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 “nthropogenic asin losure and groundwater inization” (ABCSAL). We then 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 in agricultural regions worldwide constitutes a largely unrecognized constraint on groundwater sustainable yield on similar timescales to aquifer depletion, and poses a serious challenge to global groundwater quality sustainability, even where water levels are stable.

Anthropogenic Basin Closure and Groundwater Salinization (ABCSAL)

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Key Points:

- Groundwater pumping may close a basin, leading to TDS accumulation in the production aquifer.
- We describe the process of ABCSAL, which can only be reversed by opening the basin.
- We develop a mixing model to estimate the rate of ongoing ABCSAL in California's Tulare Basin.

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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 then 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 in agricultural regions worldwide constitutes a largely unrecognized constraint on groundwater sustainable yield on similar timescales to aquifer depletion, and poses a serious challenge to global groundwater quality sustainability, even where water levels are stable.

Plain Language Summary

Although pumped groundwater is widely used for drinking water and irrigation, it is generally unrecognized that groundwater pumping at rates sufficient to prevent overdraft may still render the groundwater resource nonsustainable because of negative effects on the basin salt balance. We describe how groundwater pumping may convert an open, fresh basin into a closed-basin system that gradually salinates, even if modern groundwater management halts falling water levels. We then examine the time scales over which an unconsolidated sedimentary aquifer system is degraded over its entire vertical extent. We use the Tulare Lake Basin, California, as a case study of an irrigated agricultural basin in a semi-arid climate with historic groundwater overdraft. Even for modern water management practices that successfully arrest historic water level decline, our mixing-cell based

model indicates that groundwater salinity exceeds safe drinking water limits within decades for shallow aquifers (36 m), and two to three centuries for intermediate (132 m) to deep aquifers (187 m). Increasingly saline pumped irrigation water can negatively impact crop yield, necessitating significant land use change or the eventual desalination of pumped groundwater. Timescales are of similar order as those found for aquifer depletion in other basins, with or without ABCSAL. ABCSAL itself, even absent of aquifer depletion, therefore poses a serious threat to long-term quality and sustainability of global groundwater resources.

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 & 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, 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, 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 et al., 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, 2019b). 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, 2019a, 2019b). 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), (2) naturally-occurring salinization in closed surface-water basins (i.e., endorheic basins and playas) (Eugster & Hardie, 1978; Hardie & Eugster, 1970), (3) high water tables causing groundwater evaporation and soil salinization via capillary rise (Datta & De Jong, 2002; Barrett-Lennard, 2003; Chaudhuri & Ale, 2014; Hillel, 1992), and (4) soil salinization due to irrigation (Hanson et al., 1999; Bernstein & 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 et al., 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 & Eugster, 1970; Eugster & Hardie, 1978; Jones & 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 & 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 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 & Lall, 2017; Barlow & Leake, 2015; Hunt, 1999) and reverse existing groundwater gradients at subsurface outflow boundaries (Figure 1A). Progressively greater closed basin conditions diminish and eventually entirely eliminate natural TDS export from the groundwater basin (Figure 1C). 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 & Eugster, 1970; Jones & 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 by pumping wells to the land surface and the process repeats.

[FIGURE 1 about here]

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 & 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 300 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 & 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 (Figure 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 the clastic sedimentary aquifer system deposited mainly by fluvial and alluvial processes. 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 & Swain, 1995; K. R. Belitz & Heimes, 1990; Deverel & 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 & 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, 2018).

[FIGURE 2 about here]

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 (Figure 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 & 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; 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 & 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 (Figure 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

A lumped parameter approach based on upscaling water fluxes of a fully three-dimensional groundwater model was employed as an appropriately parsimonious modeling tool, given the large space and time scales of interest, and the large-scale effectively one-dimensional flow conditions in the basin. While local hydrogeologic conditions vary widely and lead to locally complex three-dimensional flow and transport conditions, our focus here is on large scale salinization behavior and time scales, which can be well-captured with a one-dimensional approach. We assess results against existing fully three-dimensional flow and salt transport models that also address aquifer heterogeneity, albeit at spatial scales significantly smaller than the TLB, to ascertain the appropriateness of the mixing cell approach chosen here.

