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Anthropogenic basin closure and groundwater salinization (ABCSAL)

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ABSTRACT

Global food systems rely on irrigated agriculture, and most of these systems in turn depend on fresh sources of groundwater. In this study, we demonstrate that groundwater development, even without overdraft, can transform a fresh, open basin into an evaporation dominated, closed-basin system, such that most of the groundwater, rather than exiting via stream baseflow and lateral subsurface flow, exits predominantly by evapotranspiration from irrigated lands. In these newly closed hydrologic basins, just as in other closed basins, groundwater salinization is inevitable because dissolved solids cannot escape, and the basin is effectively converted into a salt sink. We first provide a conceptual model of this process, called "**A**nthropogenic **B**asin **C**losure and groundwater **SAL**inization" (ABCSAL). We examine the temporal dynamics of ABCSAL using the Tulare Lake Basin, California, as a case study for a large irrigated agricultural region with Mediterranean climate, overlying an unconsolidated sedimentary aquifer system. Even with modern water management practices that arrest historic overdraft, results indicate that shallow aquifers (36 m deep) exceed maximum contaminant levels for total dissolved solids on decadal timescales. Intermediate (132 m) and deep aquifers (187 m), essential for drinking water and irrigated crops, are impacted within two to three centuries. Hence, ABCSAL resulting from groundwater development constitutes a largely unrecognized constraint on groundwater sustainable yield on similar timescales to aquifer depletion in the Tulare Lake Basin, and poses a serious challenge to groundwater quality sustainability, even when water levels are stable. Results suggest that agriculturally intensive groundwater basins worldwide may be susceptible to ABCSAL.

1. Introduction

Groundwater from major aquifer systems supplies 43% of the world's irrigation water (Siebert et al., 2010). As a result of excessive groundwater development and land use change, groundwater quantity and quality in these agriculturally intensive groundwater basins has been significantly impacted. Numerous global and regional studies document aquifer depletion related to agricultural withdrawal (Brush et al., 2013; Döll et al., 2012; Famiglietti, 2014; Faunt et al., 2009; Gleeson et al., 2012; Russo and Lall, 2017; Scanlon et al., 2012; Siebert et al., 2010; Vörösmarty et al., 2000; Wada et al., 2014). Anthropogenic contaminants to groundwater include nitrates, which originate from agricultural fertilizers (Burow et al., 2008), pesticides (Burow et al., 2008; Burow et al., 1998), and animal farming (Harter et al., 2012). Groundwater pumping may even mobilize naturally-occurring contaminants such as arsenic (Winkel et al., 2011; Smith et al., 2018) and uranium (Jurgens et al., 2008; Jurgens et al., 2010).

Another class of groundwater contaminants are total dissolved solids (TDS), also referred to as salts or salinity. TDS are sourced both naturally (e.g., produced by rock-water interactions) and anthropogenically (e.g., imported by surface water for irrigation). Elevated TDS is an indicator of human impact on freshwater systems (Ayers and Westcot, 1985; Kaushal et al., 2014, and reduces agricultural productivity Lopez-Berenguer et al., 2009; Munns, 2002; Pessarakli, 2016), which has prompted states to set agricultural irrigation water quality goals, (e.g., 450 mg/L in California) (CSWRCB, 2019a). For drinking water, the United States Environmental Protection Agency and the state of California recommend a secondary maximum contaminant level of 500 mg/L TDS (CSWRCB, 2019b; CSWRCB, 2019a). Water high in TDS may exhibit discoloration, unpleasant odor and taste, and may be unsuitable for human consumption or irrigation (Hem, 1985). Fresh water is defined as containing TDS less than 1,000 mg/L, brackish water ranges from 1,000 to 10,000 mg/L, and saline water ranges from 10,000 to 100,000 mg/L (Fetter, 2001).

Groundwater salinization is widely studied (Greene et al., 2016) in terms of (1) seawater intrusion (Bear et al., 1999; Werner et al., 2013),

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(2) naturally-occurring salinization in closed surface-water basins (i.e., endorheic basins and playas) (Eugster and Hardie, 1978; Hardie and Eugster, 1970), (3) high water tables causing groundwater evaporation and soil salinization via capillary rise (Datta and De Jong, 2002; Barrett-Lennard, 2003; Chaudhuri and Ale, 2014; Hillel, 1992), and (4) soil salinization due to irrigation (Hanson et al., 1999; Bernstein and Francois, 1973; Hillel, 2000). This study describes a fifth type of groundwater salinization that remains largely unexplored: salinization of an entire groundwater basin created by historically excessive pumping, then sustained by the inability of a closed groundwater system to discharge salts. Henceforth, we refer to this fifth type as "**A**nthropogenic **B**asin **C**losure and groundwater **SAL**inization" (ABCSAL).

This fifth type of salinization, ABCSAL, is related to naturally-occurring closed basin salinization (case (2) above), but has significantly different phenomenology. It is therefore useful to first consider the difference between an open, fresh hydrologic basin, and a naturally closed, saline basin.

An open, fresh groundwater basin has sufficient natural outlets for TDS, such as baseflow to streams and lateral subsurface flow across basin boundaries, which maintains a balance between salinity that is naturally generated within the basin (i.e., mineral dissolution) and salinity that is exported out of the basin. Basins containing fresh groundwater exist only because they have outlets for both the circulating groundwater and the dissolved salts therein, originating from intrabasin rocks and sediments Domenico and Schwartz, 1998.

In contrast, closed hydrologic basins – common in arid to semiarid regions worldwide – naturally form when (a) outflow by surface water or groundwater flows is absent or small, and (b) evaporation is the dominant mechanism by which water exits the basin (Hardie and Eugster, 1970; Eugster and Hardie, 1978; Jones and Deocampo, 2003). Because TDS concentrations in precipitation are low (around 10^1 mg/L), most TDS originates from rock-water reactions in surface runoff and in the subsurface. Salts may accumulate at the evaporative boundaries of the basin: at or immediately below the surface where discharging groundwater evaporates or at the bottom of a surface depression in terminal and sometimes ephemeral lakes that collect runoff, baseflow, and spring outflow (Wooding et al., 1997; Richter and Kreitler, 1986). Examples of naturally closed hydrologic basins with saline features at or near the land surface are found worldwide: playas and salt flats such as those in the Great Basin (USA) and Salar de Uyuni (Bolivia); saline lakes like the Great Salt Lake (USA) and the Dead Sea (Middle east); in extremely arid deserts such as the Arabian and Atacama; and in the unsaturated subsurface of semi-arid regions with insufficient precipitation to recharge groundwater (Scanlon et al., 1997; Kreitler, 1993).

In this paper, we argue that sufficient groundwater development can lower groundwater levels in an open to semi-open and relatively fresh basin, thus converting it into a closed basin, which then salinates in a distinctly different manner from those described in (1)–(4). First, moderate to large amounts of groundwater development may result in sufficient reduction of groundwater levels that reduce or eliminate natural baseflow to streams (Russo and Lall, 2017; Barlow and Leake, 2015; Hunt, 1999) and reverse existing groundwater gradients at subsurface outflow boundaries (Fig. 1A). Progressively greater closed basin conditions diminish and eventually entirely eliminate natural TDS export from the groundwater basin (Fig. 1B). Furthermore, if the basin is irrigated, crop evapotranspiration becomes the dominant water outflow from the basin, leaving behind salts that are returned to the groundwater basin via irrigation return flows and recharge from precipitation. Across the globe, water level stabilization in such overdrafted basins is sometimes achieved by importing additional surface water. However, water imports can add significant salt to the basin. Moreover, even when balancing the water budget with imported water, this does not stop the ABCSAL process if groundwater does not have exits (e.g., baseflow to streams or lateral subsurface outflow), and if water continues to leave the basin predominantly through evapotranspiration, which leaves behind salts. Although these latter two conditions are similar to those in a naturally closed basin (2) (Hardie and Eugster, 1970; Jones and Deocampo, 2003), vertical groundwater fluxes under ABCSAL are in the opposite direction from natural basin salinization and thus, the location of salinization is different. In a naturally closed basin, salinization occurs at the land surface due to upward groundwater discharge. Under ABCSAL, pumping and recharge from irrigation lead to a net downward flux, then mobilize salts left behind by irrigated crops downward into the production zone of the groundwater basin, before they are recycled

