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Isotopic and hydrologic responses of small, closed lakes to climate variability: Hydroclimate reconstructions from lake sediment oxygen isotope records and mass balance models

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Abstract

Hydroclimate sensitivity simulations were conducted with a lake-catchment hydrologic and isotope mass balance model adapted to two small, closed lakes (Castor and Scanlon) located in the Pacific Northwest. Model simulations were designed to investigate the combined influences of persistent disequilibrium, reddening, and equifinality on lake water and sediment (i.e., biogenic and endogenic carbonate mineral) oxygen isotope ($\delta^{18}O$) values and to provide a basis for quantitative, probabilistic climate reconstructions using lake sediment $\delta^{18}O$ records. Simulation results indicate that within closed-basin lakes changes in long-term (i.e., multi-decadal) precipitation amounts produce inconsistent responses in lake water and sediment $\delta^{18}O$ values that are strongly influenced by lake basin outseepage and morphometry. Simulations of variable initial conditions in which randomly generated monthly climate data (i.e., precipitation, temperature, and relative humidity) were used to force the model during the equilibration period (which precedes the application of instrumental climate data) demonstrate that Castor Lake and Scanlon Lake have a somewhat limited isotopic 'memory' of ~10 years. Additional tests conducted using a Monte Carlo ensemble (in which random climate data were used to force the model) combined with $\delta^{18}O$ analyses of water samples collected from 2003 to 2011 AD, indicate that within small, closed lakes in the Pacific Northwest November–February precipitation is the strongest seasonal, climatic control on sediment oxygen-isotope values. Further, a Monte Carlo based reconstruction of 20 year average November–February precipitation amounts strongly correlates ($R^2 = 0.66$) to instrumental values from the 20th century (with all observed values falling within modeled 95% prediction limits), indicating that probabilistic, quantitative paleoclimate interpretations of lake sediment $\delta^{18}O$ records are attainable.

1. INTRODUCTION

The oxygen ($\delta^{18}O$) and hydrogen (δD) isotopic composition of lake water is influenced by multiple factors including lake morphology, catchment size and hydrologic characteristics, groundwater throughflow rates, and climate variations (Gonfiantini, 1986; Almendinger, 1990; Smith et al., 1997; Leng and Marshall, 2004; Steinman et al., 2010a). Despite the multitude of controls on lake hydrology, isotopic processes within lakes are well understood due to a large number of modeling and observational studies (e.g., Gat, 1970; Hostetler and Benson, 1994; Donovan et al., 2002) in which the effects of climate, catchment and groundwater controls have been described and reproduced through model experiments. This abundance of research has formed a basis for qualitative and semi-quantitative paleoclimate interpretations of lake sediment (i.e., biogenic or endogenic carbonate) $\delta^{18}O$ and δD records (Sachs et al., 2009; Bird et al., 2011; Nelson et al., 2011) as well as model based quantitative reconstructions of paleoprecipitation and humidity (Ricketts and Johnson, 1996; Cross et al., 2001; Anderson et al., 2007; Jones et al., 2007). To date however quantitative paleoclimate reconstructions have not been produced in which modern sediment core isotope

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records were interpreted using lake models to reproduce instrumental (i.e., measured) hydroclimatic variability for the purposes of model validation/verification (i.e., similar to exercises conducted in tree-ring based paleoclimate reconstructions). Obstacles to such efforts have included the complexity inherent in the combined influence of mean state (i.e., multidecadal to century scale) and stochastic (i.e., random, monthly to inter-annual) climate variations on lake hydrologic and isotopic evolution (Benson and Paillet, 2002; Leng et al., 2005; Steinman et al., 2010b) and the difficulty in accounting for these controls in lake models. Other problems have included a lack of instrumental climate data (Jones et al., 2005), insufficient observation or measurement of lake hydrologic and isotopic processes (e.g., outseepage rates), and the confounding effects of catchment alteration on modern lake hydrology. Overcoming these obstacles could potentially allow lake sediment isotope records to be quantitatively interpreted to produce statistically validated paleoclimate reconstructions.

Lake surface water $\delta^{18}O$ values in seasonal climates are primarily controlled by the balance between the inflow of fresh meteoric water (e.g., via the subsurface and catchment runoff) during the cooler seasons (when evapotranspiration rates are low and catchment water balance is positive), water losses through evaporation during the warmer seasons, and outflow through non-fractionating pathways (i.e., overflow and outseepage through the lake bed). The isotopic composition of inflowing water is an additional control that is influenced by various factors including temperature at the time of precipitation (e.g., in the mid-latitudes summer rainfall $\delta^{18}O$ values are high and winter values are low) and the origin and rainout history of the air mass (Dansgaard, 1964; Rozanski et al., 1992, 1993). Hydrologically closed lakes lose the majority of water through evaporation and are more isotopically enriched relative to open lakes, which lose the majority of water through non-fractionating outflows. The limiting cases are a closed lake that loses all water through evaporation, and therefore is at the limit of isotopic enrichment (Gat, 1984), and an open lake that loses no (or very little) water through evaporation, and maintains an isotopic composition similar (or identical) to that of meteoric water (von Grafenstein et al., 1996; Hammarlund et al., 2002; Anderson et al., 2005).

In seasonal climates, closed lakes are in a state of persistent intra- to inter-annual (i.e., transient) hydrologic and isotopic disequilibrium (Gibson et al., 2002; Shapley et al., 2008; Jones and Imbers, 2010) resulting from disparity in seasonal water inflow and losses through evaporation. This shorter-term disequilibrium, however, exists within a longer-term equilibrium state in which short timescale (e.g., seasonally derived) isotopic variations produce average $\delta^{18}O$ values over many years that are primarily a function of mean state climate conditions (Steinman et al., 2010b). Given that smaller lakes typically have shorter residence times, seasonality can exert a strong influence on small lake water isotopic evolution, leading to large intra-annual variance in surface water $\delta^{18}O$ values. Because of these circumstances, small lakes respond more quickly to decadal to sub-decadal climate variability, and therefore have the potential to produce sediment paleoclimate archives that contain information on a wide range of timescales.

Although small lakes exhibit greater seasonal sensitivity, they have several characteristics that make them relatively easier to model and therefore understand in the context of paleo-interpretations of lake sediment isotope records. Small lakes, for example, generally have simple catchments that span a narrow elevation range and that contain relatively few soil and vegetation types, factors that are difficult to account for when determining hydrologic characteristics such as soil available water capacity. Additionally, small lakes are typically overlain by advected air masses that contain minimal moisture derived from lake surface evaporation, an important distinction given that measuring the fraction of advected air over a large lake is difficult and adds complexity to calculations of the isotopic composition of evaporation (Benson and White, 1994). Accounting for the fundamental differences between large and small lakes is important in model design and potentially precludes the use of large lake models on smaller lakes, and vice versa, in paleoclimate applications.

The relative hydrologic and isotopic simplicity of small lakes notwithstanding, there are several inherent aspects of all lake systems that can potentially complicate paleoclimate interpretations of sediment geochemical variations: namely, the persistent state of disequilibrium (in both lake hydrology and isotopic processes), reddening (i.e., red noise in water and sediment geochemical values), and equifinality. Persistent disequilibrium occurs to a greater degree in small lakes and can lead to extensive water geochemical variance over seasonal to inter-annual timescales. This may cause difficulties when interpreting, for example, isotope records from lakes in which the timing of carbonate mineral formation changes substantially from year to year or when mixing (e.g., through bioturbation) changes the temporal resolution of sediment. Reddening is a condition in which lake responses to climatic perturbation depend on the hydrologic and isotopic state of the lake at the time of the climatic change. Reddening has been investigated through modeling studies (Benson et al., 2002; Steinman et al., 2010a) and is a generally well understood complicating factor for the interpretation of lake sediment geochemical records, especially in the case of large lakes with long residence times. Equifinality results in part from reddening and persistent disequilibrium and is defined as a state in which many possible climate (and resulting lake hydrologic and isotopic) scenarios can produce a specific lake water or sediment geochemical (i.e., $\delta^{18}O$ or salinity) value. Equifinality occurs to a greater extent in lakes that have a similar sensitivity to two or more hydroclimatic variables. One example would be a lake system in a monsoonal setting in which the majority of precipitation falls during the warmer season, a scenario that could lead to a high level of both temperature and precipitation sensitivity (Henderson and Shuman, 2009) and could lead to an inability to distinguish in the sediment record between past wetter or cooler (and hotter or drier) conditions. All three of these characteristics (disequilibrium, reddening, and equifinality) can produce "noise" in lake sediment geochemical records that, depend-

ing on the attributes of the lake system, has the potential to obscure the climate signal in paleo-interpretation studies. One way to account for these complicating factors is the application of a Monte Carlo ensemble approach to lake-catchment modeling in which random climate data are applied over the course of a large number of simulations to develop a range of potential climatic scenarios that can explain each measured sediment $\delta^{18}O$ value. Such an approach would fundamentally integrate persistent disequilibrium, reddening and equifinality in the model based solutions and would provide a probability based method for producing quantitative paleoclimate information using lake sediment isotope records.

This manuscript is the second (along with Steinman and Abbott, 2012) in a two part series designed to show how lake hydrologic and isotope mass balance modeling can be combined with sediment core isotope data from small, closed lakes to produce quantitative paleoclimate reconstructions. In the first paper we conducted validation exercises to demonstrate that model reconstructions of lake hydrologic and isotopic responses to climate forcing correspond with observations (or measurements). In this manuscript we develop new methods for interpreting sediment core oxygen isotope records using Monte Carlo based model simulations (in which climate data are randomly varied within realistic limits) to produce probabilistic, quantitative solutions of past hydroclimatic conditions. The overall objective of this two part series is to provide guidelines for producing quantitative interpretations of lake sediment paleorecords, and to show how characterizing uncertainty (e.g., in physical parameters of the lake-catchment system or in climate datasets) is essential for constraining error in model predictions of lake geochemical dynamics. Markedly, the analytical methods described herein could potentially be used to produce probabilistic, quantitative estimates of past hydroclimatic conditions in drought prone regions such as the Pacific Northwest where water management strategies are based on instrumental datasets spanning a relative short time period.