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 (M. E. Campana, 1975; M. Campana & Simpson, 1984; M. E. Campana, 1987; Carroll et al., 2008; Kirk & Campana, 1990). A mixing cell approach segments the system into a set of control volumes. In each time step of the model, water and salt masses are passed between the cells, and new concentrations are calculated at each cell. 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 equation 4) below, thus avoiding numerical dispersion issues. 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 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 SI Appendix Table S4 and Figure S3; 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 $[ML^{-3}]$. Dissolution of mass along groundwater flow paths is well documented in sedimentary aquifers (Palmer & 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 & 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. For sensitivity analysis, 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 (Figure 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...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 equation (4) – the pumping, P_j, lateral basin flow I_j, or subsidence flow C_j (Figure 3) conceptually flow into or out of the mixing cell bottom. The following sections provide further details on the parametrization of (3) and (4).

[FIGURE 3 about here]

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, 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 SI Table S1). 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 equations (3) - (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 (Figure 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 (SI Table S8) and found that it had a negligible impact on resulting salt concentrations presented in this study (SI Table S7).

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 (SI Appendix Figure S1 and Table S2). 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) (SI Appendix Table S2). 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. 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 (SI Table S3), 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 (Figure 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 SI Table S6.

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 SI Appendix Table S3), 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 (SI Table S1) 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:

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

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 ($m_{P,k}$):

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

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 (Figure 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 & Jackson, 2016; Kharaka & Thordsen, 1992; DeSimone et al., 2010).

[FIGURE 4 about here]

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 (SI Table S5), 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 model would be computationally prohibitive. Parameter ranges are estimated from literature for rock-water interaction coefficient (Williamson et al., 1989; Kang & 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).

Rock-water interactions is the perhaps most uncertain parameter, thus, in order to understand its influence 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).

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.

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

[Figure 5 about here]

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 1.7 to 3.5 depending on the timeframe considered. When rock-water interactions are present ($\rho > 0$), they initially contribute a comparable mass to groundwater pumping (around 4 metric $Mtons$), but with time, salt accumulates in the aquifer, and the mass recycled by groundwater pumping exceeds the mass imparted by rock-water interactions (Figure 5A).

Annually, surface water diversions add 1.5 metric $Mtons$ of salt to the study site. This is around 4 times the amount of all other non-pumping water budget terms combined (I, M, R, B), which add only 0.35 metric $Mtons$. We estimate that rock-water interactions add between 3.3 metric $Mtons$ and 4.6 metric $Mtons$ of salt annually. 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 (Figure 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 (Figure 6). Intermediate (132 m) and deep aquifers (187 m) exceed 1,000 mg/L within century-long timescales.

[FIGURE 6 about here]

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.

At the beginning of the simulation (year 1960), initial TDS concentration increases gradually with depth (Figure 4 and SI Appendix Table S7). Shallow aquifer salinity is 506 mg/L. After 50 yrs with $\rho = 0$, average shallow aquifer salinity reaches a median concentration of 975 mg/L with an interquartile range (IQR) of 871 - 1,124 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 1,314 mg/L (IQR: 1,100 - 1,654 mg/L). Finally, after 300 yrs (year 2310), median shallow aquifer TDS approaches nearly 1,574 mg/L (IQR: 1,264 - 2,103 mg/L).

Intermediate and deep aquifers are impacted much later than shallow systems, and exceed the freshwater TDS threshold on timescales of two to three centuries. After 200 yrs (year 2160), intermediate aquifer median TDS exceeds 1,000 mg/L (IQR: 861 - 1,048

mg/L). After 300 yrs (year 2260), deep aquifers (IQR: 867 - 1020 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 slightly amplified, 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 (Figure 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 representing the dominant hydrologic features of a system. These strengths come with some tradeoffs. Mixing cell models 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), 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 would be more appropriate (Zhang et al., 2006; Guo et al., 2019, 2020; Henri & 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 (SI Figure S2) 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.