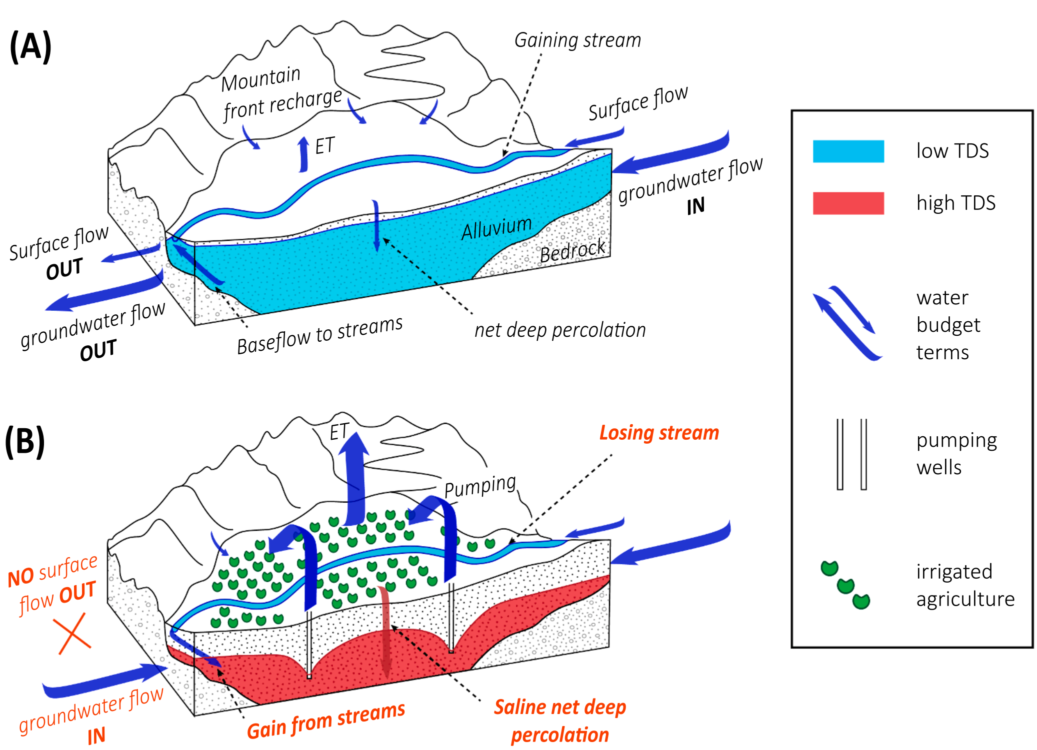

Fig. 1. Conceptual model of ABCSAL. (A) Open basin, pre-groundwater development: surface and groundwater systems connect. Groundwater discharges dissolved solids into surface water which exits the basin. Groundwater at this stage is predominantly fresh (e.g., <1,000 mg/L). (B) Closed basin: groundwater pumping causes elimination of baseflow to streams. Lower groundwater levels cause subsurface inflow to drain adjacent basins. Pumped groundwater is concentrated by evapotranspiration (ET) when applied for irrigation. Salts migrate into the production zone of the aquifer, driven by vertical hydraulic gradients from recharge and pumping. Although these figures showtwo extremes (open and closed), partially-closed basins also exist.

by pumping wells to the land surface and the process repeats.

Importantly, we point out that the long-term continuous decline of groundwater storage is not a necessary condition for ABCSAL. Rather, even in basins where groundwater levels are stable and hence assumed to be free of overdraft, as long as they remain physically closed, they will salinate. Furthermore, although for simplicity we describe basins as either “open” or “closed”, in reality, closure ranges from 0–100 % (i.e., fully open to fully closed), and gradations of basin closure exist, which impact the rate of salinization and hence, the long-term temporal and vertical spatial salt distribution. Except for the most extremely exploited aquifers (one of which we explore in this study), many aquifers will fall somewhere between fully open to fully closed and not exactly at one extreme.

In this research, we illustrate the development of ABCSAL in a historically open, freshwater basin using the agriculturally intensive Tulare Lake Basin (TLB) in California’s southern Central Valley as a case study. Previous research in the TLB has shown evidence of salt accumulation in groundwater via simple water and salt budgets (Schmidt, 1975), and shallow aquifer salt accumulation from sediment dissolution processes in highly-soluble calcium and magnesium carbonates and sulfates (Schoups et al., 2005). Other studies have shown that TDS concentrations in TLB groundwater have increased over the last century (Hansen et al., 2018; Lindsey and Johnson, 2018), and suggested this is the result of pumping for municipal and irrigation supply which has caused shallow, higher TDS groundwater to be driven downward into deeper aquifers. We are not aware of prior work that has placed these trends into the context of ABCSAL, or quantified potential rates of salinization across a range of aquifer depths and timescales.

Our aim in this study is to assess the first order salt balance and timescales over which the TLB as a large production aquifer system becomes regionally degraded over most of the vertical extent of its nearly 200 m thick main production zone. We conservatively assume that, under recent state regulation, groundwater overdraft is arrested, but not reversed. We compare timescales of ABCSAL degradation against the estimated lifespan of the greater Central Valley aquifer (i.e., 390 years at historical overdraft rates) (Faunt et al., 2009), challenge the notion that the depletion of groundwater storage is a more urgent issue than the degradation of groundwater quality in the TLB (and in other basins with ABCSAL conditions), and consider the water management implications and the steps required to reverse extensive basin-scale groundwater salinization. The management would likely involve both hydrologic opening of the basin to provide natural outlets for salt, a reduction of sources of salinity, and the development of regional groundwater quality management models (Fogg and LaBolle, 2006; CRWQCB, 2018). The adaptation might involve the eventual desalination of most groundwater pumped from the basin, producing a future economic burden that should be anticipated and evaluated, as it bears on the security of water, food, and energy resources.

This paper is organized as follows: first, we describe the hydrogeology, water budget, and water quality of the study site. Then we describe and justify our approach involving a simple 1D mixing cell solute transport model. Next, we present our results, and finally, we discuss the implications of the research, the limitations of our approach, and the extensibility of the study to other areas.

2. Methods

2.1. Study area

In selecting the TLB as our study site, we looked for (1) a history of intensive groundwater pumping and irrigation, (2) availability of historical water budget and water quality data, and (3) social and economic significance. The TLB (Fig. 2) occupies the southern third of the Central Valley, California and is bounded by the Coast Ranges to the west, the Tehachapi Mountains to the south, and the southern Sierra Nevada to the east. Geology strongly influences dissolved solid concentrations in

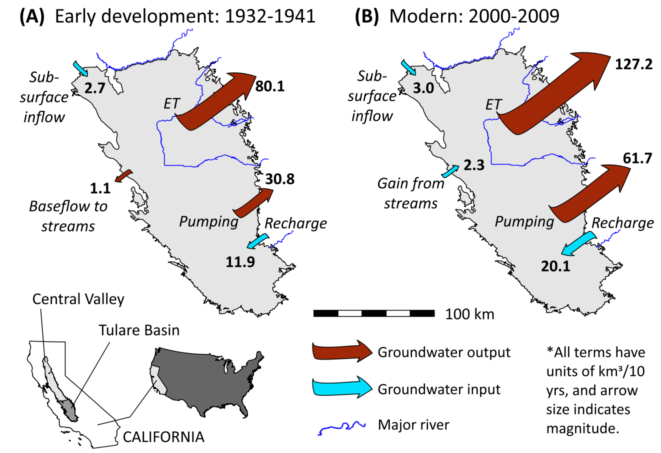

Fig. 2. The TLB overlies an agriculturally intensive sedimentary aquifer in California’s southern Central Valley. Significant changes are observed in selected decadal hydrologic year water budget terms derived from C2VSim at (A) early-groundwater-development (not to be confused with pre-groundwater-development) and (B) post-groundwater-development timescales in the TLB. Notably, gaining streams transition to losing streams, and increases are observed in pumping, evapotranspiration (ET), and recharge (from diversions and natural sources, like streams, lakes, and watersheds). All terms are aggregated at the scale of the TLB, except for subsurface inflow, which is calculated at the northern TLB boundary. Note that this is not the TLB groundwater budget (Table 1) nor the land surface and rootzone budget (Appendix Table A.2), but rather, a combination of ground and surface water budget terms that illustrate hydrologic change and show the main inputs (recharge) and outputs (pumping and evapotranspiration). Major rivers (shown in blue) from north to south include the San Joaquin, Kings, Kaweah, Tule, and Kern. Minor streams and tributarie.s are not shown. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

the clastic sedimentary aquifer system composed of fluvial and alluvial fan deposits. Calcium and magnesium sulfates and carbonates in Coast Range sediment in the western TLB are more soluble than sediments from the predominately crystalline rocks of the Sierra Nevada to the east, thus the groundwater in the western basin tends to have higher TDS (Fujii and Swain, 1995; Belitz and Heimes, 1990; Deverel and Millard, 1988). Fresh groundwater in the TLB spans depths from land surface to around 1,000 m where brackish water and marine deposits limit the development of groundwater resources (DeSimone et al., 2010; Kang and Jackson, 2016). Above this deep brackish zone is a major freshwater aquifer system. In combination with a natural endowment of significant, but intermittent runoff from surrounding uplands, abundant fresh groundwater has transformed the TLB into one of the most heavily irrigated and economically productive agricultural regions in the world (Hanak et al., 2011). At its peak in the 1980s, approximately 14,164 km^2 of its 44,110 km^2 were irrigated (TNC, 2014). Today roughly 12,140 km^2 remain irrigated, with a total gross value of all agricultural crops and products at $23.4 billion USD in 2017 (Fankhauser, 2018; Hook, 2018; L. Wright, 2018; M. Wright, 2017).

Although a TLB water budget from pre-development times is not available, the surface and subsurface hydrologic characteristics of the basin, which is a part of the larger Central Valley sedimentary basin (Fig. 2), indicate that it was hydrologically open. We first discuss the surface hydrologic aspects. Despite the shallow topographic depression in which Tulare Lake used to exist, the freshwater lake periodically filled up and overflowed northward into the San Joaquin River (Grunsky, 1898; Davis et al., 1959), providing an outlet for any accumulated salts. Reconstructions of historical Tulare Lake level indicate that in 19 of the 29 years from 1850 to 1878, it filled up and flowed out of the basin to the north (USBR, 1970). This water and salt exit via intermittent surface inundation would be different than, say, baseflow to a stream, but would accomplish the same flushing function. No overflows are documented after 1878 due to the diversion of tributary waters for agricultural irrigation and municipal water use (ECORP, 2007).