Here we use the lake-catchment model of Steinman et al. (2010a) to simulate the hydrologic and isotopic responses of two lakes in central Washington, Castor Lake and Scanlon Lake, to hypothetical climate scenarios derived from instrumental weather data. In the first set of simulations, average relative humidity (RH), temperature, and precipitation amount values from the time period of instrumental observation (i.e., 1900 to 2007 AD) were altered to demonstrate lake responses to combined mean-state and stochastic climate change. Specifically, average climate values were changed by a constant amount (i.e., monthly precipitation, temperature, and RH were changed by ±50%, ±2 °C, and ±5%, respectively) to assess the degree of control of individual climate variables on lake water and sediment $\delta^{18}O$ values. We use $\delta^{18}O$ measurements of water samples collected from 2003 to 2011 AD to provide support for these test results and to identify the climate variables most effectively reconstructed using a Monte Carlo algorithm. In the second set of tests, standard deviations of annual average instrumental temperature and precipitation were altered by a fixed percentage (i.e., the standard deviations were changed by ±50% while maintaining the annual average values and monthly trends) to demonstrate the effect of variance changes in hydroclimate. In the third series of simulations, the influence of initial hydrologic and isotopic conditions (which produce reddening) on model reconstructions of lake sediment $\delta^{18}O$ values were investigated by applying randomly generated monthly climate data during the equilibration period that preceded the application of the instrumental datasets. Results from these experiments provide information on the expected duration of the influence of past climate changes on lake hydrologic and isotopic states. In the final series of simulations, a Monte Carlo approach (in which random, monthly precipitation, temperature, and relative humidity values were applied as model inputs) was used to reconstruct November–February precipitation amounts at CL over the 20th century and to demonstrate the potential of this method for producing probabilistic, quantitative paleoclimate records.

2. METHODS

2.1. Study sites

Scanlon Lake (SL) (48.542 N, 119.582 W) and Castor Lake (CL) (48.539 N, 119.561 W) are located in north-central Washington, within 14 km and ~300 m elevation of two National Climatic Data Center (NCDC) and one Pacific Northwest Cooperative Agricultural Weather Network (AgriMet) weather stations (Fig. 1). The seasonal, semi-arid climate in this region is largely controlled by the interaction between the Pacific westerlies and the Aleutian low- and north Pacific high-pressure systems. A more detailed description of the climate setting is provided by Steinman and Abbott (2012). Monthly average climate data for the CL/SL region appears in Table EA1.

The observational and modeling studies of Steinman et al. (2010a) identified substantial hydrologic and morphological differences between SL and CL that affect lake isotopic and hydrologic responses to climatic forcing. For example, at SL, lake bed outseepage rates (estimated on the basis of mass balance modeling) are much lower than at CL, leading to a greater proportion of water loss through evaporation. The smaller lake water volume at SL relative to CL (despite similar surface areas) (Steinman et al., 2010a) also leads to a larger volumetric percentage change (and hence enhanced sensitivity) in response to seasonal and inter-annual hydroclimatic forcing. At CL, surface overflow occurs when lake level reaches ~13.5 m, providing an additional non-fractionating outflow pathway that produces exceptionally different hydrologic and isotopic responses to mean state precipitation increases at CL relative to SL (Steinman and Abbott, 2012).

2.2. Water isotope samples

Water samples for hydrogen and oxygen isotope analyses were collected from lake surfaces and from shallow wells within the lake catchments at irregular intervals between 2003 and 2011 (see Fig. 1 of Steinman and Abbott, 2012

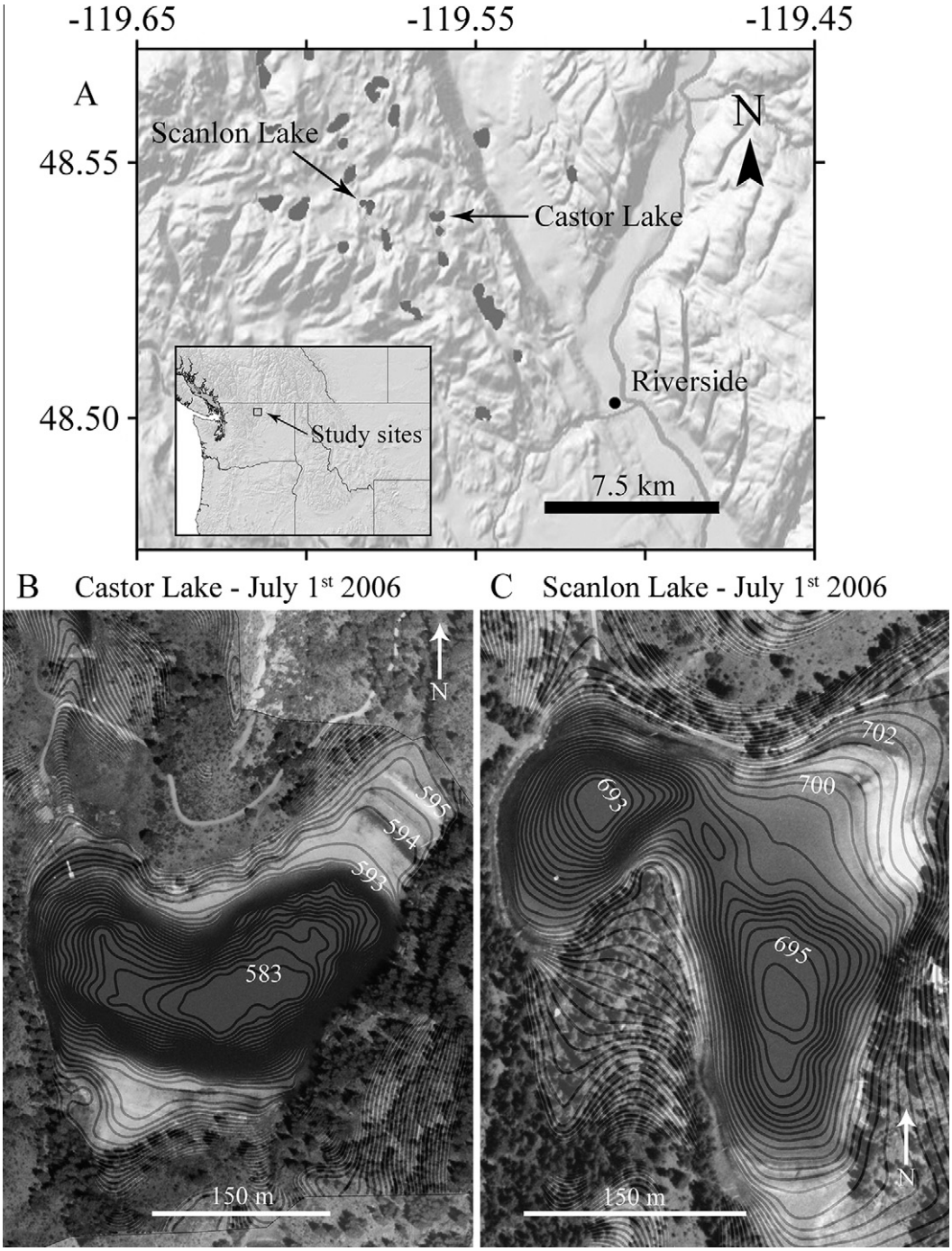

Fig. 1. Regional basemap (A) showing the location of Castor Lake and Scanlon Lake in north-central Washington, USA. Catchment topography and land cover map of Castor Lake (B) and Scanlon Lake (C) adapted from aerial photographs taken on July 1st, 2006 (Steinman and Abbott, 2012). Catchment maps are in UTM coordinates expressed in meters. 0.5 m elevation contours (above mean sea level) are displayed.

for well locations). Samples were collected in 30 mL polyethylene bottles by rinsing three times with sample water and then filling and capping the bottle underwater to remove any trapped air. After collection, all samples were immediately stored in a cooler and kept refrigerated until analysis. Isotopic ratios of lake water oxygen were measured at the University of Arizona Environmental Isotope Laboratory by CO_2 equilibration with a Finnigan Delta S isotope ratio mass spectrometer. For hydrogen, samples were reacted at 750 °C with Cr metal using a Finnigan H/Device coupled to the mass spectrometer. The reported precision is better than 0.1‰ for $\delta^{18}O$ and 1.0 ‰ for δD. Water sample isotope results are available in Electronic Annex EA-3 of Steinman and Abbott (2012).

Table 1
Climate-lake sediment $\delta^{18}O$ regression data from Monte Carlo simulations.

Time interval (year)	Months	Precipitation			Temperature		RH	
		Slope	y-Intercept	R^2	Slope	R^2	Slope	R^2
20	November–February	−42.25	−54.74	0.49	0.90	0.35	−1.16	0.20
20	March–June	−15.49	43.04	0.07	0.69	0.29	−1.03	0.19
20	July–October	−14.18	1.96	0.09	0.61	0.23	−1.13	0.19

2.3. Model structure

In this study, modeled lake hydrology and isotope mass balance dynamics for the CL and SL catchments were determined using a system of 12 ordinary differential equations compiled using STELLA (isee systems) software. Additional detail regarding the model is provided by Steinman et al. (2010a) and Steinman and Abbott (2012). All model simulations were conducted using modified versions of the Castor/Scanlon models available as Electronic Annex EA-2 in Steinman and Abbott (2012). Unless otherwise stated, May–July average values of model simulation results are presented.

2.4. Mean state RH, temperature, and precipitation forcing simulations

To explore the influence of mean state hydroclimatic changes on lake hydrology and isotope dynamics, we conducted simulations using continuous temperature, precipitation, and RH datasets (Fig. 2, Steinman and Abbott, 2012) in which the monthly values of each climate variable were changed by a fixed amount. Simulations in which monthly climate data were continuously offset during the test were conducted for precipitation (±50%), temperature (±2 °C) and RH (±5%). Monthly precipitation amounts were increased by 50% in one set of model tests, and decreased by 50% in a second set of simulations. Similarly, monthly average temperature was both increased and decreased by 2 °C, in two independent models tests. Relative humidity was also changed (increased or decreased by 5%) in separate model runs. Note that relative humidity in the model is defined as a function of temperature for all years prior to 1989 (Steinman and Abbott, 2012) and therefore that temperature sensitivity tests also include changes in relative humidity (of ~1.5%/°C). These model tests were conducted using the Castor/Scanlon model without uncertainty algorithms (i.e., uncertainty in climate data and the timing of carbonate mineral formation were not represented in the model) (Steinman and Abbott, 2012).

