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Stable isotope mass balance of lakes: a contemporary perspective

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ABSTRACT

The theoretical basis for application of stable isotope mass balance of lakes is described for a range of climatic situations including low latitude, high latitude, high altitude, continental and coastal systems, as well as cases where the atmospheric boundary layer is significantly modified by the lake evaporation process. The effects of seasonality on isotopic offset between precipitation and atmospheric vapour and the slope of the local evaporation line are described. Atmospheric feedback and its role in labelling the isotopic composition of the Laurentian Great Lakes and tropical lakes is discussed. Several important considerations are suggested to improve parameterization of quantitative paleoclimatic reconstructions including use of assumptions that are appropriate for the climatic setting, for the atmospheric feedback situation, for salinity, and headwater setting. Potential for use of dual-isotopes to trace past changes in seasonality and input, and a dual-lake index method that can potentially be used to trace connectivity of lakes are presented. In cases where modern or paleo-evaporation systems may be under-defined there are inherent limitations in the degree of quantification that can be attained.

1. Introduction

The investigation of stable water isotopes (mainly $\delta^{18}O$) records from terrestrial material, as an alternative to marine sediments, was propelled to the forefront of paleoclimate studies in the early 1990s (Swart et al., 1993). The development of water isotope records from lake sediments, in substrates such as carbonates, biogenic silica, kerogen, sedimentary cellulose, and lipids, is typically undertaken with the goal of reconstructing the isotopic history of lake water (Edwards et al., 2004). Analogous to marine studies (Emiliani, 1955; Shackleton and Opdyke, 1973) stratigraphic changes in $\delta^{18}O$ in the lacustrine environment were initially attributed largely to changes in lake or air temperature (Eicher and Siegenthaler, 1976) and interpreted based on the understanding of the temperature dependence of isotopic fraction in the formation of carbonate (Epstein et al., 1953; Friedman and O'Neil, 1977). A number of studies have also utilized modern analogues to develop empirical relationships between temperature and various isotopic proxies (von Grafenstein et al., 1996; Schleser et al., 1999; Wooller et al., 2004). However, an increasing number of studies have pointed out the apparent limitations of this simplified transfer function approach, including the inherent control that water balance and climatic conditions play in determining lake water isotopic composition (Edwards et al., 2004; Jones et al., 2005; Henderson and Shuman, 2009; Steinman et al., 2010a).

More recently, as an emerging trend in the paleoclimate community, multi-disciplinary investigations have sought to gain insight from more holistic contemporary monitoring of spatial and temporal isotope systematics to inform the interpretation of isotope records in the lake sediment record (Benson, 1994; Roberts et al., 2008; Shapeley et al., 2008; Wolfe et al., 2012; Steinman et al., 2013; Zolitschka et al., 2013). Several studies have developed coupled hydrology and isotope mass balance (IMB) models for specific lake systems (Hostetler and Benson, 1994; Benson and Paillet, 2002; Shapley et al., 2008; Jones and Imbers, 2010; Steinman et al., 2010a,b), while others have demonstrated the sensitivity of paleo-archives to hydrological setting (Jones et al., 2007; Shapley et al., 2009; Steinman et al., 2010b, 2012, 2013; Steinman and Abbott, 2013). IMB models were found to be informative for these purposes as they provide a theoretical framework to simulate and quantitatively interpret isotopic signals in lakes (e.g. Dincer, 1968; Gonfiantini, 1986; Gat, 1995). Increasingly, it has been realized that utilization of IMB becomes essential for the

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proper interpretation of lacustrine isotope records as paleoclimate and paleoenvironmental proxies (Cross et al., 2001; Leng and Marshall, 2004; Anderson et al., 2007; Wolfe et al., 2007; St. Amour et al., 2010). Other requirements include specific understanding of local isotope–climate relations and substrate-specific fractionations (e.g. Marshall, 1992; Swart et al., 1993; Kim and O'Neil, 1997; Sauer et al., 2001; Huang et al., 2004).

This review is a process-based summary of important controls on $\delta^{18}O$ or δ^2H in the IMB model. Here we describe IMB as applied in many contemporary studies, discuss the assumptions made to characterize atmospheric moisture and fractionation factors, present calibration approaches such as use of an index lake, describe the influence of headwater conditions, seasonality and atmospheric feedbacks, and configurations of the IMB for specific settings including low latitude, high-latitude, continental and coastal systems, and chain of lakes. Appropriate assumptions for application of IMB to paleo-environmental reconstructions are also described.

2. Theory

2.1. Isotope mass balance

The annual water-mass and isotope-mass balance for a well-mixed lake may be written respectively as

$$dV/dt = I - Q - E \quad \left(\mathrm{m}^3 \cdot \mathrm{year}\right) \tag{1}$$

$$V\frac{d\delta_L}{dt} + \delta_L\frac{dV}{dt} = I\delta_I - Q\delta_Q - E\delta_E \quad \left(\text{‰} \cdot \mathrm{m}^3 \cdot \mathrm{year}\right) \tag{2}$$

where V is the volume of the lake, t is time, dV is the change in volume over time interval dt, I is instantaneous inflow where $I = I_U + I_R + I_G + P_L$; I_U is channelized inflow from upstream lakes, I_R is surface inflow from the catchment area, I_G is groundwater inflow, P_L is precipitation on the lake surface; $Q = Q_Y + Q_Z$ is instantaneous outflow where Q_Y is surface outflow, Q_Z is groundwater outflow; E is evaporation; δ_L is the isotopic composition of the lake; δ_I, δ_U, δ_R, δ_G and δ_P are the isotopic compositions of total inflow and its components, i.e. channelized inflow from upstream lakes, surface inflow, groundwater inflow, and precipitation, respectively; and δ_Q, δ_Y and δ_Z are the isotopic compositions of total outflow and its components, i.e. surface outflow and groundwater outflow, respectively. Here, δ values are defined as isotopic ratios reflecting deviation in per mil (‰) from Vienna-SMOW (Standard Mean Ocean Water), where $\delta_{\mathrm{SAMPLE}} = 1000((R_{\mathrm{SAMPLE}}/R_{\mathrm{SMOW}}) - 1)$, and R is $^{18}O/^{16}O$ or $^{2}H/^{1}H$. Values presented here are normalized on the SMOW-SLAP (Standard Light Arctic Precipitation) scale.