The subsurface characteristics also indicate open hydrologic conditions. There is significant evidence that groundwater flowed northward into the adajacent San Joaquin Basin in pre-development times (circa early 1900s). This evidence includes (1) historical measurements of Central Valley groundwater TDS showing lowest TDS values in the TLB, with increasing TDS to the north into the San Joaquin Basin (Mendenhall et al., 1916, Table 23), consistent with northward groundwater flow and the accompanying down-hydraulic-gradient groundwater chemistry evolution that is routinely observed in sedimentary basins, e.g., Palmer and Cherry (1984); (2) the regional, south-to-north topographic gradient to provide the driving force for gravity-driven flow in the same direction, out of the TLB, even if there existed shallower, local groundwater flow components from north to south at the subtle depression that collected Tulare Lake (e.g., refer to classic work of Tóth (1970) on topographically controlled, gravity-driven flow systems); and (3) horizontal stratification of fine- and coarse-textured sediments in the Central Valley sedimentary basin that results in much lower effective hydraulic conductivities in the vertical direction than the horizontal e.g., Weissmann et al. (2002) and Faunt et al. (2009), thereby minimizing influence of subtle topographic features like the Tulare Lake depression on all but the shallowest groundwater flow components (e.g., refer to Tóth (1970) and related work).

Summarizing, our conceptual model of the pre-development TLB hydrologic system is one in which the subtle topographic depression that collected the typically 12 m deep Tulare Lake (Preston, 1990), together with the periodic overflow of the lake and discharge to the north, resulted in a partly open surface drainage system. Further, the larger topographic and geologic structure of the basin, together with groundwater chemistry evidence, indicates there was net-northward groundwater flow, making the TLB groundwater system an open hydrologic basin in pre-development times.

Parts of TLB may have been salinating to some degree before development due to shallow evaporation of groundwater and surface water (case (3) in Introduction), in contrast to the ABCSAL process that we describe in this paper. Portions of the TLB closed under pre-development conditions would lead to salt accumulation in and near its playas (e.g., Buena Vista Lake, Tulare Lake): an evaporative boundary of the basin and endpoint to all surface water discharge (case (2) above). This is consistent with observations of high salinity near and in these lakebeds (Hansen et al., 2018; Fujii and Swain, 1995). Although there exist local areas of shallow groundwater with elevated salinity on the west side of the TLB, these areas are typically associated with salt mobilization out of alluvial sediments originating from marine sedimentary source rocks in the Coast Ranges, and not from basin closure.

By the time regional groundwater levels were mapped in the early twentieth century, the TLB showed signs of closure: groundwater flow across the northern boundary was minimal, and flowed north to south, into the TLB (Mendenhall et al., 1916; Ingerson, 1941). Although pre-groundwater-development (pre-1850) water budgets are unavailable, two large-scale, regional groundwater flow models of the Central Valley (Brush et al., 2013; Faunt et al., 2009) provide decadal groundwater budgets for early- (1932–1941) and post-groundwater-development (2000–2009) timescales.

Relative to the decadal hydrologic water year budgets of early-groundwater-development, post-groundwater-development water budgets show much higher pumping, crop evapotranspiration, and recharge (Brush et al., 2013). As groundwater levels fell, gaining streams transitioned to losing streams, and subsurface inflow along the northern basin boundary slightly increased (Fig. 2). Groundwater discharge to surface water almost entirely ceased. Surface water exits the basin in rare years when the Kings, Kaweah, and Kern rivers produce sufficiently large floods, mostly runoff from the surrounding uplands. Evapotranspiration from irrigated crops has become the dominant water outflow, and this flow is much greater than it was during early-groundwater-development (Brush et al., 2013). Taken together, these hydrologic changes have transitioned the TLB into an anthropogenically closed groundwater system with commensurate onset of ABCSAL.

2.2. Mixing cell model development

Given the large space and time scales of interest, and the large-scale effectively one-dimensional vertical flow conditions in the basin due to pumping and recharge, we used a lumped parameter approach based on upscaling water fluxes of a fully three-dimensional groundwater model. Although local hydrogeologic conditions vary and can lead to locally complex three-dimensional flow and transport, our focus here is on large scale salinization behavior and time scales, thus an upscaled model was appropriately parsimonious. Moreover, upscaling the advection dispersion equation to regional scales remains a scientific and computing challenge (Guo et al., 2020; Guo et al., 2019) beyond the scope of this study.

Mixing cell models, also called discrete-state compartment models, are computationally inexpensive and have successfully been used in place of complex flow models to provide rapid, first-order estimates of water budgets, mass flux, and contaminant concentrations (Campana, 1975; Campana and Simpson, 1984; Campana, 1987; Carroll et al., 2008; Kirk and Campana, 1990; TC, 1982). A mixing cell approach segments the system into a set of control volumes. In each iteration the incoming water displaces an equivalent volume of water, then mixes with the remaining cell contents, and new concentrations are calculated at each cell. Specifically, we use the "modified mixing cell model" consistent with Fick's Law (TC, 1982). Here, we represent the TLB groundwater system through a one-dimensional, vertical column of discrete control volumes (cells), given the predominance of vertical downward flow at the aquifer system scale. We assume that each cell consists of a fraction f of sediments participating in groundwater flow and salt transport with porosity η. We neglect flows and rock-water interactions in sediments not participating in transport, of proportion $1-f$ (more details below). The thickness of each cell is chosen such that the advective travel time (Δt) of water and salt downward through each cell is exactly 50 years (synchronized tipping bucket model, see Eq. 4) below, thus full mixing occurs at each cell even as the groundwater flow velocity decreases with depth. To determine the mixing cell parameters, water fluxes throughout the vertical domain (e.g., recharge, vertical flow rate, pumping) are obtained by averaging (i.e., mass-conservative upscaling) the TLB portion of a fully three-dimensional, heterogeneous groundwater flow model of the Central Valley (Brush et al., 2013).

The salt accumulation in a mixing cell at a discrete time k is a mass balance of the initial mass (m_k) $[M]$, incoming mass (m_k^{in}) and exiting mass (m_k^{out}).

$$m_{k+1} = m_k + m_k^{in} - m_k^{out} \tag{1}$$

Input and output mass terms can be calculated for each term in the water and salt budget (Table 1), from their input and output concentration (C_k^{in}, C_k^{out} $[ML^{-3}]$) and input and output volumetric flow (Q_k^{in}, Q_k^{out} $[L^3]$):

$$m_k^{in} = C_k^{in} Q_k^{in} \; ; \; m_k^{out} = C_k^{out} Q_k^{out} \tag{2}$$

Finally, the concentration in a mixing cell at time step k is:

$$C_{k+1} = \frac{m_k + m_k^{in} - m_k^{out} + \rho V}{V f \eta} \tag{3}$$

where V $[L^3]$ is the total cell volume, f $[-]$ is the fraction of sediments actively participating in groundwater flow and salt transport, η $[-]$ is the porosity of those sediments, and ρ $[ML^{-3}]$ is rock-water interaction coefficient. The fraction f is found to be 0.99 (Brush et al., 2013), which in the C2VSim model includes all textures but the Corcoran clay, a relatively impermeable clay layer comprising around 1% of the model

Table 1
Average annual groundwater and salt budget for the TLB (Eq. (5)) from C2VSim (1961-10-31 to 2001-09-30), and the modified no-overdraft budget used in this analysis (Eq. (7)).

	Source	Q $((km)^3/yr)$	C (mg/L)	m (kt/yr)
Historical budget	R	2.451	32.5	80.3
	B	0.236	32.5	7.5
	C	0.572	32.5	18.5
	I	0.011	32.5	0.3
	P	−6.761	*	*
	N	1.883	*	*
	RWI	–	–	*
	ΔS	−1.608	–	–
Alternate budget	R	2.451	32.5	80.3
	B	0.236	32.5	7.5
	C_{alt}	0	–	–
	M	0.678	32.5	22.0
	I	0.011	32.5	0.3
	P_{alt}	−5.259	*	*
	N	1.883	*	*
	RWI	–	–	*
	ΔS_{alt}	0	–	–

* non-constant term calculated at each time step
Q is the volumetric flow rate, C is the concentration of TDS, and m is the mass flux of salt (where t represents "tonne" which is 1000 kg). Groundwater budget terms are: R = recharge from streams, lakes, and watersheds, B = lateral mountain front recharge from streams and watersheds, C = subsidence flow, C_{alt} = subsidence flow to eliminate overdraft (along with M_{alt} and P_{alt}), M = managed aquifer recharge to eliminate overdraft (along with C_{alt} and P_{alt}), I = subsurface inflow from the north, P = groundwater pumping, P_{alt} = alternate groundwater pumping to eliminate overdraft (along with M and C_{alt}), N = net deep percolation (recharge from the land surface through vadose zone and into saturated groundwater), RWI are rock-water interactions. ΔS = change in groundwater storage. ΔS_{alt} = change in groundwater storage for the modified budget. The modified budget eliminates overdraft by reducing P to P_{alt} according to Eq. (12), and introducing recharge M.