2.5. Temperature and precipitation variance forcing simulations

To investigate shorter-term climate forcing effects, we conducted experiments that utilized alterations to the inter-annual standard deviation of precipitation and temperature while maintaining their respective annual averages (Fig. 2). For each model year, average annual temperature and annual total precipitation amounts were calculated using instrumental datasets (Steinman and Abbott, 2012). These values were then altered so that the dataset average

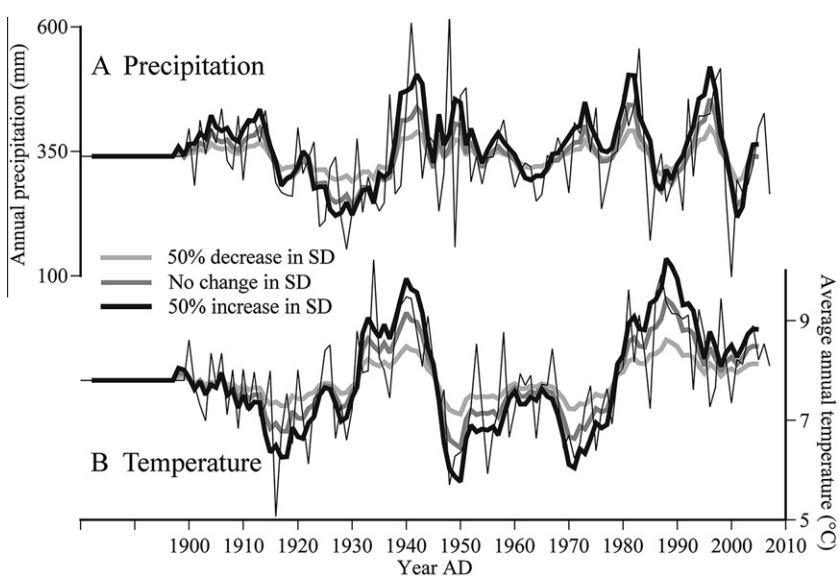

Fig. 2. Continuous annual (A) total precipitation and (B) average temperature instrumental datasets unaltered and modified by changing the standard deviation by ±50%. Fine lines depict annual values. Coarse lines depict 5 year moving averages. Climate dataset description is provided by Steinman and Abbott (2012).

remained unchanged while the inter-annual standard deviation was either increased or decreased by 50%, according to the following equations:

$$\left(\frac{\text{Annual total } P \text{ in year } x}{\text{20th century average annual } P} - 1\right) \times (1.5 \text{ or } 0.5) + 1 = APM \quad (1)$$

$$\left(\frac{\text{Annual averagre } T \text{ in year } x}{\text{20th century average annual } T} - 1\right) \times (1.5 \text{ or } 0.5) + 1 = ATM \quad (2)$$

where P is precipitation, T is temperature, APM is the annual precipitation modifier, and ATM is the annual temperature modifier. To produce the final altered dataset, the annual modifier for each year was multiplied by the instrumental monthly values for that year. Note that negative temperatures were offset by a positive amount if the $ATM > 1$ and by a negative amount if $ATM < 1$. These model tests were conducted using the Castor/Scanlon model (without uncertainty algorithms) (Steinman and Abbott, 2012).

2.6. Initial condition forcing simulations

To investigate the influence of reddening on the simulation results, we conducted experiments in which randomly generated monthly precipitation, temperature, and RH data were applied during the equilibration period, after which instrumental climate data were applied. In these tests, modeled $\delta^{18}O$ values and lake volumes from each individual simulation approached equality after the application of instrumental data. We defined this threshold of near equality as the convergence point. To establish a quantitative basis for this inference, we calculated the standard deviation of sediment $\delta^{18}O$ values and volumes for each year of the 50 simulation ensemble and established 2σ limits of 0.25‰ for lake sediment $\delta^{18}O$ values and 2.5% of total average volume (defined as the average volume over the course of all 50 simulations) for volumetric estimates. These model tests were conducted using an altered Castor/Scanlon model (available in Electronic Annex EA-2) in which uncertainty algorithms (described above) were removed and Monte Carlo components were added to produce random climate data over the first 20 years of the model test prior to the application of instrumental data.

2.7. Monte carlo simulations

We selected the CL model and sediment core $\delta^{18}O$ record to conduct Monte Carlo simulations for application to the 20th century precipitation reconstruction because of the considerable sensitivity of CL to precipitation changes on a variety of timescales (from intra-annual to multidecadal) (Steinman et al., 2010a). CL sediment core recovery, sampling methods, and dating are described by Nelson et al. (2011) and Steinman and Abbott (2012). Uncertainty in the timing of aragonite mineral formation (which occurs during the spring and summer months, primarily May–July) in the CL water column is also addressed by Steinman and Abbott (2012). To simulate lake hydrologic and isotopic responses to random climate variability, we forced the CL model using white noise (with both even and normal distributions) applied to monthly climate variables, namely precipitation, temperature, and relative humidity (see Figs. EA1–EA4 in Electronic Annex EA-1). Prior to each simulation, randomly selected average values for each climate variable for each calendar month were determined (within the model) using an even distribution random number algorithm. Climate variable input limits (i.e., the domain of the Monte Carlo simulations) were based on the largest 20 year average instrumental data variations from the mean. January precipitation, for example, averaged 36 mm over the instrumental period with maximum and minimum 20 year average values of 47 and 27 mm, respectively (which differed by 11 and 9 mm from the mean) (see Table EA1 in Electronic Annex EA-1). The model therefore selected an average precipitation amount of 36 ± 11 mm for January at the beginning of each test that established the mean values for this month in all following simulation years. The model then imparted stochastic variance to the monthly mean values using a randomly generated, lognormal distribution (based on the inter-annual standard deviation for each month over the instrumental period) to produce random, inter-annual variance in precipitation. Similarly, the model generated average monthly temperatures within a range defined by variations in 20 year instrumental data averages and then imparted stochastic variance to these mean monthly values over the course of the simulation. RH values were defined as a function of temperature and then randomly varied (using an even distribution) within the 95% prediction intervals of the RH–temperature regression relationship (Steinman and Abbott, 2012). These model tests were conducted using an altered Castor/Scanlon model (available in Electronic Annex EA-2) in which all climate data was produced by the Monte Carlo algorithm described above EA.

One hundred simulations were conducted on a monthly time step over 60 years of which the first 20 were a model equilibration period. The relationship between modeled lake sediment $\delta^{18}O$ values and average climate values within the remaining model simulation years were then split into discrete 20 year periods to investigate the influence of variable (or unknown) initial conditions on longer-term relationships between hydroclimate and lake sediment and water $\delta^{18}O$ values. For example, for each individual dataset in the ensemble, the final 20 annual sediment $\delta^{18}O$ values were averaged to produce a 20 year average $\delta^{18}O$ value, which was then related to climate variable averages over the common time period. Specifically, these 20 year average modeled sediment $\delta^{18}O$ values were compared to November–February, March–June, and July–October average precipitation, temperature and RH. We used linear regression to analyze the relationship between the simulated climate data averages and sediment $\delta^{18}O$ values. The linear regression equation for November–February precipitation (which had the highest correlation coefficient) was then applied to the CL sediment core $\delta^{18}O$ values to reproduce corresponding 20 year average 20th century precipitation amounts. Prediction limits (at the 95% level) were calculated for the

precipitation-sediment $\delta^{18}O$ relationships in order to constrain uncertainty. The model and sediment core derived precipitation amounts were then compared to instrumental data to assess veracity of the reconstruction.

3. RESULTS AND DISCUSSION

3.1. Water isotope measurements

CL and SL water $\delta^{18}O$ values exhibited considerable seasonal variability, ranging from −0.4‰ to −8.1‰ for CL and −13.7‰ to 2.7‰ for SL (Fig. 3). SL waters were in almost all cases more isotopically enriched than CL waters collected on the same or similar dates indicating that SL has a higher degree of hydrologic closure than CL. Isotopic values of waters collected from wells located in the lake catchments plot at the intersection point of the local evaporation line (LEL) and the local meteoric water line (LMWL), reinforcing the assertion that both lakes experience significant evaporative losses (see Fig. 6 in Steinman and Abbott, 2012). Qualitative comparison of lake water $\delta^{18}O$ values (from the months of July and August) to seasonal temperature and precipitation amounts (beginning in November of the year prior to the sample collection date) indicates that November–February precipitation is the most influential seasonal climatic control on both CL and SL $\delta^{18}O$ values on interannual timescales and that March–June precipitation is also important (Fig. 4). Interestingly, there does not appear to be a strong coherence between summer water $\delta^{18}O$ values and temperature in any season at either lake. Both temperature and well water $\delta^{18}O$ values generally decrease over the period of observation, an expected result given the established positive correlation between precipitation $\delta^{18}O$ values and temperature in the mid latitudes (Dansgaard, 1964; Rozanski et al., 1992, 1993). Markedly, the $\delta^{18}O$ values of well water samples do not match the precipitation weighted mean isotopic composition of precipitation (−13.2‰, calculated using data in Table EA1) but are similar to interpolated December–March values, which vary between −15‰ and −16‰ (Fig. 5) (waterisotopes.org; Bowen and Wilkinson, 2002; Bowen and Revenaugh, 2003; Bowen et al., 2005). This result contrasts with other studies (Henderson and Shuman, 2009, 2010) which demonstrate that in most areas of the American West, lake input waters have an isotopic composition that reflects the weighted mean isotopic composition of precipitation rather than the water surplus weighted mean $\delta^{18}O$ value.

3.2. Simulations of mean state variations in RH, temperature, and precipitation

Lake hydrologic responses to mean state changes in RH, temperature, and precipitation were generally consistent between CL (Fig. 6) and SL (Fig. 7), in that precipitation rates more strongly influenced average lake level and volume than did RH and temperature. For example, the average change in lake depth (at the deepest part of the lake) in the precipitation decrease scenario was much greater for CL and SL (with lake level declines of 737 and 431 cm, respectively) than lake level changes in the RH decrease and temperature increase tests (see Table EA2 in Electronic Annex EA-1). RH and temperature adjustments did however, produce less appreciable average lake level changes at CL relative to SL due to the lower outseepage rate, greater SA:V ratio, and the fact that evaporation represents a greater proportion of the total outflow at SL relative to CL. These results suggest that RH and temperature control of long-term lake level and volume is in part a function of the amount of water loss through non-fractionating pathways such as overflow and throughflow, which are strongly influenced by lake morphology and lake bed outseepage (and overflow) rates.