4 Discussion

4.1 ABCSAL: an urgent threat to regional groundwater quality degradation, and constraint on sustainable yield

In this study we show that at its most fundamental level, anthropogenic groundwater basin closure and salinization, or ABCSAL, is a hydrologic process where salts are recycled within a basin because of the elimination of exits for the salts due to basin closure. For the TLB, our calculated ABCSAL timescales proceed at similar rates to those of aquifer depletion. ABCSAL also agrees with and provides a model for observed decadal changes in shallow groundwater salinity of the TLB. Thus, ABCSAL constitutes an unrecognized form of regional groundwater quality degradation, with uncounted, significant constraints on groundwater sustainable yield that may be as urgent as aquifer depletion.

Scanlon et al. (2012) used the CV Hydrologic Model (Faunt et al., 2009) to estimate the lifespan of the Central Valley aquifer at 390 years, based on a remaining water storage in the year 2000 of 860 km^3 and a depletion rate of 2.2 km^3/yr. Scanlon et al. (2012) also note that aquifer lifespan is likely shorter in the TLB (southern CV) due to focused groundwater depletion in the area. 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 approximately 390 year timescale of aquifer depletion.

Measured TDS change from historic (1910) to modern (1993-2015) time periods in the TLB (Hansen et al., 2018) agree with this study's modeled changes in TDS over similar timescales (1960 to 2010). Hansen et al. (2018) found that median shallow aquifer (less than 50 m deep) TDS increased by 143 - 241 mg/L, with an IQR increase of 110 - 850 mg/L, depending on the region considered in the TLB. Across the entire TLB, our "No rock-water interactions" ($\rho = 0$) results indicate a median increase in shallow aquifer (36 m deep) TDS of 469 mg/L with an IQR increase of 365 - 618 mg/L. When rock-water interactions are considered ($\rho > 0$), the median increase in shallow aquifer TDS is 605 mg/L with an IQR increase of 601 - 759 mg/L. Our IQR ranges of TDS increase with and without rock-water interactions are well within the IQR range measured by Hansen et al. (2018). Differences in median TDS and IQR may be explained in a number of ways. First, a careful examination of the spatial distribution of TDS samples reported by Hansen et al. (2018) reveals that sampling was not entirely representative of the entire TLB; in particular, missing samples from the west side of the valley where shallow TDS should be higher might explain why our lumped model (which averages conditions across the TLB) estimates a higher median increase. Moreover, the smaller size of our IQR of TDS increase compared to Hansen et al. (2018) may suggest that our model parameters are not constrained enough to reproduce the distribution of observed TDS 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). We point out that the original question of the study was not to perfectly predict increases in shallow aquifer TDS (which is why we do not calibrate the model), but rather to explore the timescales of regionally downward salinization of the entire production aquifer under ABCSAL. In this sense, the findings of Hansen et al. (2018) substantiate the mass balance evolution described by our model.

The historical and modern periods considered by Hansen et al. (2018) and this study do not exactly align with one another, but most groundwater development for agriculture, and hence ABCSAL, commenced in the mid-twentieth century, thus the timelines are quite comparable in terms of the duration of groundwater development. This study's predicted salinization time frames (i.e., decades for shallow systems, centuries for deep systems) are also consistent with random walk salt and nitrate particle transport simulations in detailed 3D heterogeneous alluvial aquifers (Henri & Harter, 2019; Zhang et al., 2006), which suggests that the simple mixing cell model captures key transport dynamics. Thus, it may serve as a useful benchmark for future research with more complex, distributed parameter, regional-scale transport models incorporating geologic heterogeneity and transient boundary conditions.