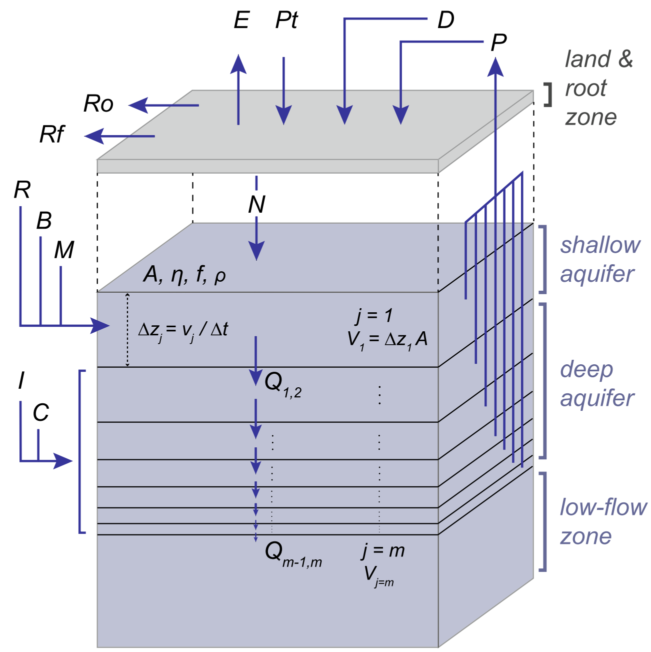

Fig. 3. Conceptual land-root zone model and groundwater mixing cell model with surface area A, porosity η, aquifer fraction f, rock-water interaction coefficient ρ, and m cells. The cell thickness Δz_j is given per Eq. (4), where average linear velocity $v_j = q_j/f\eta$. The cell volume V_j is the total bulk volume of the rock including aquifer and non-aquifer material. The TDS in cell j is calculated by Eq. (3). The land and root budget (Appendix Table A.2) accounts for pumping (P), surface water diversions (D), precipitation (Pt), evapotranspiration (E), runoff (Ro), return flow (Rf), and net deep percolation (N). N enters the top of the groundwater mixing cell model along with recharge from streams, lakes, and watersheds (R), boundary inflow from mountain front recharge (B), and managed aquifer recharge (M). Internal flows from subsurface inflow from the north (I), subsidence flow (C), and pumping (P) are distributed proportional to cell volume, e.g., Eq. (12). The average annual groundwater and salt budget is reported in Table 1.

volume. Porosity, η, is set to 0.40, the average for the TLB. Coarse and fine sediment porosities do not appreciably differ, averaging around 0.40 with an interquartile range of 0.39–0.41 for all textures, as demonstrated in abundant core analyses (Johnson et al., 1968), and discussed further in Appendix Table A.5 and Fig. A.9; hence, we did not consider varying η across aquifer layers.

To account for mass contribution from natural dissolution of geologic minerals, we define a zero order source term called the rock-water interaction coefficient ρ $[ML^{-3}]$. Rock dissolution along groundwater flow paths is well documented in sedimentary aquifers (Palmer and Cherry, 1984; Oetting et al., 1996; Tóth, 1999; Mahlknecht et al., 2004; Cloutier et al., 2008). We obtain a representative mass dissolution rate from the slope of a representative TDS profile for the TLB from land surface to the base of fresh water (Williamson et al., 1989; Kang and Jackson, 2016). The product of the rock-water interaction coefficient ρ and the cell volume (V) is the additional mass accumulated from rock-water interactions in the cell. We also evaluate an alternative scenario with $\rho = 0$.

We solve (3) sequentially over the stacked mixing cells from top to bottom and across seven 50-year time steps from 1960 (initial condition) to 2310 (synchronized tipping bucket approach) to obtain the variation of salinity with depth and time.

The discretization, Δz_j, of the stacked series of mixing cells (Fig. 3) is driven by the time step, $\Delta t = 50$ years, and the representative basin-scale vertical Darcy velocity, q_j, within the j^{th} mixing cell:

$$\Delta z_j = \frac{q_j}{f\eta} \cdot \Delta t \tag{4}$$

Since q_j is depth dependent, we solve (4) sequentially for $j = 1 \ldots m$, beginning at the water table to compute the vertical discretization of the stacked mixing cell model. Here, we assume that the inflow into a mixing cell, $q_{j-1,j}$ is representative of the flow rate q_j throughout the cell. Thus – to compute cell thicknesses with Eq. 4 – the pumping, P_j, lateral basin flow I_j, or subsidence flow C_j (Fig. 3) conceptually flow into or out of the mixing cell bottom. The following sections provide further details on the parametrization of (3) and (4).

2.3. Boundary conditions, model parameters, and stochastic simulation

Initial conditions, boundary conditions, and model parameters are informed by the C2VSim groundwater flow model developed by the California Department of Water Resources (Brush et al., 2013), publicly available water quality data (CSWRCB, 2019c), and previous field studies of the TLB. The following describes methods used to determine (1) water and salt budgets, (2) salt fluxes from evaporative concentration and pumped groundwater, (3) the groundwater velocity-depth profile, (4) the initial TDS-depth profile, and (5) spatial parameters and aquifer properties. Lastly, we discuss the simulation timescale and the role of stochastic simulation.

2.3.1. Water and salt budgets

The water budget is based on C2VSim version 3.02, a 3 layer and 1,392 element, regional scale, finite-element groundwater flow model of California's Central Valley alluvial aquifer system (Brush et al., 2013). C2VSim is an application of the Integrated Water Flow Model (IWFM) (Dogrul et al., 2018), a water resources management and planning model that simulates surface water, stream-groundwater interaction, vadose zone flow, and groundwater flow. In the C2VSim model, California's Central Valley aquifer is separated into 21 subregions, and detailed land surface, root zone, and groundwater budgets for each subregion are calculated at monthly time steps from the 1923 to 2009 hydrologic years. The TLB is represented by subregions 14–21. Because of its detailed representation of surface–groundwater interaction, groundwater pumping, three-dimensional aquifer structure, representation of significant land subsidence in the study site, and calibration, C2VSim was chosen as a reasonable representation of the TLB water budgets, groundwater velocities, and thus chosen to develop the mixing cell model.

The C2VSim model was run for the 40-year period from 1961-10-31 to 2001-09-30 to obtain an average annual TLB groundwater budget (an equivalent average annual landscape/root zone budget is provided in Appendix Table A.2). This post-groundwater development water management time frame is characterized by pumping and overdraft, in addition to wet, dry, above normal, below normal, and critical water year types. The C2VSim change in groundwater storage is defined as:

$$\Delta S = R + B + C + I + N - P \tag{5}$$

where ΔS is change in groundwater storage $[L^3]$, R is basin recharge from streams, lakes, and watersheds $[L^3]$, B is lateral mountain front recharge from streams and watersheds $[L^3]$, C is subsidence based flow from clay compaction $[L^3]$, I is subsurface inflow from the north $[L^3]$, N is net deep percolation predominately from irrigation water $[L^3]$, and P is groundwater pumping $[L^3]$. The dominant budget terms are P, R, and N (Table 1).

To demonstrate ABCSAL under long-term conditions that avoid further overdraft (but not basin closure), we solve the mixing cell model Eqs. (30 and (4) alternatively for $\Delta S_{alt} = \Delta C_{alt} = 0$. Overdraft is eliminated with an alternate budget (Table 1), which adds managed aquifer recharge, m as inflow to the top mixing cell (Fig. 3), and reduces pumping to an alternative pumping level, P_{alt}. We add $M = 0.68\ km^3$, which was determined by a prior study as the maximum theoretical recharge available to the San Joaquin Valley (which includes the TLB), assuming unlimited infrastructure and water transfer ability (Hanak et al., 2019). Eliminating overdraft in this way effectively maintains a steady-state, saturated model that remains closed to due to lack of baseflow and groundwater outflow. Hence, the water level is immobile, but the salt front can move, thus simulating salt migration without drying out cells due to overdraft.

Since M represents captured surface water flow, we assign it the same TDS as natural water (32.5 mg/L), discussed below. We also simulated M with a TDS of 0 mg/L (Appendix Table A.9) and found that it had a negligible impact on resulting salt concentrations presented in this study (Appendix Table A.8).

The alternate, reduced pumping P_{alt}, is computed by rearranging (5), adding M, and setting $\Delta S_{alt} = C_{alt} = 0$:

$$P_{alt} = R + B + M + I + N \tag{6}$$

Therefore, the modified no-overdraft alternate groundwater budget is:

$$\Delta S_{alt} = R + B + C_{alt} + M + I + N - P_{alt} = 0 \tag{7}$$

The salt budget is calculated by assigning a TDS concentration to each term in the groundwater budget (7). TDS for natural waters (e.g., stream, lake, and managed aquifer recharge budget terms) were determined to be 32.5 mg/L, by computing the median of the sampling distribution of sample TDS medians in TLB stream samples (USGS, 2016) from 1951–2019 (Appendix Fig. A.7 and Table A.3). Similarly, the TDS of diverted surface water was calculated to be 264.5 mg/L, as the average annual water and salt budget from 1985–1994 of two major surface water conveyance structures, the California State Water Project and the State Water Project (Cismowski et al., 2006) (Appendix Table A.3). Salt and water budgets are detailed in Table 1.