The effects of mean state precipitation change on average sediment $\delta^{18}O$ values were inconsistent between CL (Fig. 6) and SL (Fig. 7) and depended upon outseepage

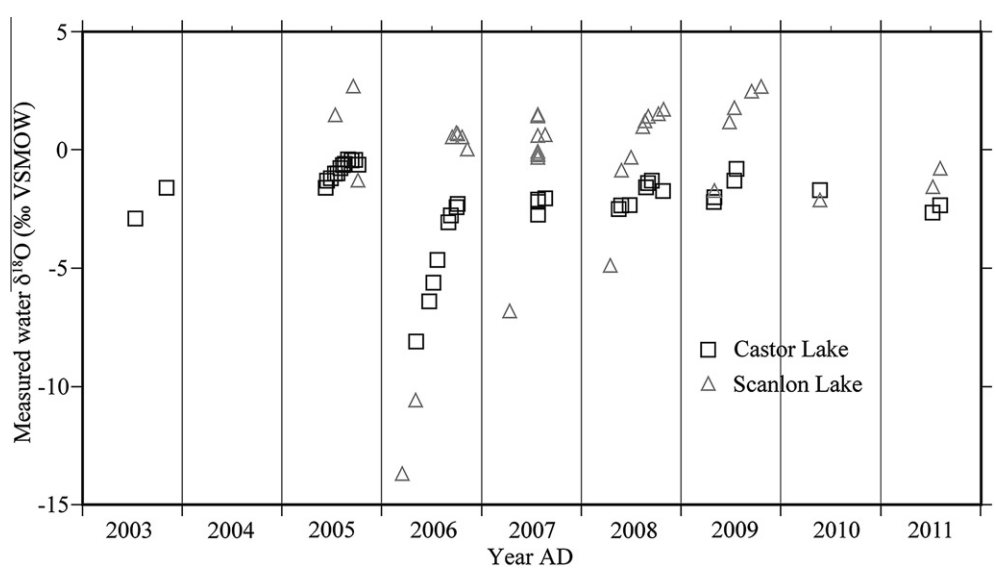

Fig. 3. $\delta^{18}O$ values of Castor Lake (open squares) and Scanlon Lake (open triangles) water samples collected from 2003 to 2011 AD.

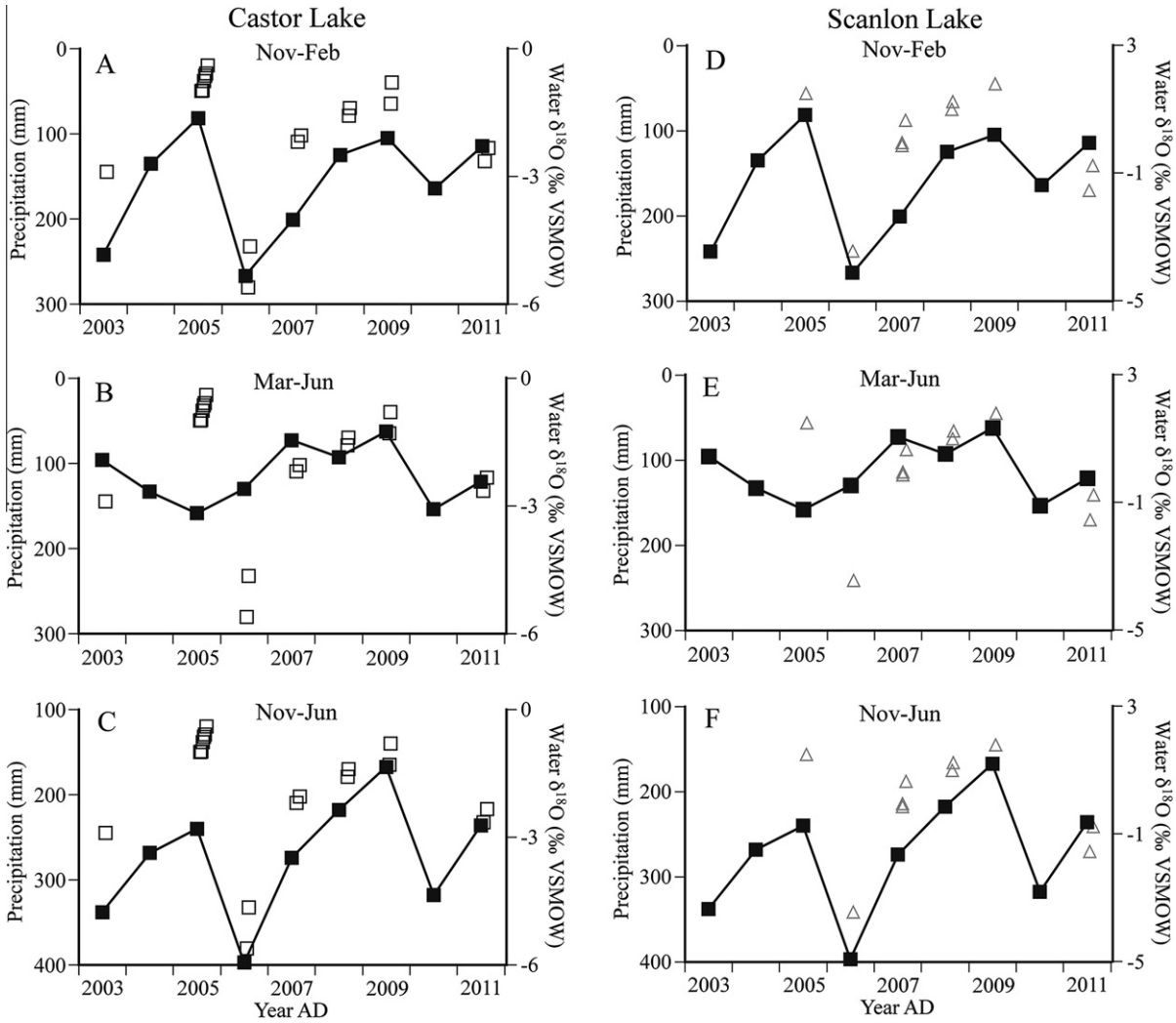

Fig. 4. Castor Lake (A–C) and Scanlon Lake (D–F) water sample $\delta^{18}O$ value (open squares) comparison with seasonal (November–February, March–June, and November–June) total precipitation amounts (closed squares) (Steinman and Abbott, 2012) from the period of observation (2003–2011 AD). Only July–August water sample data are shown. Note that lake $\delta^{18}O$ values correlate most strongly to November–February precipitation totals.

dynamics and basin morphology. For CL, a mean precipitation increase of 50% resulted in lake level increases, surface overflow (a non-fractionating outflow pathway), and subsequently lower lake water and sediment $\delta^{18}O$ values (with decreases of ~0.9 ‰ for both water and sediment May–July average values) (Table EA2). Precipitation decreases of 50%, in contrast, produced lower lake levels, greater proportional water loss through evaporation (a fractionating outflow pathway) and consequently higher $\delta^{18}O$ values (with increases of ~2‰ for both water and sediment May–July average values). For SL however, large increases in precipitation produced higher lake levels and an increase in the surface area to volume (SA:V) ratio (due to basin morphology), which in turn lead to greater water loss through evaporation and an increase in lake water and sediment $\delta^{18}O$ values (by ~0.6‰). Large decreases in precipitation at SL produced lower lake levels and a similar increase in the SA:V ratio, which (as in the precipitation increase scenario) lead to greater proportional evaporative outflow and higher $\delta^{18}O$ values (by ~0.4‰). The more pronounced $\delta^{18}O$ changes at CL relative to SL in the precipitation decrease scenario largely resulted from the different outseepage rates of the two lakes. At CL, where the outseepage rate is higher, basin morphology leads to an increased SA:V ratio with a decrease in lake level, and a corresponding increase in the proportion of water lost through evaporation. In contrast, at SL, where the outseepage rate is lower, a decrease in lake level does not lead to a large change in the proportion of water lost through evaporation because low outseepage rates produce a lower sensitivity (relative to CL) to SA:V ratio changes. These results demonstrate that closed-lake outseepage (and overflow) rates are a fundamental control of long-term lake isotopic responses to hydrologic forcing (i.e., lower outseepage and overflow rates cause decreased lake isotopic sensitivity to mean state hydrologic forcing and vice versa), a finding that

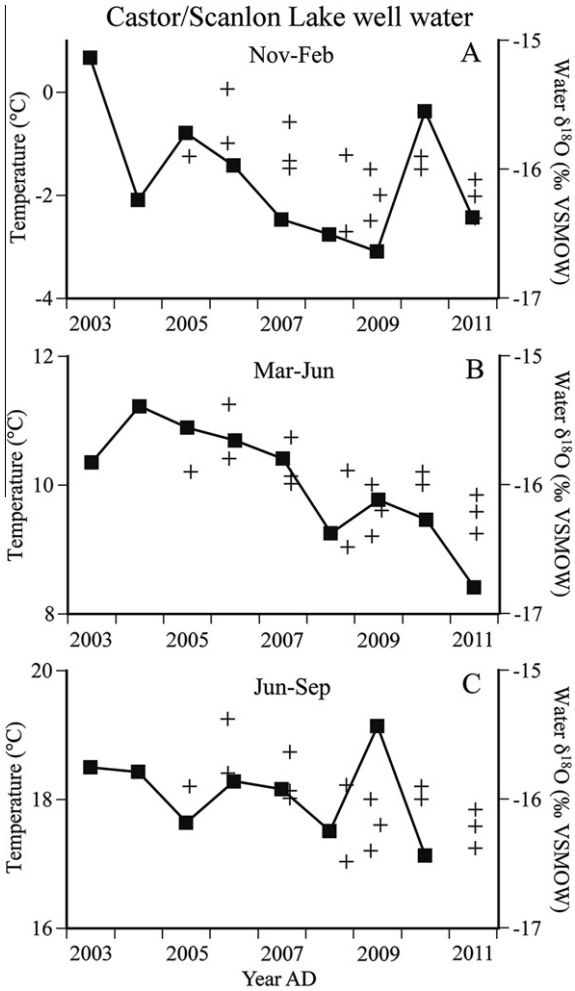

Fig. 5. Castor Lake and Scanlon Lake well water sample $\delta^{18}O$ value (plus signs) comparison with seasonal temperature data (closed squares) (Steinman and Abbott, 2012), Electronic Annex EA-3 from the period of observation (2003–2011 AD). November–February, March–June, and June–September average temperature (A–C, respectively) are displayed.

has been reproduced using other models (Gat and Bowser, 1991; Gat, 1995; Gibson, 2001).

