe, 1995; Arnauld et al.1997; Correa Metrio et al., 2012
Hoya Rincon de Parangueo, Mexico	40	Lake -pollen, magnetic susceptibility, geochemistry	20°23'N	101°15'W	1800	Park et al., 2010
Lago de Patzcuaro, Mexico	41	Lake -geochemistry, magnetic susceptibility, diatoms	19°36'N	101°39'W	2035	Bradbury, 2000; Metcalfe et al., 2007; Metcalf and Davies, 2007; Caballero et al., 2010
Lago de Zirahuen	42	Lake -pollen, magnetic susceptibility	19°26'N	101°44'W	2075	Ortega et al., 2010; Vazquez et al., 2010; Lozano-Garcia et al., 2013

Laguna de Juanacatlan	43	Lake-geochemistry	20°37'N	104°44'W	2000	Metcalfe et al., 2010
Lago de Santa Maria del Oro	44	Lake - geochemistry, magnetic susceptibility	21°22'N	104°34'W	750	Vazquez-Castro et al., 2008
Northern Region						
Las Cruces, Mexico	45	Lake -geochemistry	22°39'N	101°53'W	2106	Roy et al., 2013b
Bolson de Mapimi	46	Middens	25°54'N	103°40'W	1300	Van Devender, 1990; Felstead et al., 2014
Sierra San Francisco	47	Middens	27°32'N	113°6'W	780	Rhode, 2002
Lake Tulane, Florida	48	Lake -pollen, hydrogen isotopes	27°35'N	81°30'W	36	Grimm et al., 2006
Ciénega de Camilo, Mexico	49	Bog -pollen	28°25'N	108°34'W	1534	Ortega Rosas et al., 2008a; 2008b
Babicora, Mexico	50	Lake -diatoms, pollen, geochemistry, sediment	29°20'N	108°W	2200	Metcalfe et al.,1997; 2002; Roy et al., 2013; Ortega Ramirez et al., 1998
San Pedro Martir, Mexico	51	Middens	30°5'N	115°W	650-900	Holmgren et al., 2011
Laguna Seca San Felipe, Mexico	52	Lake -diatoms, sediment	31°08'N	115°15'W	400	Ortega Guerrero et al., 1999; Roy et al., 2010
Palomas Basin, Mexico	53	Lake -levels	31°10'N	107°25'W	1184	Castiglia and Fawcet, 2006
Cave of the Bells, Arizona	54	Speleothem	31°26'N	110°28'W	1700	Wagner, 2006
Lake Cloverdale, New Mexico	55	Lake -sediment	31°30'N	108°50'W	1573	Krider et al., 1998
Murray Springs, Arizona	56	Aroyo wall -pollen	31°34'N	110°11'W	1260	Mehringer, 1967
Eagle Eye Mountains, Arizona	57	Middens	31°53'N	113°10'W	800-825	McAuliffe and Van Devender, 1998
Pelonxillo Mountains, Arizona	58	Middens	31°58'N	108°57'W	1287-1422	Holmgren et al., 2006
Pink Panther Cave, New Mexico	59	Speleothem	32°05'N	105°10'W	1300	Asmerom et al., 2007
Bat Cave, New Mexico	60	Speleothem	32°10'N	104°27'W	1004	Asmerom et al., 2013
Fort Stanton Cave, New Mexico	61	Speleothem	33°30'N	105°12'W	1900	Asmerom et al., 2010
Lake Cochise, Arizona	62	Lake -levels	32°10'N	109°50'W	1274	Waters, 1989

Salton Basin, California	63	Tufa -isotopes	33°27'N	116°03'W	-69	Li et al., 2008
Lake Elsinore, California	64	Lake -sediment	33°40'N	117°21'W	377	Kirby et al., 2010; 2012; 2013
San Agustin Plain, New Mexico	65	Lake -pollen, ostracods	33°58'N	108°34'W	2065	Markgraf et al., 1984
Joshua Tree Nat. Monument, California	66	Middens	34°02'N	116°11'W	930-1357	Holmgren et al., 2009
Dry Lake, California	67	Lake -sediment, geochem	34°07'N	116°50'W	2763	Bird and Kirby, 2006; Bird et al., 2009
Lower Bear Lake, California	68	Lake -sediment, geochem, ostracods	34°14'N	116°58'W	2050	Kirby et al., 2012
Estancia Basi, New Mexico	69	Lake/playa -levels	34°64'N	105°97'W	1846	Menking and Anderson, 2003
Potato Lake, Arizona	70	Lake -pollen	35°03'N	111°21'W	2222	Anderson, 1993
Stewart Bog, Sangre de Christo Mnts, New Mexico	71	Bog -pollen, magnetic susceptibility	35°40'N	105°45'W	3100	Jiménez Moreno et al., 2008
Mojave Desert, California	72	Alluvium -sediment	35°N	117°W	~300	Miller et al.2010
Silver Lake (Lake Mojave), California	73	Lake -level, sediment	35°12'N	116°08'W	276	Enzel et al., 1992; Wells et al., 2003; Kirby et a 2012
Owens Lake, California	74	Lake -level, sediment	36°13'N	117°37'W	1084	Benson et al.,2002; Bacon et al.,2006
Bear Lake, Arizona	75	Lake -pollen	36°22'N	112°09'W	2778	Weng and Jackson, 1999
Fracas Lake, Arizona	76	Lake -pollen	36°38'N	112°14'W	2518	Weng and Jackson, 1999
Mono Lake, California	77	Lake -level, tree rings	38°01'N	119°0'W	1955	Stine, 1994
Pyramid Lake, Nevada	78	Lake -isotope, magnetic susceptibility	40°03'N	119°34'W	1157	Benson et al., 2002