2.3.2. Velocity-depth profile

To explicitly solve for the mixing cell discretization (4), we fit a linear model to the C2VSim vertical Darcy velocities, reported for each finite element cell in the three layer C2VSim grid at the layer-to-layer boundaries. Due to increases in recharge and pumping caused by groundwater development and irrigation, the groundwater flow system is vertically dominant, and thus supports the application of a 1D, vertically oriented model. To account for groundwater velocity change in the alternate groundwater budget (7), groundwater velocity is scaled proportional to the decrease in vertical volumetric flow rate, $P_{alt}/(P+C) = 0.85$ (a 15 % reduction). This is equivalent to the ratio of net downward volumetric flow in the alternate budget to the net downward volumetric flow in the historical budget (Table 1).

$$q(z) = (\beta_0 + \beta_1 z)\cdot\frac{P_{alt}}{P + C} \tag{8}$$

where β_0 and β_1 are the regression coefficients (Appendix Table A.4), and the overall change (reduction) in velocity is -15%. Mixing cell thickness (4) is determined by computing q_j from (8) for the depth, z, of the bottom of the mixing cell $j-1$ (top of cell j). To ensure consistency between the water balance terms in (5) and the approximated vertical velocity profile (8), we compute the water mass balance error, $MB_{error,j}$, for each mixing cell j:

$$MB_{error,j} = q_{j-1,j} + I_j - P_{alt,j} - q_{j,j+1} \tag{9}$$

For the uppermost mixing cell $j = 1$, we rearrange (9), replacing $q_{j-1,j}$ for the sum of N, R and B, and ignoring subsurface inflow I_j (Fig. 3):

$$MB_{error,1} = N + R + B + M - P_{alt,1} - q_{1,2} \tag{10}$$

The cell by cell budget and mass balance errors (which are effectively zero, and equivalent to the cell-by-cell change in storage) are reported in Appendix Table A.7.

2.3.3. Evapoconcentration and pumping

Evapotranspiration removes a majority of total applied water, leaving behind dissolved solids in the crop rootzone that eventually migrate into groundwater. We model the evapoconcentration of TDS in total applied water (a combination of pumped groundwater and imported surface water diversions) by accounting for the application efficiency (Burt et al., 1997), and thus the fraction of water that remains after evapotranspiration:

$$C_N = \left(\frac{m_D + m_P}{V_D + V_P}\cdot\frac{1}{1 - E_a}\right) = \frac{C_{D,P}}{1 - E_a} \tag{11}$$

C_N is the concentration of net deep percolation after accounting for evapotranspiration. m_D and m_P are the mass, and V_D and V_P are the volume of surface water diversions (D) and pumping (P), respectively. $C_{D,P}$ is the concentration of total applied water from surface water diversions and pumping (calculated by mixing diversions and pumped groundwater in their respective proportions, see Appendix Table A.4), and E_a is the application efficiency, which has a measured regional average of 0.78 in the Tulare Basin (Sandoval-Solis et al., 2013), and agrees with measured values in hydrologically similar areas (Hanson et al., 1995; Howell, 2003). Alternatively, the C2VSim landscape/soil water budget (Appendix Table A.2) provides an application efficiency, E_a, of 0.88 when considering the amount of water infiltrating into the soil and deep percolation. For sensitivity analysis, we run simulations for several E_a between 0.78 and 0.88 to further explore model outcome uncertainty.

For the stacked mixing cell model, we assume that P_{alt} in the no-overdraft groundwater budget (6) is distributed uniformly with depth, from the water table to the last mixing cell m. Similarly, we assume lateral inflow I is uniformly distributed across depth, from cell 2 to cell

m. Therefore, pumping is proportional to mixing cell thickness, and the salt mass flux due to pumping during time step k in mixing cell j is:

$$m_{j,k} = \frac{V_j f\eta}{f\eta \sum_{i=1}^{n} V_i} PC_{j,k} \tag{12}$$

Noting that the $f\eta$ term drops out, and summing over all mixing cells at time k gives the total mass flux from groundwater pumping ($m_{P,k}$):

$$m_{P,k} = \sum_{j=1}^{n} \frac{V_j}{\sum_{i=1}^{n} V_i} PC_{j,k} \tag{13}$$

2.3.4. Initial TDS-depth profile

The initial TDS-depth profile is determined by fitting a linear model to the pre-1960 TDS-depth measurements (Fig. 4) (CSWRCB, 2019c). Due to the influence of freshwater recharge at the land surface and rock-water interactions, pre-1960 TDS generally increases with depth, consistent with observations of increasing TDS with depth in the region (Kang and Jackson, 2016; Kharaka and Thordsen, 1992; DeSimone et al., 2010).

2.3.5. Ensemble simulation

We assign a uniform probability distribution to the parameters of which we are least certain and discrete values to those that are measured (Appendix Table A.6), then perform Monte Carlo simulation to generate an ensemble output. The mixing cell model is evaluated 1,000 times – which the computational simplicity of a lumped model permits; modeling uncertainty in this way with a distributed parameter, 3D flow and transport model would be computationally prohibitive. Parameter ranges are estimated from literature for rock-water interaction coefficient (Williamson et al., 1989; Kang and Jackson, 2016), detailed in Section 2.2. As described in Section 2.3.3, application efficiency is both measured (Sandoval-Solis et al., 2013), and calculated from C2VSim (Brush et al., 2013).

To show the influence of rock-water interactions on the progression of closed basin salinization, we simulate two basic scenarios:

1. No rock-water interactions: mass accumulates from water budget inputs.
2. Rock-water interactions are present: mass accumulates from water budget inputs, but also internally via rock-water interactions (see Section 2.2 for details).

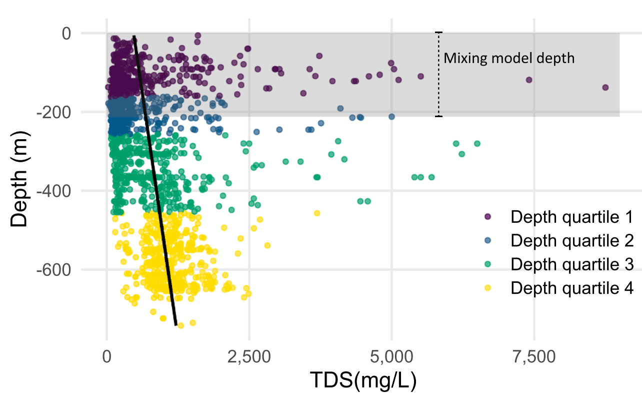

Fig. 4. Pre-1960 groundwater quality generally decreases with depth, reaching an average concentration of 1,000 mg/L at 526 m deep. The initial TDS-depth concentration at $t = 0$ is approximated by a linear model, shown as a black line. The transparent, grey rectangle shows the depth of the mixing cell model (212 m).

3. Results

3.1. Groundwater and salt budget

The average historical C2VSim groundwater budget in the TLB from 1961-10-31 to 2001-09-30 (Table 1) reflects post-groundwater development conditions. Pumping removes an average of $-6.76\ km^3/yr$ from the groundwater system. Natural recharge from streams, lakes, and watersheds adds an average of 2.45 km^3/yr, and net deep percolation of agricultural irrigation adds an average of 1.89 km^3/yr. Smaller sources of water inflow include subsidence flow (0.57 km^3/yr), lateral mountain front recharge from streams and watersheds (0.24 km^3/yr), and subsurface inflow from the north (0.01 km^3/yr).

The alternate budget (Table 1) used in this study eliminates overdraft ($\Delta S = 0$), and is identical to historical budget described above, except that pumping P_{alt} is reduced to $-5.26\ km^3/yr$, managed aquifer recharge M is added at a rate of 0.68 km^3/yr, and subsidence flow C_{alt} is reduced to 0. Importantly, in this alternative budget the basin remains closed.

Salt inputs to the system (Fig. 5A) come from pumped groundwater, water budget terms, and rock-water interactions.

Groundwater pumping for agriculture is unlike other water budget terms (I, M, R, B) and rock-water interactions in that it does not *add* new salt into the system, but rather *recycles* existing salt from deeper layers to the land surface and back into shallow groundwater via irrigation (discussed in Section 3.2). In the no rock-water interactions scenario ($\rho = 0$), the median mass recycled by pumped groundwater exceeds the mass input of all other water budget terms by a factor of 2.0 to 3.9 depending on the timeframe considered. When rock-water interactions are present ($\rho > 0$), they initially contribute a comparable mass to groundwater pumping (around 4 Mt/yr), but with time, salt accumulates in the aquifer, and the mass recycled by groundwater pumping exceeds the mass imparted by rock-water interactions (Fig. 5A).

Annually, surface water diversions add 1.5 Mt/yr of salt to the study site. This is around 13 times the amount of all other non-pumping water budget terms combined (I, M, R, B), which add only 0.11 Mt/yr. We estimate that rock-water interactions add between 3.3 and 4.6 Mt/yr of salt. This exceeds the mass introduced by imported surface water and is comparable to the mass recycled by groundwater pumping.

Due to the closed-basin hydrology of the study site, there are no exits for salt to leave the system. Instead, pumping and irrigation recycle salts within the basin, and evapotranspiration by crops at the land surface increases the concentration of net deep percolation, which recharges groundwater (Fig. 5B).