At both CL (Fig. 6) and SL (Fig. 7), mean temperature changes of ±2 °C resulted in significantly altered water $\delta^{18}O$ values but relatively smaller changes to sediment $\delta^{18}O$. At CL and SL for example mean temperature forcing of ±2 °C produced average lake water $\delta^{18}O$ values that were different by $\sim \pm 1.7‰$ (with positive values corresponding to the temperature increase scenario, and vice versa) and average sediment $\delta^{18}O$ changes of only $\sim \pm 1.3‰$. This discrepancy exists because of distinct seasonal isotopic influences that have the potential to offset one another. More specifically, in the winter and spring months, higher than normal temperatures lead to an increase in the isotopic composition of rainfall and or snowpack (by ~0.6‰/°C) (Dansgaard, 1964; Rozanski et al., 1992, 1993) which subsequently flows into the lake after spring thaw, but only if catchment soils are saturated. In model simulations of increased temperature, a greater amount of isotopically heavier spring and winter precipitation is removed from catchment soils due to greater water loss via evaporation and increases in transpiration during spring, partially canceling the effects of isotopically heavier precipitation on lake water isotopic mass balance. This catchment modeling result is supported by regional scale observational (Dettinger and Cayan, 1995; Regonda et al., 2005; Stewart et al., 2005) and modeling (Lettenmaier and Gan, 1990; Dettinger et al., 2004; Hamlet et al., 2007) studies that demonstrate earlier spring runoff associated with higher temperatures. The primary summer temperature influences on sediment isotope values result from evaporative concentration of oxygen-18 in lake water (with higher temperatures resulting in higher evaporation rates) (Gat, 1981; Gibson et al., 1996; Leng et al., 2005) and control of the equilibrium fractionation factor for the calcite/aragonite–water system, in which higher water temperatures lead to a decrease in the fractionation factor and lower carbonate mineral $\delta^{18}O$ values (Friedman and O'Neil, 1977; Kim and O'Neil, 1997; Kim et al., 2007). These two summer controls therefore partially offset one another such that summer temperature increases produce appreciably higher lake water $\delta^{18}O$ values but smaller positive adjustments in sediment $\delta^{18}O$. The slightly larger water and sediment $\delta^{18}O$ changes at SL in response to temperature offsets (Table EA2) highlights the greater control of temperature on isotope dynamics in closed lakes with low outseepage rates.

An additional, notable result of the mean state change experiments is that the variance of sediment $\delta^{18}O$ values was dependent in part on lake level and volume, with periods of lower lake level resulting in an increased variance in sediment $\delta^{18}O$ values and vice versa (Figs. 6 and 7). Results from these tests therefore suggest that in volumetrically smaller closed lakes (which have larger SA:V ratios than larger lakes with similar proportions) greater variance in water and sediment isotope values can be expected in response to mean state changes in hydroclimate. This assertion is supported by the larger standard deviation of Scanlon Lake water $\delta^{18}O$ values relative to that CL over the observational period (Figs. 3 and 4). This supports the conclusion of Steinman et al. (2010b) that percentage changes in lake volume will be greater at lower lake levels and in smaller lakes, resulting in comparably larger isotopic variance in response to stochastic precipitation (and vice versa).

3.3. Simulations of changes in RH, temperature, and precipitation variance

For both CL and SL, the variance of modeled, inter-annual sediment isotope values was largely proportional to the variance of annual precipitation, an expected result given the control that precipitation exerts over lake water and sediment $\delta^{18}O$ values on inter-annual to decadal timescales (Fig. 8). Conversely, simulations of changes in the variance of temperature demonstrate that inter-annual sediment $\delta^{18}O$ values are not strongly influenced by stochastic temperature changes. This result can be explained by the much lower standard deviation of average annual tempera-

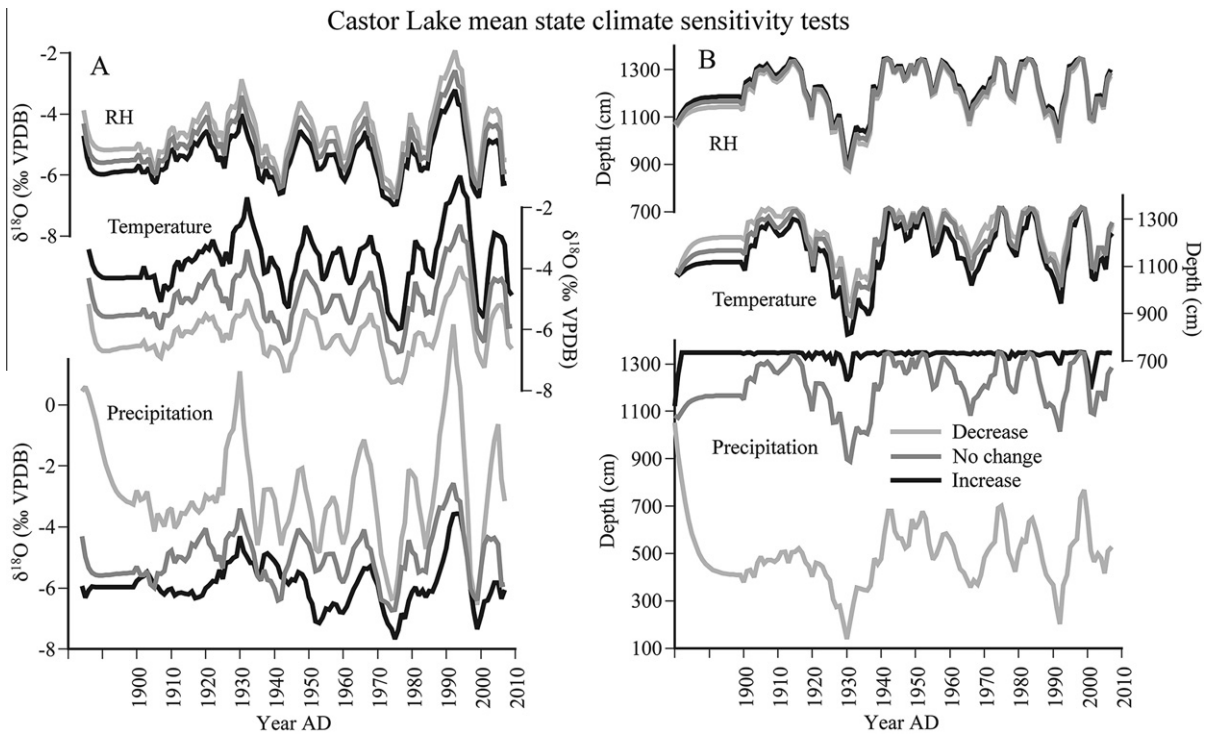

Fig. 6. CL average May–July (A) sediment $\delta^{18}O$ values and (B) depths (at the deepest part of the lake) from mean state sensitivity simulations in which monthly, instrumental RH (±5%), temperature (±2 °C), and precipitation (±50%) datasets were separately increased (black line), held constant (gray line), or decreased (light gray line). Five-year moving averages of $\delta^{18}O$ data are shown to facilitate comparison of multi-year trends. Monthly average climate data altered by the amounts described above were applied during the equilibration period (i.e., during the 20 simulation years prior to 1900) and maintained thereafter.

ture relative to that of precipitation in north-central Washington (at least over the instrumental period) (Table EA1) as well as the counteracting seasonal isotopic effects of constant temperature offsets (as previously discussed) (Steinman et al., 2010a). Note also that alterations in the standard deviation of temperature produced minimal change in both short-term and long-term lake levels whereas similar variance changes in precipitation produced relatively extensive changes in lake level at all timescales. For both lakes, the average modeled instrumental period lake level and volume was greater in the increased precipitation standard deviation scenario, suggesting that variability in stochastic precipitation can also influence long-term lake hydrology through non-linear catchment runoff responses to large precipitation events.

Interestingly, isotopic shifts in response to lake level change in the altered climate variance simulations did not follow the generally accepted interpretation of sediment core $\delta^{18}O$ records in which higher lake levels are associated with lower $\delta^{18}O$ values (and vice versa) (Fig. 8; Table EA2). For example, lake level increases (of ~35 cm) resulted in a small increase in average lake water and sediment $\delta^{18}O$ values (of <0.1‰) for CL with larger changes (of <0.2‰) for SL. This is due, in part, to the volumetric outseepage configuration applied here (Steinman et al., 2010a) and the fact that the SA:V ratio increases slightly over the range of lake level variability produced by the variance change simulations. As lake levels rise, for example, the SA:V ratio increases, leading to greater water loss through evaporation, less water loss through outseepage, and the enrichment of water isotopes. These simulations therefore demonstrate that in lakes with SA:V ratios that increase with increasing depth, it may be possible for lake $\delta^{18}O$ values to rise in response to larger precipitation amounts. This scenario is, however, contingent on lake throughflow and/or outseepage characteristics, and may be (at least to some extent) an artifact of the outseepage algorithms applied in the Steinman et al. (2010a) model. Studies have demonstrated that outseepage through the lake bed decreases exponentially with offshore distance from the shoreline (McBride and Pfannkuch, 1975; Lee, 1977; Pfannkuch and Winter, 1984), a phenomenon that should be accounted for in future lake modeling efforts.