ABCSAL can explain early observations of salt accumulation California's TLB (Schmidt, 1975) through the process of basin closure. Moreover, ABCSAL's impact on shallow groundwater is well supported by field-based observations that TDS has increased in the TLB over the past half century, and that most of this increase is observed in shallow aquifers

(Hansen et al., 2018; CRWQCB, 2018). These results extend previous modeling efforts to estimate shallow aquifer salt transport (Schoups et al., 2005) by including transport into deeper aquifers and multi-century simulation to evaluate the long-term consequences of basin closure in an agriculturally intensive basin.

Our findings indicate that the long-term fate of basins closed by groundwater pumping and salt recycling are similar to that of naturally closed basins (Hardie & Eugster, 1970; Jones & 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). This change in groundwater movement coupled with basin closure drives the migration of TDS 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 critical point (and prevent drying out the model), we eliminate overdraft via equation (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 %, and still observe groundwater salinization although the water budget remains in steady state. Even applying completely clean recharge with TDS = 0 mg/L (SI Appendix Table S8), makes little difference in the long term salt balance. Arresting overdraft is insufficient to stop or reverse ABCSAL because it does not fix the underlying basin closure. An area may accumulate salts if groundwater storage is stable or even increasing, as long as the basin remains closed and salts cannot exit.

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 to ABCSAL because it is the pumping itself that lowers groundwater levels and cuts off lateral outflow and subsurface baseflow exits, thus initiating ABCSAL.

Unsustainable groundwater management eventually leads to undesirable effects (Giordano, 2009; *Sustainable Groundwater Management Act, California Water Code sec. 10720-10737.8*, 2014), such as: chronic groundwater level declines and depletion of groundwater storage, which may cause well failure and increase 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 (Foster et al., 2000; Smith et al., 2018). The negative externalities above are recognized consequences of unsustainable groundwater extraction. However, ABCSAL, which progressively deteriorates groundwater quality, may be considered an unrecognized threat to regional groundwater quality and sustainability, and significant constraint on groundwater sustainable yield in many food production regions of the world.

4.2 Implications for groundwater management

ABCSAL is driven by the simultaneous processes of hydrologic basin closure and salt input from groundwater applied as irrigation water. The only way to prevent this process from salinating the entire groundwater basin is to re-open the basin by sufficiently filling it up to the point where baseflow to streams and/or lateral flow to adjacent basins resumes, or resolve to eventually desalinate most of the pumped groundwater. Thus the mitigation of ABCSAL may require increasing groundwater storage in the basin by reducing pumpage, increasing recharge, or both. The increased recharge would have to be accomplished with relatively clean (low TDS) sources of water, such as, in the TLB case, high-magnitude flood flows from streams draining the Sierra Nevada (Kocis & Dahlke, 2017) or the Coast Ranges. As long as a basin remains closed, and most of the recharge comes from applied irrigation water, groundwater quality will only worsen due to the salinity of applied water, as well as nitrates (Harter et al., 2012). The short- and long-term consequences on groundwater quality of introducing increasing clean recharge and reducing pumping need to be investigated. This in turn would require the development of regional groundwater quality management models (Fogg & LaBolle, 2006; Kourakos & Harter, 2014) that can represent the effects of heterogeneity and non-Fickian transport.

The above-described changes in basin water resources management need to happen within a carefully managed scheme in which the pumpage 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 (e.g., Schoups et al., 2005; K. Belitz & Phillips, 1995). For many basins, including the TLB, this would be a challenging proposition and would require decades or more of integrated water resources planning and management in which greater emphasis is put on subsurface storage of water rather than surface storage. In reality, additional sources of clean recharge water within the TLB watersheds are not voluminous 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 the wetter Central Valley watersheds located north of the TLB (Hanak et al., 2019).