Evapoconcentration by crops at the land surface increases the average concentration of total applied water (pumped groundwater combined with surface water diversions) by 5.1–6.8 times its original amount, regardless of whether rock-water interactions are absent or present. As previously discussed, since pumped groundwater concentration increases with time, total applied water and thus net deep percolation also become increasingly saline over time.

3.2. Progression of groundwater salinization

The shallow aquifer (36 m) is heavily impacted by the recycling of salts via pumping and irrigation, and exceeds the freshwater concentration threshold (1,000 mg/L) within decadal timescales (Fig. 6). Intermediate (132 m) and deep aquifers (187 m) exceed 1,000 mg/L within century-long timescales.

Uncertainty in the salt balance results from parameter uncertainty expressed in the Monte Carlo simulation (Section 2.3.5), which affects the distribution of calculated salt concentrations at the salt front. Deeper layer insensitivity results from being insulated from the salt front – a top down source. Accordingly, shallow layer uncertainty increases over time because salt is continuously added through top-down irrigation and recharge.

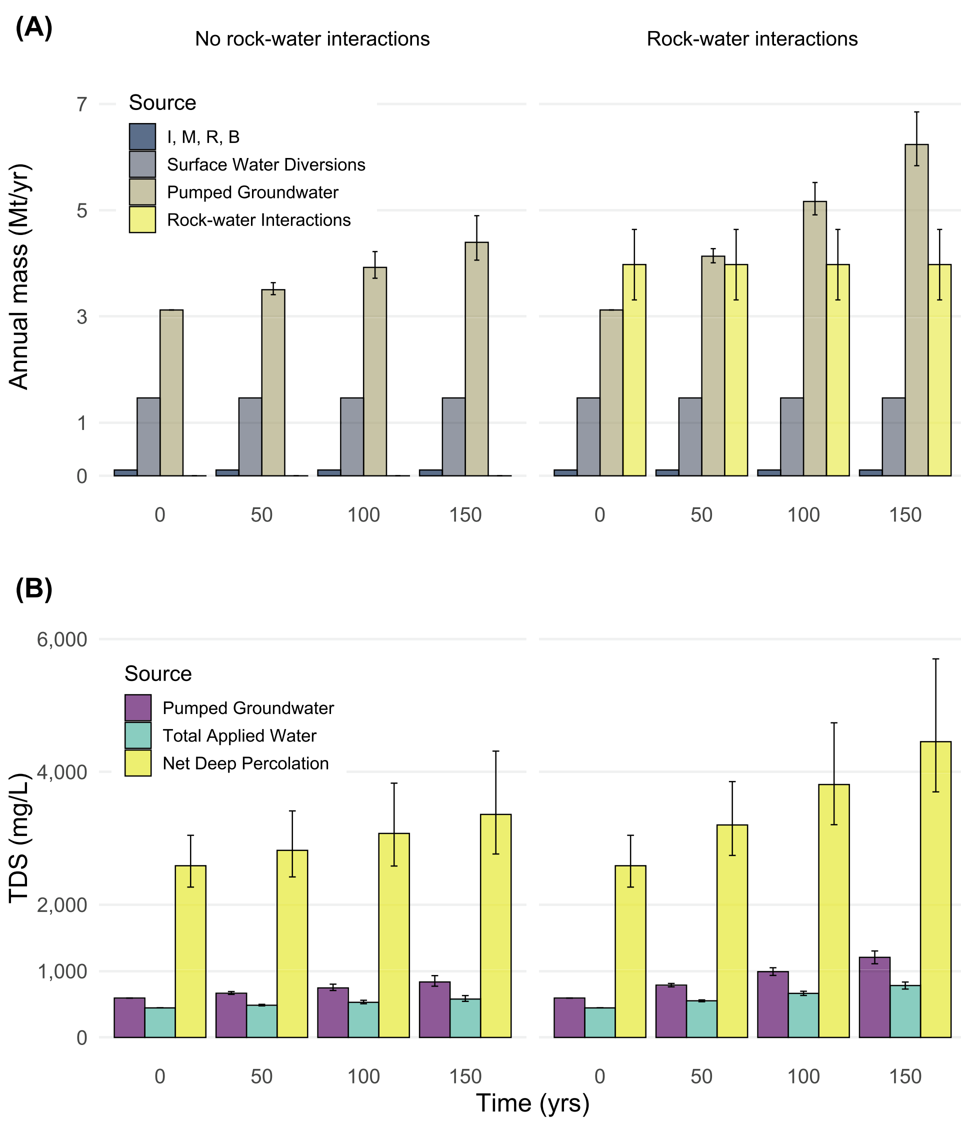

Fig. 5. Annual mass flux and TDS of selected budget terms. The height of each column is the ensemble median result, and the width of error bars, if present, is the interquartile range of the ensemble distribution. (A) Pumped groundwater contributes more mass than surface water diversions and all other water budget terms combined (represented by their symbol: I, W, R, B). (B) TDS of *pumped groundwater* is diluted when mixed with imported surface water, which forms *total applied water*. However, evapotranspiration concentrates total applied water, which enters the groundwater system as *net deep percolation*. Over time in a closed basin system, the groundwater salinates, which in turn increases the concentration of total applied water and net deep percolation.

Let us first summarize the results with no rock-water interactions. At the beginning of the simulation (year 1960), initial TDS concentration increases gradually with depth (Fig. 4 and Appendix Table A.8). Shallow aquifer salinity is 506 $\mathrm{mg/L}$. After 50 *yrs* with $\rho = 0$, average shallow aquifer salinity reaches a median concentration of 934 $\mathrm{mg/L}$ with an interquartile range (IQR) of 829–1,083 $\mathrm{mg/L}$. Thus, the TDS-depth profile at $t = 50$ begins to invert (i.e., shallow aquifer salinity exceeds deep aquifer salinity), consistent with modern-day observed TDS-depth relationships in the TLB (Hansen et al., 2018). After 200 *yrs* (year 2160), shallow aquifers reach brackish TDS levels with a median TDS of 1,241 $\mathrm{mg/L}$ (IQR: 1,031–1,576 $\mathrm{mg/L}$). Finally, after 300 *yrs* (year 2310), median shallow aquifer TDS approaches nearly 1,477 $\mathrm{mg/L}$ (IQR: 1,175–1,993 $\mathrm{mg/L}$).

Intermediate and deep aquifers are impacted much later than shallow systems, and approach the freshwater TDS threshold on timescales of two to three centuries. After 200 *yrs* (year 2160), intermediate aquifer median TDS is 907 $\mathrm{mg/L}$ (IQR: 830–1,017 $\mathrm{mg/L}$). After 300 *yrs* (year 2260), deep aquifers (IQR: 841–995 $\mathrm{mg/L}$) experience the first arrival of the lumped salt front.

In the "rock-water interactions present" scenario ($\rho > 0$), the progression of groundwater salinization follows approximately the same trend and timescale as the scenario without rock-water interactions (described above), but the resulting concentrations are significantly greater, and deep groundwater salinates faster. In both scenarios, the greatest change in salinity occurs in the shallow aquifer within the first 50 *yrs*, which is due to the introduction of mass from total applied water (i.e., diversions and pumped groundwater), and the inability for that mass to exit because of basin closure. Moreover, regardless of whether rock-water interactions are included, the slope of the TDS-depth profile (Fig. 6) gradually inverts and amplifies, and shallow groundwater becomes saltier than deep groundwater. Thus, even in the absence of rock-water interactions, moderate and constant salt inputs (mostly due to recycled groundwater and imported surface water) are sufficient to salinate shallow aquifers within decades, and deep aquifers within centuries.

3.3. Additional perspective on the model

Lumped mixing cell models have a relatively small number of parameters, are computationally inexpensive, conceptually simple, and importantly, can represent the dominant hydrologic features of a system. These strengths come with some tradeoffs. Mixing cell models can be used to simplify groundwater flow and contaminant transport by ignoring horizontal flow, geologic heterogeneity, dispersion, diffusion, sorption, and reactive transport. Strong vertical hydraulic gradients induced by pumping in agriculturally dominant systems (like the TLB),