3.4. Variable initial conditions simulations

The hydrologic and isotopic mass balance of a lake is dependent upon the hydrologic and isotopic conditions of the lake in many prior years, with the duration of this influence controlled in part by the volume (and associated proportions of volumetric change in response to climate forcing events) and residence time of the lake. Consequently, interpretation of sediment oxygen isotope values cannot be limited to the characteristics of local climate conditions during the time of sediment formation, but must instead consider the effects of cumulative hydroclimatic

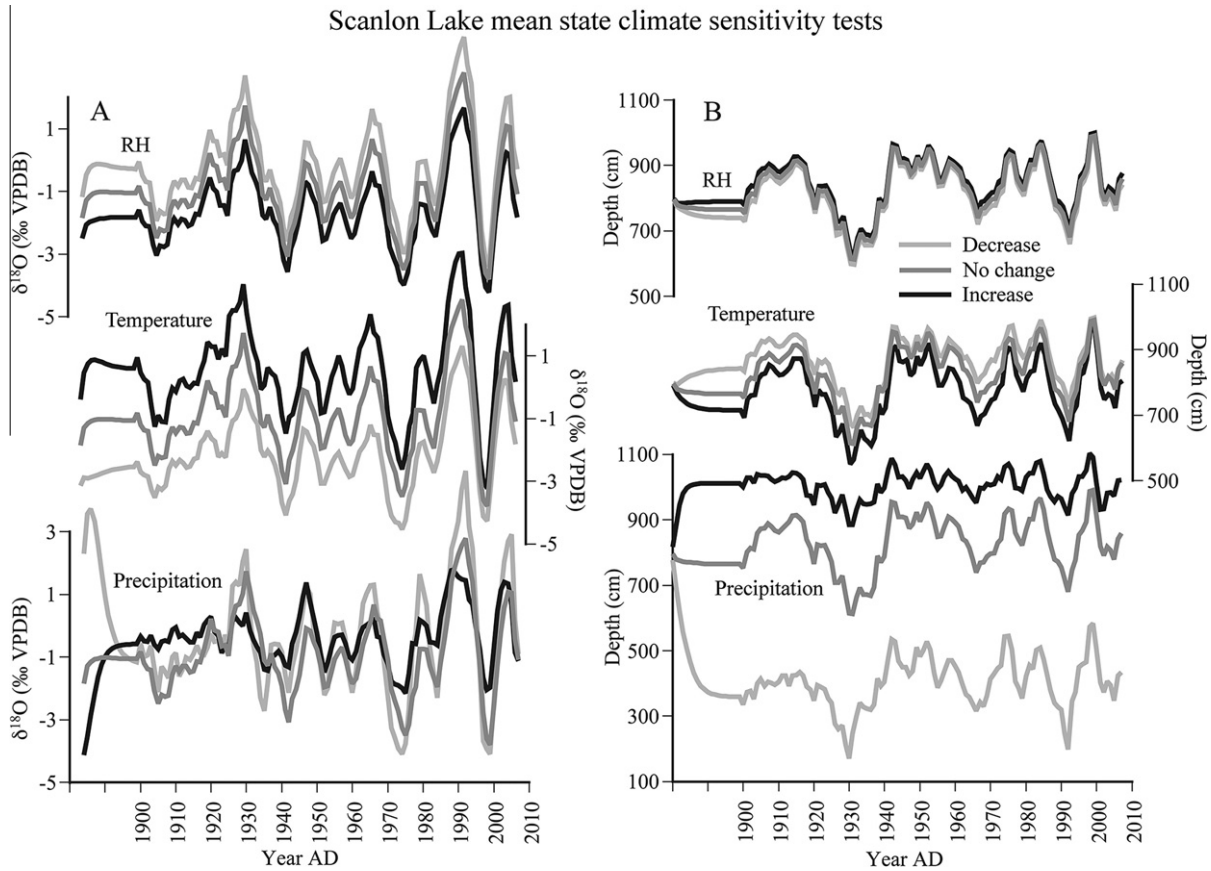

Fig. 7. SL average May–June (A) sediment $\delta^{18}O$ values and (B) depths (at the deepest part of the lake) from mean state sensitivity simulations in which monthly, instrumental RH (±5%), temperature (±2 °C), and precipitation (±50%) datasets were separately increased (black line), unchanged (gray line), or decreased (light gray line). Five year moving averages of $\delta^{18}O$ data are shown to facilitate comparison of multi-year trends. Monthly average climate data altered by the amounts described above were applied during the equilibration period (i.e., during the 20 simulation years prior to 1900) and maintained thereafter.

conditions. This conclusion (i.e., that lakes exhibit a reddened response to climate change) has significant implications for both qualitative and quantitative interpretations of sediment core isotope records. Within high resolution $\delta^{18}O$ or δD records (i.e., lake isotope records with annual or near-annual resolution) from closed-basin lakes, large isotopic excursions may not always be attributable to proportionately large magnitude droughts or wet periods in the years immediately preceding the time of sediment formation. That is, short-term, large scale isotopic variations may not necessarily be representative of equivalently large climate variations. Extracting useful quantitative information from sediment isotope records on mean state hydroclimatic conditions may therefore require model based analysis of average $\delta^{18}O$ values over time periods long enough to minimize the potential influence of the preceding hydrologic and isotopic conditions.

In the random initial condition simulations presented here, both CL and SL exhibited considerable variance (as defined by the 2σ threshold for $\delta^{18}O$ and volume values of 0.25‰ and 2.5% of total volume, respectively) in lake volume and $\delta^{18}O$ values between individual model runs for approximately the first 10 years of 20th century. More specifically, modeled CL and SL sediment isotopic values produced through the application of randomly generated hydroclimate data converged in 1908 and 1912, respectively (Fig. 9); whereas modeled lake volumes converged in 1907 and 1911, respectively (Fig. 10). As discussed above, in volumetrically smaller lakes with shorter residence times (e.g., SL, which has a residence time of ~2 years, Steinman et al., 2010a), the percentage volumetric change, and consequently, the isotopic response to hydrologic forcing is larger than in volumetrically larger lakes with longer residence times (e.g., CL, ~2.5 years). In larger lakes, however, the long residence times prolong the influence of any individual hydroclimatic perturbation relative to smaller lakes. The combination of these two characteristics (i.e., longer residence times in the case of volumetrically larger lakes, and larger percentage volumetric changes in response to hydrologic forcing in the case of smaller lakes) leads to a memory effect (i.e., persistence of the lake response to a climatic forcing event) of 8–12 years (i.e., 3–6 residence times) in closed lakes with volumes similar to those of this study. In other terms, the maximum duration of a hydrologic or isotopic anomaly derived from a climatic event is no greater than ~12 years, meaning that reddening of lake sediment $\delta^{18}O$ records is not substantially controlled by the isotopic and hydrologic state of the lake or climatic changes that occur before this 12 year time period.

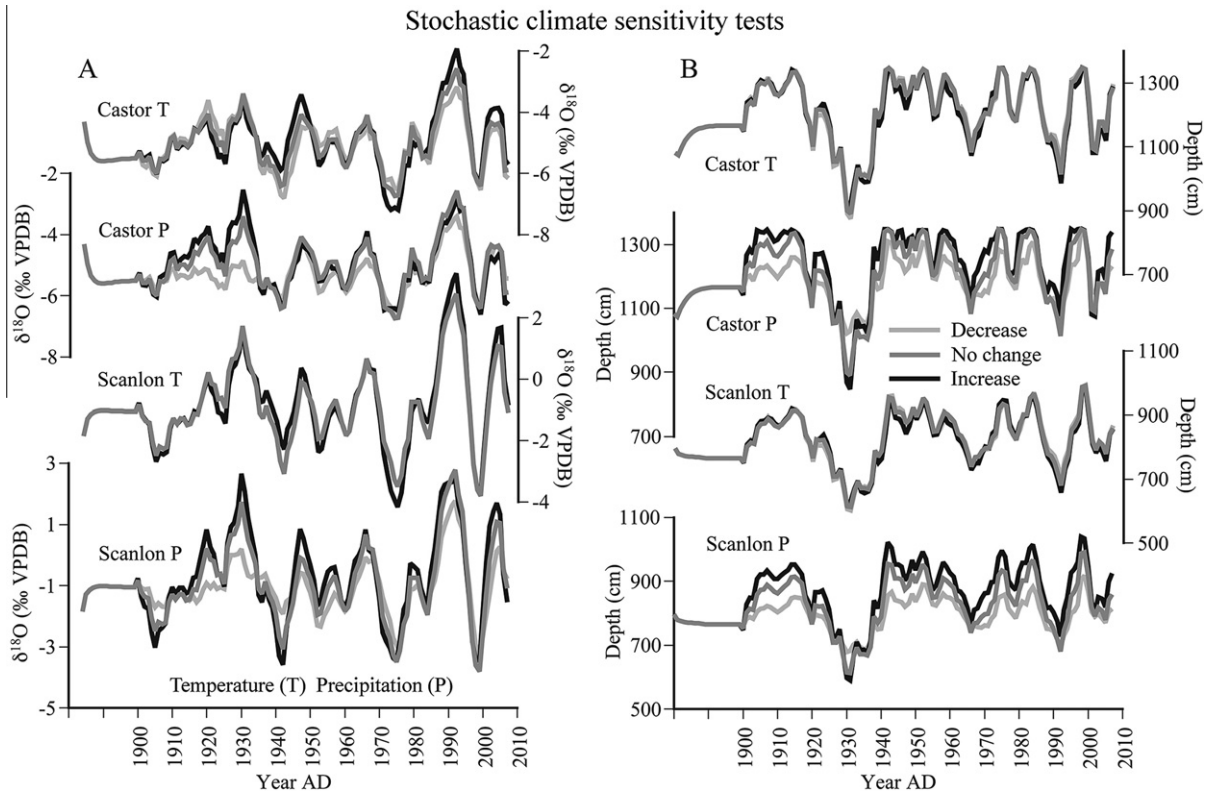

Fig. 8. CL and SL average (A) spring/summer (May–July) sediment $\delta^{18}O$ values and (B) depths (at the deepest part of the lake) from stochastic forcing simulations in which the standard deviation of average annual instrumental temperature and total annual precipitation were altered by a fixed percentage (±50%). Five year moving averages of $\delta^{18}O$ data are shown to facilitate comparison of multi-year trends.

If moving averages are applied to modeled $\delta^{18}O$ values, differences between the initial condition simulation results become smaller (Fig. 9), suggesting that model simulations can be used to quantitatively analyze sediment oxygen isotope records over discrete time periods with limited concern for initial conditions (and resulting reddening). For example, the 2σ value of the 10 year moving averages of the modeled CL and SL sediment $\delta^{18}O$ results suggest that ~95% of all randomly generated initial condition scenarios will produce values within $\sim \pm 0.6‰$ and $\sim \pm 1.0‰$, respectively, of the 10 year average $\delta^{18}O$ value in 1905. If 20 year moving averages are applied, this uncertainty range decreases to $\sim \pm 0.3‰$ and $\sim \pm 0.5‰$, respectively, in the year 1910. Effectively, this means that for CL reddening due to unknown initial conditions contributes at most ±0.3‰ of uncertainty to sediment $\delta^{18}O$ values if 20 year averages are applied and climatic changes do not exceed the limits of the Monte Carlo domain applied here. Notably, we conducted three additional sets of simulations in which random data was applied until simulation year 1910, 1920, and 1930, respectively (instead of until year 1900 as in the simulation results presented above), with no significant change, demonstrating that decadal variance in temperature and precipitation trends during the interval that immediately follows the period of random data application do not substantially affect the initial condition analysis. These results provide the basis for the application of 20 year averages to the CL $\delta^{18}O$ record in the Monte Carlo based climate reconstruction.