Note that $\delta_I = (\delta_U I_U + \delta_R I_R + \delta_G I_G + P_L\delta_P)/I$ and $\delta_Q = (\delta_Y Q_Y + \delta_Z Q_Z)/Q$, the latter of which, δ_Q and its subcomponents, δ_Y and δ_Z will be identical and approximately equal to δ_L in well mixed lakes to maintain the isotope balance. Three main hydrologic settings can be distinguished including (i) desiccating water bodies ($dV/dt < 0$), where inflow occurs once or periodically, and water bodies at hydrologic steady state ($dV/dt = 0$), including (ii) terminal lakes, where inflow is balanced by evaporation ($I = E$), and (iii) throughflow lakes, where inflow is balanced by evaporation and outflow ($I = E + Q$) (Fig. 1; see also Horita, 1990; Gat, 1981).

Based on the linear resistance model of Craig and Gordon (1965) and using the convention of Gonfiantini (1986) for describing the equilibrium fractionation α^+ as a quantity slightly greater than 1, δ_E can be estimated by:

$$\delta_E = \left(\frac{\delta_L - \varepsilon^+}{\alpha^+} - h\delta_A - \varepsilon_K\right) \Big/ \left(1 - h + 10^{-3}\cdot\varepsilon_K\right) \quad (\text{‰}) \tag{3}$$

where h is the relative humidity normalized to water surface temperature (decimal fraction), δ_A is the isotopic composition of atmospheric moisture, ε^+ is the equilibrium isotopic separation (see Horita and Wesolowski, 1994) being related to α^+ by $\varepsilon^+ = (\alpha^+ - 1)\bullet 1000$, and $\varepsilon_K = \theta\cdot C_K\bullet(1-h)$ is the kinetic isotopic separation, where θ is a transport resistance parameter, commonly assumed to be unity when δ_A and h are measured or estimated close to the interface (Gat, 1995). $\theta = 1$ indicates that the evaporation rate is controlled by molecular transport of water through the laminar sublayer (see Horita et al., 2008). C_K is a kinetic constant described later on. Note that air–water isotopic exchange will proceed and the lake water will enrich (or deplete) to an isotopic steady-state reflective of the isotopic and hydrologic characteristics of the system, as described below.

Substitution of eq. (3) into eq. (2) assuming well-mixed conditions yields:

$$V\frac{d\delta_L}{dt} + \delta_L\frac{dV}{dt} = I\delta_I - Q\delta_L - \frac{E}{1 - h + \varepsilon_K/1000} \times \left(\frac{\delta_L - \varepsilon^+}{\alpha^+} - h\delta_A - \varepsilon_K\right) \quad \left(\text{‰}\cdot\mathrm{m}^3\cdot\mathrm{year}\right) \tag{4}$$

which under constant atmospheric and hydrologic conditions (i.e. hydrologic steady state, such that $dV/dt = 0$) simplifies to (Gonfiantini, 1986):

$$d\delta_L/dt = -\left[(1 + mx)\delta_L - \delta_I - x\delta^*\right](I/V) \quad (\text{‰}) \tag{5}$$

where $x = E/I$ is the fraction of lake water lost by evaporation, $1-x = Q/I$ (being the fraction of water lost to liquid outflows), and

$$m = \left(h - 10^{-3}\cdot\left(\varepsilon_K + \varepsilon^+/\alpha^+\right)\right) \Big/ \left(1 - h + 10^{-3}\cdot\varepsilon_K\right) \quad (\text{dimensionless}) \tag{6}$$

is the temporal enrichment slope, and

$$\delta^* = \left(h\delta_A + \varepsilon_K + \varepsilon^+/\alpha^+\right) \Big/ \left(h - 10^{-3}\cdot\left(\varepsilon_K + \varepsilon^+/\alpha^+\right)\right) \quad (\text{‰}) \tag{7}$$

is the limiting isotopic composition. It is significant to note that δ^* is the isotopic composition that a desiccating water body would approach under non-steady-state conditions as it dries up (i.e. $V \to 0$).

Integrating eq. (5) with respect to time, assuming constant values for δ_I, δ_A, ε^+, ε_K, h, I, Q, E yields (Gonfiantini, 1986):

$$\delta_L(t) = \delta_S - (\delta_S - \delta_0)\exp[-(1 + mx)(It/V)] \quad (\text{‰}) \tag{8}$$

where $\delta_L(t)$ describes the change in isotopic composition of the lake with time t, δ_0 is the initial isotopic composition of the lake at t_0, and δ_S is the steady-state isotopic composition that the lake approaches as $t \to \infty$. The steady state isotopic composition δ_S is given by Gonfiantini (1986) and Gat (1995) as:

$$\delta_S = (xm\delta^* + \delta_I)/(1 + mx) \quad (\text{‰}) \tag{9}$$

which can be rearranged to provide an expression to estimate x (or E/I):

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