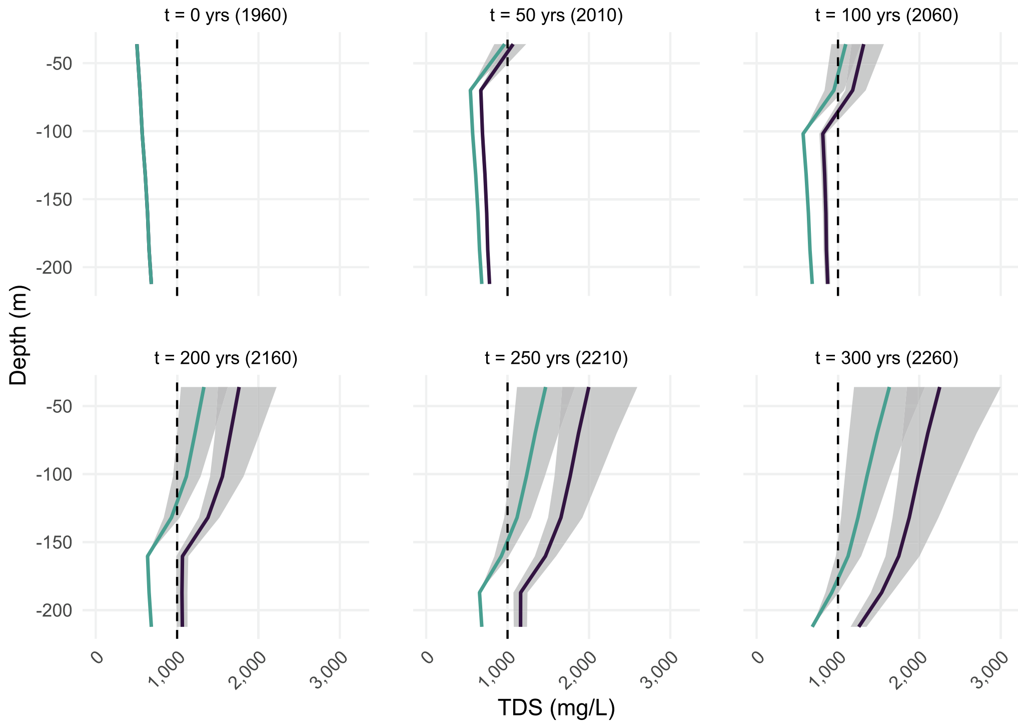

Fig. 6. Progression of groundwater salinization ensemble results for two scenarios (with and without rock-water interactions). RWI stands for rock-water interactions. The blue and purple lines show the ensemble median concentration for the two scenarios, and the interquartile range (IQR) of the ensemble simulations is shown as a grey shaded area. Complete statistics are provided in Appendix Table A.8.

produce vertically dominated flow systems (Brush et al., 2013; Faunt et al., 2009). In upscaling these distributed models to the regional scale, the dominant role of vertical flux becomes apparent and explains why the mixing cell model captures the salient features of regional ABCSAL degradation. For more sub-regional or local applications, a fully three-dimensional distributed parameter model representing the effects of preferential flow and tailing on solute transport would be more appropriate (Zhang et al., 2006; Guo et al., 2019; Guo et al., 2020; Henri and Harter, 2019).

Additionally, we assume that the early-groundwater-development TDS-depth relationship is approximately equal to observed pre-1960 TDS data. Over the model domain (212 m deep), these measurements (Appendix Fig. A.8) are well distributed. We experimented with different values for the initial TDS-depth profile, and found that the results were relatively insensitive to the initial conditions, as the imported salt and the salt generated by rock-water interactions greatly exceeds the initial salt load.

Moreover, in this study we model TDS as a lumped term, yet it should be noted that TDS is a combination of many solutes which differ in their impact to crops, toxicity, and reactivity with the subsurface. In the TLB, the dominant salts include cations and anions from geochemical weathering of multi-mineralic, clastic sedimentary deposits containing carbonate as well as silicate minerals (e.g., calcium, magnesium, sodium, carbonate) (e.g., sodium, calcium, bicarbonate) (Schoups et al., 2005; Hansen et al., 2018), but in other basins, this may not be the case. Therefore, the impact to fresh groundwater in other basins depend on the types and relative abundances of solutes present. Nevertheless, the geochemical make up of sedimentary particles in the Central Valley of California is typical of that found in many other sedimentary basins.

Lastly, in our TLB study site, historical groundwater pumping has reduced groundwater levels such that basin outflow even in rare wet years is essentially negligible, hence, we found it appropriate and parsimonious to use an average water budget that maintains constant hydrologic basin closure over time. However, other basins, both open and closed, will exhibit seasonal, annual, and decadal hydrologic variability. For instance, extreme rainfall and net deep percolation can temporarily induce basin outflows that export some of the accumulated salts, the degree of which is a function of the salinity of the source(s) responsible for basin discharge (e.g., surface water, shallow or deep groundwater). These cycles are not incorporated in the model presented, but in sites where they are important, they should be included.

4. Discussion

4.1. ABCSAL threatens regional groundwater quality and sustainable yield

In this study we show that ABCSAL is a progressive, regional-scale hydrologic process where salts accumulate within an aquifer because basin closure eliminates exits for the salts. In the TLB, our calculated ABCSAL timescales have similar timescales to aquifer depletion, are consistent with 3D random walk salt transport simulations, and agree with observed decadal changes in shallow groundwater salinity in the TLB.

Our estimates of decadal timescales for shallow aquifer (36 m) salinization, and two to three centuries for intermediate (132 m) and deep aquifers (187 m) are similar to the estimated 390 year timescale of Central Valley aquifer depletion by Scanlon et al. (2012), who assumed a remaining water storage of 860 km^3 in the year 2000, and a depletion rate of 2.2 km^3/yr. Scanlon et al. (2012) also noted that aquifer lifespan is likely shorter than 390 years in the TLB due to focused groundwater depletion in the area. Thus, ABCSAL, which constitutes a slow-moving form of regional groundwater quality degradation may significantly constrain groundwater sustainable yield on similar timescales to aquifer depletion in the TLB.

This study's predicted salinization time frames (i.e., decades for shallow systems, centuries for deep systems) are consistent with random walk salt and nitrate particle transport simulations in detailed 3D heterogeneous alluvial aquifers (Henri and Harter, 2019; Zhang et al., 2006), which suggests that the simple mixing cell model captures key transport dynamics. Thus, these results provide a useful basis for future research using more complex, distributed parameter, regional-scale transport models incorporating geologic heterogeneity and transient boundary conditions.

Moreover, measured TDS change from historic to modern time periods in the TLB agree with this study's modeled changes in TDS over similar time periods and horizontally averaged depth scales. Hansen et al. (2018) measured a 110–850 mg/L interquartile range (IQR) increase in shallow aquifer (<50 m) TDS from historic (1910) to modern (1993–2015) time periods in the TLB. Our results indicate an IQR increase in shallow aquifer (<37 m) TDS of 323–717 mg/L depending on the inclusion of rock-water interactions (ρ in Eq. 3), and over similar timescales (1960 to 2010), especially given that groundwater development for agriculture in the TLB largely commenced around 1950. This study's smaller IQR compared to Hansen et al. (2018) may suggest that our model parameters are over-constrained, and thus, do not reproduce the wider distribution of observed TDS IQR increase. However, it is also possible that the larger IQR from Hansen et al. (2018) indicates insufficient sampling (i.e., a perfectly random spatial sample with enough observations might yield a more constrained distribution of TDS measurements that more closely approximate the true population IQR). Nonetheless, given the broad aim of this study to estimate the approximate timescales of regionally downward salinization of the production aquifer under ABCSAL, the evolution of mass flux described by our model generally agrees with observations of shallow aquifer TDS increase in the TLB.

Unsustainable groundwater management eventually leads to undesirable effects (Giordano, 2009; Sustainable Groundwater Management Act, 2014), such as: chronic groundwater level declines and depletion of groundwater storage; well failure (Pauloo et al., 2020); increased energy costs for pumping (Wada et al., 2010); land subsidence (Smith et al., 2017); sea water intrusion (Zektser et al., 2005); desiccation of groundwater dependent ecosystems (TNC, 2014); and groundwater quality degradation (Smith et al., 2018; Foster et al., 2000). The negative externalities above are recognized consequences of unsustainable groundwater extraction. However, ABCSAL, which progressively deteriorates groundwater quality over decades to centuries, may be considered an additional, unrecognized threat to regional groundwater quality and sustainability in the TLB, and a constraint on groundwater sustainable yield in other food production regions of the world.

4.2. Key features of ABCSAL

ABCSAL arises from groundwater development, and is sustained by basin closure. Once a basin in closed, salinization does not depend on groundwater overdraft per se, but rather, on the closure itself, which prevents the basin from discharging salts.

Our findings indicate that the long-term fate of basins closed by groundwater pumping may be similar to that of naturally closed basins (Hardie and Eugster, 1970; Jones and Deocampo, 2003). However, unlike naturally-occurring closed basins, salt cycling in agriculturally intensive closed basins is driven by human-made water management decisions, and may progress more rapidly. Near the onset of the 21st century, average vertical groundwater movement in the Central Valley increased by about 6 times the rate from pre-development conditions, mainly as a result of agricultural recharge and withdrawal from public-supply and irrigation wells (Williamson et al., 1989). Strong vertical transport coupled in a closed basin drives TDS migration into deeper aquifers.

Although groundwater levels in the TLB are in chronic decline (Scanlon et al., 2012), groundwater overdraft is not a necessary condition for ABCSAL to occur. To illustrate this point, we eliminated overdraft (Eq. (7)) by increasing clean recharge M (TDS = 32.5 mg/L) at 0.68 km^3/yr following Hanak et al. (2019), and reducing pumping by 15%. We still observed groundwater salinization, even though the water budget remained in steady state. We also applied completely clean recharge with TDS = 0 mg/L (Appendix Table A.9), and found that it was insufficient to stop or reverse ABCSAL because it did not fix the underlying basin closure. Thus, an area will accumulate salts if groundwater storage is stable or even increasing, as long as the basin remains closed and salts cannot exit. As our model assumes no-overdraft conditions, the results presented herein may be more severe if overpumping were instead to continue.