3.5. Monte carlo simulations

November–February precipitation exhibited the strongest correlation to simulated 20 year average $\delta^{18}O$ values ($R^2 = 0.49$) (Fig. 11; Table 1). The strongest temperature correlation to sediment $\delta^{18}O$ values also occurred in the November–February interval ($R^2 = 0.35$). RH in all seasons correlated with similar strength to average sediment $\delta^{18}O$ (with R^2 values of ~0.2). Prediction intervals (at the 95% level) in the November–February precipitation–$\delta^{18}O$ correlations averaged ±37 mm.

Modeled precipitation values (developed using the Monte Carlo method) for the 1900–2007 period strongly correlate to instrumental observations with an R^2 value of 0.66 (Fig. 12). The magnitude of model reconstructed precipitation variations was larger than that of observations, with differences between maximum and minimum values of 14 and 27 mm, respectively. Nonetheless, reconstruction limits of the modeled precipitation values overlap with observations throughout the instrumental period.

The most conspicuous result of the Monte Carlo simulations is the relatively strong correlation for CL between November–February precipitation and sediment $\delta^{18}O$ values and the weaker relationship with temperature and RH in all seasons (Fig. 11; Table 1). Results from the mean state and variance model analyses, as well as observations of lake water $\delta^{18}O$ variations (Figs. 3 and 4), are largely consistent with the data generated in the Monte Carlo tests, demon-

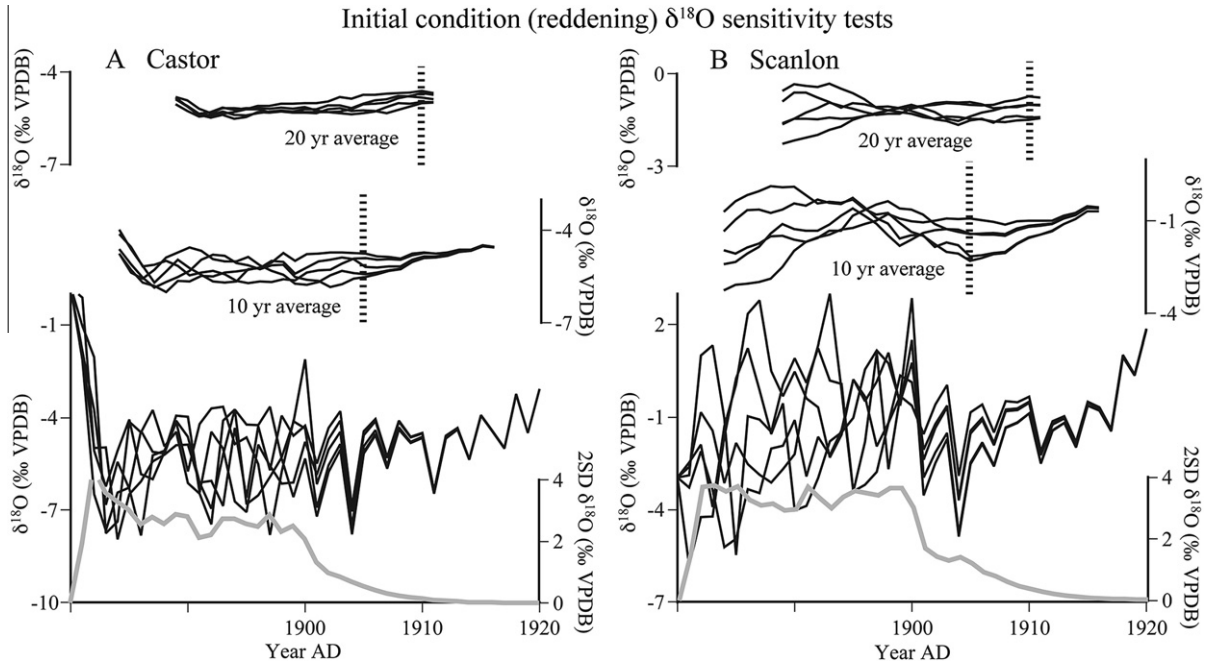

Fig. 9. (A) CL and (B) SL average spring/summer month (May–July) sediment $\delta^{18}O$ data from initial condition sensitivity simulations in which random precipitation, temperature, and RH data were applied during the equilibration period. The gray line depicts the 2σ value of the $\delta^{18}O$ values for each year of the 50 simulation dataset. Ten and 20 year moving averages are displayed to demonstrate the effect of averaging on error contributed by variable initial conditions. The vertical dashed lines mark the year in which the data comprising the moving averages all falls within the 20th century as part of the instrumental dataset. Results from the first five (of 50) simulations are displayed to maintain clarity.

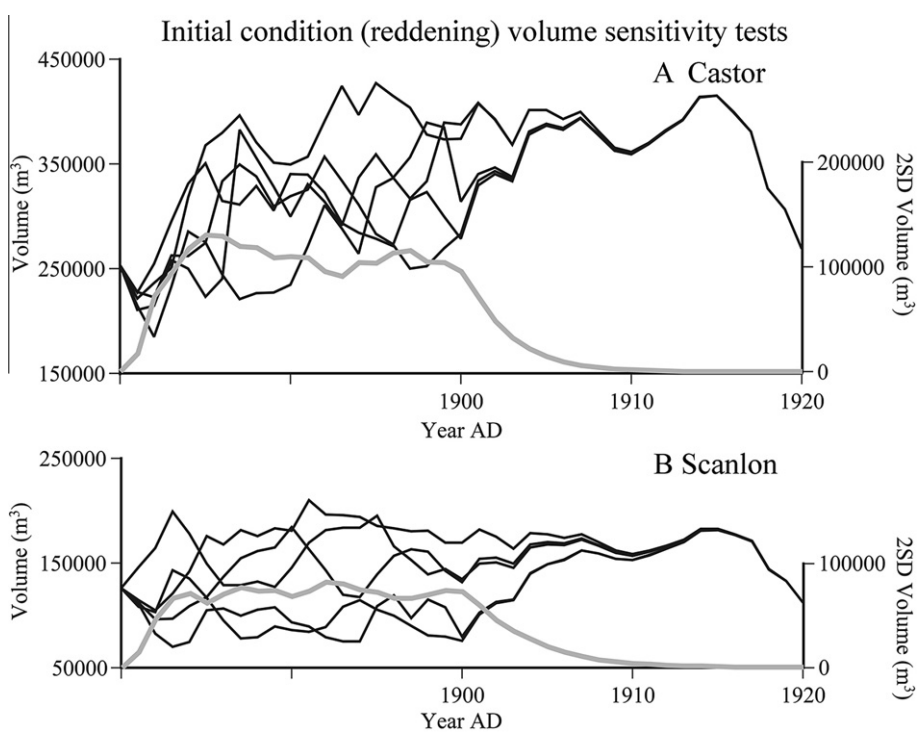

Fig. 10. (A) CL and (B) SL average spring/summer month (May–July) volume data from initial condition sensitivity simulations in which random precipitation, temperature, and RH values were applied during the equilibration period. The gray line depicts the 2σ value of the volume data for each year of the 50 simulation dataset. Results from the first five (of 50) simulations are displayed to maintain clarity.

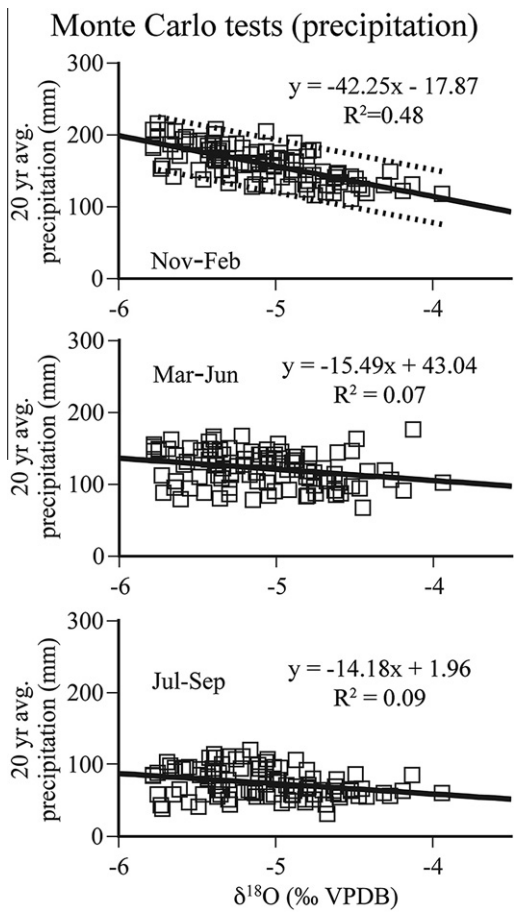

Fig. 11. Monte Carlo simulation results (open squares) relating modeled CL May–July average sediment $\delta^{18}O$ values to 20 year average November–February, March–June, and July–October precipitation amounts. Linear regression equations (solid lines) and associated 95% prediction limits (dashed lines) demonstrate the range of uncertainty in the relationship between average sediment $\delta^{18}O$ values and seasonal precipitation amounts. Changes in average climate data values (i.e., the Monte Carlo domain) were limited to the maximum offset from the observed mean of 20 year average instrumental values.