Our study shows that ABCSAL is exacerbated by imported salts in surface water for irrigation, and by groundwater pumping. Although both surface water and groundwater irrigation are present in our study area, like overdraft, they are not necessary conditions for ABCSAL. However, basins with significant groundwater irrigation are particularly susceptible because pumping lowers groundwater levels and cuts off lateral outflow and subsurface baseflow exits, thus initiating ABCSAL.

The rate and magnitude of salinization depends on a variety of factors (e.g., concentration of total applied water, evapoconcentration, vertical groundwater velocity), but fundamentally depends on the severity of basin closure. Worldwide basins range from open (i.e., natural salt exits maintain freshwater conditions), to partially closed (i.e., some salts exit, but some remain and accumulate), to fully closed (e.g., salts have no exit and hence accumulate in deep groundwater). Groundwater salinization timescales in partially closed basins may be longer than those calculated in this study for the TLB, which is completely closed. Conversely, some basins may salinate at faster rates than calculated for the TLB, depending on the hydrologic features represented in our mixing model.

4.3. Implications for groundwater management

This study demonstrates that if irrigated groundwater basins are operated in a way that hydrologically closes them, groundwater salinization (ABCSAL) is inevitable. It further demonstrates that the timescales of this phenomenon in the TLB are similar to those over which the groundwater in storage would be virtually exhausted according to classic concepts of overdraft. We know how to prevent overdraft by, for example, decreasing pumping or increasing recharge. This raises the parallel question: "How do we prevent ABCSAL?" In other words, how do we both develop groundwater resources, while also keeping groundwater basins hydrologically open?.

Conceptually, one way to both pump abundant amounts of groundwater and to keep the water table sufficiently shallow to produce groundwater discharge (via baseflow and lateral flow to adjacent basins) is to significantly increase groundwater recharge. In California this could in theory be accomplished by storing less water in surface reservoirs and storing more water in groundwater via managed aquifer recharge operations (Kocis and Dahlke, 2017; Ghasemizade et al., 2019; Gailey et al., 2019). Such an approach would be a radical shift from how our current civilization chooses to store water – mainly in surface reservoirs. In the discussion that follows we are not so much advocating such a paradigm shift in water resources management as we are suggesting the need for the beginnings of new conversations in water resources management about how to manage groundwater and surface water jointly in a way that better ensures the sustainability of both.

One challenge of filling up a groundwater basin enough to open it is to manage the water table sufficiently to prevent undesirable waterlogging effects. This would require changes in basin water resources management within a carefully managed scheme in which the pumping and recharge are optimized such that the basin opens up, while preventing the water table from getting so high that bare soil evaporation exacerbates salinization, as happened on the west side of the San Joaquin Valley (Schoups et al., 2005; Belitz and Phillips, 1995). The technology to monitor a groundwater basin and model it sufficiently to tightly manage it for optimal water table elevations does in fact exist (Calderwood et al., 2020), but would require levels of groundwater monitoring, modeling, and decision-making that are well beyond what is normally done. Additionally, achieving vast quantities of recharge will require strategic siting (Maples et al., 2019) to ensure that the subsurface geology can accommodate the recharge within a time frame that does not jeopardize the health of overlying crops (Dahlke et al., 2018) or negatively impair other land uses. In the TLB, a further challenge would be that additional sources of clean recharge water within the TLB watersheds are not large enough to accomplish the requisite amounts of recharge, as rather drastic amounts of pumping reduction would likely be necessary, unless water for recharge could be imported from wetter northern Central Valley watersheds (Hanak et al., 2019). Moreover, conditions of reduced pumping and increased recharge will still lead to vertically dominant solute migration; thus the application of clean recharge, in addition to diluting and lessening the overall salt load, will also contribute to vertical migration of the salt front. Hence, the short- and long-term consequences on groundwater quality of increasing clean recharge and reducing pumping need investigation, which in turn would require the development of regional groundwater quality management models (Fogg and LaBolle, 2006; Kourakos and Harter, 2014).

If re-operation of the groundwater basin to increase groundwater storage and open the basin does not happen, water users in the TLB will ultimately be faced with desalinating pumped groundwater for drinking water and irrigation, the ultimate costs of which remain unknown. If inland closed basin salinization proceeds at the historical rates projected in this study, the salinity of pumped groundwater may exceed thresholds safe for crop health within decades to a few centuries, depending on the depth of pumped groundwater. As prices for technology like reverse osmosis fall, and arid countries pioneer large-scale inland desalination plants for brackish groundwater (Nativ, 2004; Tal, 2006), desalination cost must be weighed against the cost of adaptive water management (e.g., fallowing fields, securing higher quality imported water, managed aquifer recharge) (Hanak et al., 2019). Moreover, economically beneficial uses of recovered salt (e.g., manufacturing table salt, road salt, and mineral commodities) may offset desalination costs. Ongoing, unmitigated ABCSAL may require new technology and irrigation methods to utilize saline water in agriculture (Beltrán, 1999), and the conversion to – and development of – genetically modified salt-tolerant crops (Yamaguchi and Blumwald, 2005).

In order to probe the full impact of ABCSAL in the TLB, particularly on shallow aquifers, which are critical to food and drinking water security worldwide, in this study we assumed no water management intervention as salinity accumulates. In reality, water users would adapt to increasingly saline aquifers by pumping from deeper, less saline aquifers, fallowing fields, mixing saline water with cleaner water, and desalinating pumped groundwater. Two and three centuries into the model, the assumption of no intervention is increasingly unrealistic as the concentration of total applied water approaches thresholds dangerous to crop health, and is likely to have prompted prior adaptive management. We deemed it necessary to evaluate the model at timescales upwards of two and three centuries in order to allow salinization to reach intermediate and deep aquifers. As our model assumes no intervention, results past 50 years of simulation (year 2010) should be interpreted as a worst case scenario.

Urban groundwater pumping might also close groundwater basins. However, there are two key differences between the hydrology of urban and agricultural areas. First, in urban areas, high evapotranspiration rates and subsequent salt concentration are unlikely unless large volumes of water are applied for landscape irrigation. Second, a substantial fraction of urban groundwater pumping (e.g., drinking water, household use, and industrial use) typically exits the basin via wastewater discharge, thus it is not returned to groundwater where it might salinate shallow aquifers (as in the case of the TLB). Hence, the threat of ABCSAL in urban basins is likely to be much less than the threat in agriculturally intensive basins where groundwater is developed and recycled internally.

5. Conclusions

Irrigated agriculture in overdrafted aquifer systems supplies much of the world's demand for food (Dalin et al., 2017). In this study, we demonstrate that intensive groundwater development can transform a fresh, open basin into an evaporation-dominated, closed-basin system. A closed basin is effectively a salt sink: aquifer salinization is inevitable because dissolved solids in groundwater cannot escape, and are recycled through pumpage, irrigation, and evapoconcentration by crops. This study provides a conceptual framework to understand this process, which we call "**A**nthropogenic **B**asin **C**losure and groundwater **SAL**inization" (ABCSAL), and a mixing cell model to provide first-order estimates of ongoing aquifer salinization in the TLB, located in California's Central Valley.

Our model indicates progressive salinization (>1,000 mg/L) of shallow aquifers (36 m) within decades. Intermediate (132 m) and deep aquifers (187 m) are impacted within two to three centuries. The TLB in California's southern Central Valley is less than one century into this "experiment" and the first signs of shallow aquifer salinization have been observed (Hansen et al., 2018; CRWQCB, 2018). Estimated salinization timescales are similar to estimated aquifer depletion timescales in the area (Scanlon et al., 2012), underscoring the urgency of regional-scale groundwater quality management. Importantly, however, while most groundwater quality management frameworks focus on contaminant source control, ABCSAL can only be prevented by allowing contaminant discharge via hydrologic opening of the basin.

This study is a first-order calculation of ABCSAL in an agriculturally intensive groundwater basin. Future research should emphasize a more comprehensive representation of subsurface transport processes through the development of groundwater quality management models. Key research questions that remain include investigating if managed aquifer recharge with relatively clean water may slow groundwater salinization. It also remains to be tested if it is possible to reverse groundwater salinization by increasing recharge until a basin "fills up" and discharges TDS into streams and lateral outflow which exit the basin. The practical likelihood of this mitigation strategy would require re-imagining integrated water resources management with a greater emphasis on subsurface storage. Ongoing ABCSAL without intervention may necessitate inland desalination to remediate saline groundwater resources, the costs of which remain presently unknown.

Traditionally, the concept of long-term sustainability of groundwater has hinged on the intuitive notion of not managing the basin in ways that result in eventual exhaustion of the groundwater stores. Herein we advance the less intuitive concept that long-term sustainability of groundwater also hinges on the salt balance, which in turn depends on how the groundwater quantity is managed. Fundamentally, ABCSAL can only be prevented by managing the basin groundwater quality in ways that open the basin.

CRediT authorship contribution statement

Richard A. Pauloo: Conceptualization, Data curation, Formal analysis, Investigation, Methodology, Software, Validation, Writing - original draft, Writing - review & editing. **Graham E. Fogg:** Conceptualization, Funding acquisition, Methodology, Project administration, Supervision, Writing - review & editing. **Zhilin Guo:** Conceptualization, Formal analysis, Investigation, Methodology, Supervision, Writing - review & editing. **Thomas Harter:** Conceptualization, Formal analysis, Methodology, Writing - review & editing.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, athttps://doi.org/10.1016/j.jhydrol.2020.125787.

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