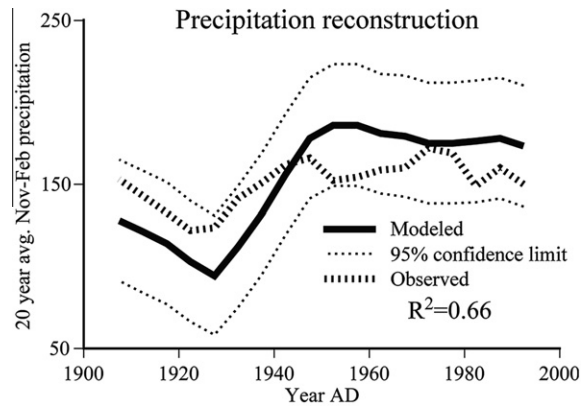

Fig. 12. Twenty year average observed (coarse dashed line) and reconstructed (solid line) November–February precipitation for the Castor Lake region over the instrumental period (1900–2007 AD) determined using the linear relationship established in the Monte Carlo tests (Fig. 11). Fine dashed lines depict estimated 95% prediction limits for the precipitation reconstruction.

strating that winter precipitation is the strongest hydroclimatic control on lake water and sediment $\delta^{18}O$ values in central Washington and potentially in other regions with Mediterranean climates (assuming similar catchment settings). This is in part a result of the seasonal distribution of precipitation in which ~45% falls during November–February and the low evapotranspiration rates during the winter months that produce large positive catchment water balance. Other studies have demonstrated similar lake hydrologic sensitivity to winter precipitation in regions with more evenly distributed (Vassiljev, 1998; Vassiljev et al., 1998) and summer dominated (Shuman and Donnelly, 2006) seasonal precipitation, indicating that in many cases lakes are more hydrologically sensitive to winter precipitation variability than to other climate variables. Interestingly, March–June precipitation amounts (which is wet relative to the summer) do not strongly correlate to sediment $\delta^{18}O$ values in the model tests. This is likely due to the influence of temperature on soil water mass balance and consequent catchment runoff amounts (as discussed above). The relatively strong correlation between winter and spring temperature (relative to summer) and sediment $\delta^{18}O$ is somewhat surprising considering that evaporation losses (which are largely controlled by temperature) are considerably higher in the summer relative to the spring and winter. This can be explained by the model algorithm that controls precipitation $\delta^{18}O$ values on the basis of temperature (0.6‰/°C) (Dansgaard, 1964; Rozanski et al., 1992, 1993). In the wetter seasons of March–June and November–February, higher temperatures lead to isotopically heavier lake water inflow and consequently larger sediment $\delta^{18}O$ values (and vice versa) such that winter and spring temperature is more isotopically influential than that of summer.

The combined effects of randomness in climate and reddening produced uncertainty in model based reconstructions of precipitation amounts, an expected result given the equifinality inherent in the climate-lake hydrologic and isotopic system. This is reflected by the large prediction limit range (at the 95% level) of approximately ±37 mm (for November–February precipitation). In turn, this suggests that quantitative precipitation reconstructions developed on the basis of Monte Carlo methods will be probabilistic in nature, in that specific quantitative climate solutions are unattainable. Reduction of uncertainty in precipitation solutions could potentially be achieved in several ways, namely, through further constraint of temperature and RH ranges in the Monte Carlo simulations or reduction in initial condition uncertainty. Climate variable constraint could potentially be realized through analysis of other paleo-proxy records such as tree-ring based reconstructions of temperature. A lessening of initial condition uncertainty could potentially be achieved through sequential analysis of climate solutions on the basis of changes in sediment core isotope values. To this end, it may be possible to establish a range of potential initial conditions through application of solutions for sediment core isotopic

values within preceding (i.e., older) sections of a given core. Increasing the averaging interval for model and sediment $\delta^{18}O$ analysis (e.g., analyzing discrete time periods of greater than 20 years) would also reduce the uncertainty contributed by reddening, albeit at the cost of decreased temporal resolution. Regardless, validation of the reconstruction through comparison to instrumental climate data is essential to the constraint of potential error, implying that increasing the length of the averaging interval may disallow an adequate comparison if the observational time period is short. An additional source of uncertainty lies in the Monte Carlo domain (i.e., the range of modeled climate data variations). It is likely, for example, that application of different limits for precipitation and temperature data would lead to changes in the regression equations that describe the relationship between modeled sediment $\delta^{18}O$ values and climate. Future paleo-interpretation applications should therefore investigate model sensitivity to the Monte Carlo domain through a series of tests in which disparate average precipitation and temperature limits are applied.

The similarity between the observed and reconstructed 20th century November–February precipitation amounts in north-central Washington validates the Monte Carlo method presented here (Fig. 12). Some of the discrepancy between the observed and modeled records is likely due to age model errors in the CL sediment core (which Steinman et al., in press estimated at ± 10 years), as well as sedimentation rate variance, which can lead to over- and under-representation of individual years in the sediment $\delta^{18}O$ record. Additional error in model reconstructions potentially results from the outseepage configuration used to describe CL groundwater throughflow, the model calibration methods (Steinman et al., 2010a) and uncertainty in instrumental records of precipitation.

Improvement in the model outseepage algorithm could be attained through the application of more complex equations that calculate outseepage as a function of offshore distance in accordance with estimates (or ideally, measurements) of lake sediment hydraulic conductivity values. Piezometer based water sampling and geochemical analyses could be used to provide salinity data for application to conservative ion mass balance modeling exercises which could improve model calibration by providing additional constraint on lake outseepage. Model calibration could also be improved through the application of additional observational data such as changes in lake level, water $\delta^{18}O$, and stratification. Collectively these additions would help to better characterize lake hydrologic and isotopic responses to hydroclimatic forcing and would thereby improve the accuracy and precision of future model based estimates of paleoclimatic conditions.

4. CONCLUSIONS

This study demonstrates that although it may not be possible to attribute a specific state of climate (i.e., a specific temperature, precipitation, and relative humidity value) to a measured lake water or sediment core $\delta^{18}O$ value, probabilistic analysis of $\delta^{18}O$ data using Monte Carlo simulations conducted with lake catchment hydrologic and isotope mass balance models can help to establish quantitative limits for past hydroclimatic conditions. Application of these methods to newly developed lake sediment $\delta^{18}O$ (and δD) records may not, however, be appropriate unless preliminary studies are undertaken to assess the lake hydrologic setting and to establish a working model that reproduces lake hydrologic and isotopic responses to climate change. Markedly, the study lake must be isolated from regional groundwater sources and receive minimal extra-catchment groundwater inflow. The well defined, elevated catchments of CL and SL appear to satisfy these conditions and likely represent a good example to follow in a preliminary assessment of lake catchment hydrology (i.e., if a lake is in a similar setting, it is likely that most inflowing water is catchment derived). In glaciofluvial settings with hummocky, karstic terrain (and poorly defined catchments) this requirement may not be met, leading to groundwater that has experienced isotopic enrichment through evaporation and/or is extra-catchment derived, which would contribute error in lake hydrologic mass balance estimates. Confirmation of these assumptions should be undertaken through a lake and catchment water isotope sampling program designed to investigate the extent of lake evaporative enrichment (and thereby the degree of hydrologic closure) through comparison to local meteoric water (ideally inflowing catchment groundwater). Perhaps more importantly, the resulting lake water isotope analyses should be used as a comparative dataset (if collected over several years and in different seasons) for validating hydrologic and isotope mass balance model estimates of lake outseepage and intra-annual isotopic and hydrologic variability. Several studies (e.g., Benson and Paillet, 2002; Jones and Imbers, 2010; Steinman et al., 2010a) describe lake hydrologic and isotope mass balance methods and provide guidelines for adapting models to specific lakes and assessing performance through comparison to observations of lake hydrologic and isotopic variability. Other studies (e.g., Roberts et al., 2008, 2012) provide useful criterion for assessing the primary controls on lake $\delta^{18}O$ on the basis of catchment and climatic setting that may be helpful for initial investigations and identification of potential study sties. In situations where extensive lake monitoring cannot be undertaken, or if one or more lake-catchment characteristics is underdetermined (e.g., lake bathymetry or catchment morphology is not well established) it may still be possible to use modeling methods to assess the geochemical and hydrologic responses of a similar (if not identical) lake to climate change (e.g., Stansell et al., 2013). Under these circumstances quantitative interpretations of sediment $\delta^{18}O$ records may not be appropriate, but modeling exercises can still provide insight into lake sensitivity to climate forcing and thereby provide a framework for producing semi-quantitative interpretations.

Of considerable importance, sediment cores with suitable analytical material for isotopic and dating analyses are required for any lake based paleoclimate investigation. Endogenic and biogenic carbonates are commonly used to produce paleoclimate records but care must be taken when modeling and interpreting sediment in which mixed mineral phases (Shapley et al., 2009), detrital contamination (Mangili et al., 2010) or multiple ostracod species are present or

when mineral formation has occurred in isotopic disequilibrium with lake water (Fronval et al., 1995; Teranes et al., 1999; Leng and Marshall, 2004). The use of aquatic cellulose (Wolfe et al., 2007), chitin (Wooller et al., 2008) and lipids (Sachse et al., 2004; Sachse and Sachs, 2008; Sachs et al., 2009) to produce isotopic records has become more common in recent years and should provide additional opportunities for paleoclimate studies in regions that lack carbonate mineral rich lakes. Regardless, a well dated surface core for use in modern reconstruction and comparison to instrumental data is essential if the objective is to produce a statistically validated climate reconstruction.

The first manuscript (Steinman and Abbott, 2012) in this two part series of papers provides a resource for assessing error contributed by uncertainty in model parameters and variables (e.g., climate data, the timing of carbonate mineral formation) and establishes methods for validating model performance through comparison to lake sediment $\delta^{18}O$ records spanning the instrumental time period. Building upon this work, the research presented here outlines Monte Carlo methods for interpreting lake sediment $\delta^{18}O$ records in the context of additional sources of uncertainty (persistent disequilibrium, reddening and equifinality) that are inherent aspects of lake-catchment physics. Future research should focus on applying these methods to lake systems in different climatic settings, ideally on lakes for which substantial observational datasets and well dated (preferably annually varved) lake sediment records exist. Although additional work is necessary to better characterize the sensitivity of model derived hydroclimatic reconstructions to unknown or underdetermined lake-catchment hydrologic characteristics (e.g., via piezometer based studies), application of this Monte Carlo method to pre-instrumental sediment records would represent a considerable advancement in the use of lake sediments as paleo-proxy archives. Such a study could potentially produce quantitative paleoclimate datasets with wide-ranging applications including sensitivity parameterization in climate model simulations (Goosse et al., 2006, 2010) or the refinement of water management policies in drought stressed regions such as the upper Columbia River basin.

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APPENDIX A. SUPPLEMENTARY DATA

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.gca.2012.11.027.

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