If re-operation of the groundwater basin to increase groundwater storage, open the basin, and introduce cleaner recharge does not happen, then water users in the TLB will ultimately be faced with having to desalinate pumped groundwater for drinking water and irrigation supplies. The ultimate costs of any future desalination on both drinking water supplies and the food supplies that come from irrigated agriculture need to be evaluated. If inland closed basin salinization proceeds at historical rates as 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), the cost of technological solutions like desalination must be weighed against those of adaptive water management (e.g., fallowing fields, securing higher quality imported water, managed aquifer recharge) (Hanak et al., 2019). Nevertheless, the prospect of possibly having to eventually desalinate much or most of the groundwater used for irrigation worldwide point to potentially catastrophic effects on long-term world food supply and economy. We should anticipate these future costs and impacts now rather than later, and consider whether the longer term stability of the Green Revolution, which occurred in part due to irrigated agriculture (Evenson & Gollin, 2003), is now in serious question.

Our study shows that the rate and magnitude of salinization depends on a variety of factors, including the concentration of total applied water, evapoconcentration at the land surface, the vertical groundwater velocity, and fundamentally, the severity of hydrologic basin closure. Local hydrogeology and water management vary across irrigated basins worldwide, and basins range from open (i.e., natural salt exits maintain freshwater conditions over long timescales), to partially closed (i.e., some salts exit, but some remain and accumulate over long timescales), to fully closed (e.g., salts have no exit and hence accumulate in deep groundwater over long timescales). The timescales of groundwater salinization in partially closed basins may be longer than those calculated in this study for the TLB, which is completely closed. Conversely, some basins may be undergoing salinization at slower or faster rates than calculated for the TLB, and these rates will depend on the hydrologic features described above.

One might suggest that urban groundwater pumping could also close groundwater basins. There are two key differences, however, in the hydrology of urban and agricultural cases. Firstly, in urban cases, high rates of evapotranspiration and subsequent salt concentration are unlikely unless perhaps water applied for landscape irrigation is very high. Secondly, a substantial fraction of urban groundwater pumping is for drinking water, household use, and industrial use, and that water typically exits the basin via wastewater discharge, thus it is not returned to groundwater where it might begin to salinate shallow aquifers (as in the case of the TLB). It therefore appears that the threat of ABCSAL in urban basins would be much less than the threat in agriculturally intensive basins where groundwater is developed and recycled internally.

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, we assume no water management intervention as salinity accumulates. In reality, as ABCSAL progresses, water users will adapt to increasingly saline aquifers in various ways and to differing degrees, including 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 becomes increasingly unrealistic as the concentration of total applied water approaches thresholds dangerous to crop health, and is highly 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.

5 Conclusions

Irrigated agriculture in overdrafted aquifer systems supplies much of the world's demand for food (Dalin et al., 2017). The conventional understanding of groundwater quality in these systems fails to acknowledge that observed changes in shallow groundwater TDS may arise from intensive groundwater development, which can transform a fresh, open basin into an evaporation-dominated, closed-basin system. A closed basin is effectively a salt sink: groundwater 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 suggests 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 troubling experiment, and the first signs of shallow aquifer salinization have been observed (Hansen et al., 2018; Schoups et al., 2005). Importantly, the estimated salinization timescales are similar to estimated aquifer depletion timescales (Scanlon et al., 2012), calling into question the urgency of regional-scale groundwater quality management.

This study is a first-order calculation of ABCSAL in an agriculturally intensive groundwater basin. Model-based uncertainty may be addressed in future research with more comprehensive representations of subsurface transport processes through the development of groundwater quality management models.

Key research questions that remain include investigating whether practices like managed aquifer recharge with relatively clean water may slow groundwater salinization. It remains to be tested if it is possible to actually reverse groundwater salinization by increasing groundwater recharge until a basin "fills up" and discharges dissolved solids into streams which exit the basin, although the practical likelihood of this would require a complete re-imagining of integrated water resources management in systems undergoing ABCSAL. Ongoing salinization without intervention may necessitate inland desalination to remediate saline groundwater resources, the costs of which remain presently unknown.

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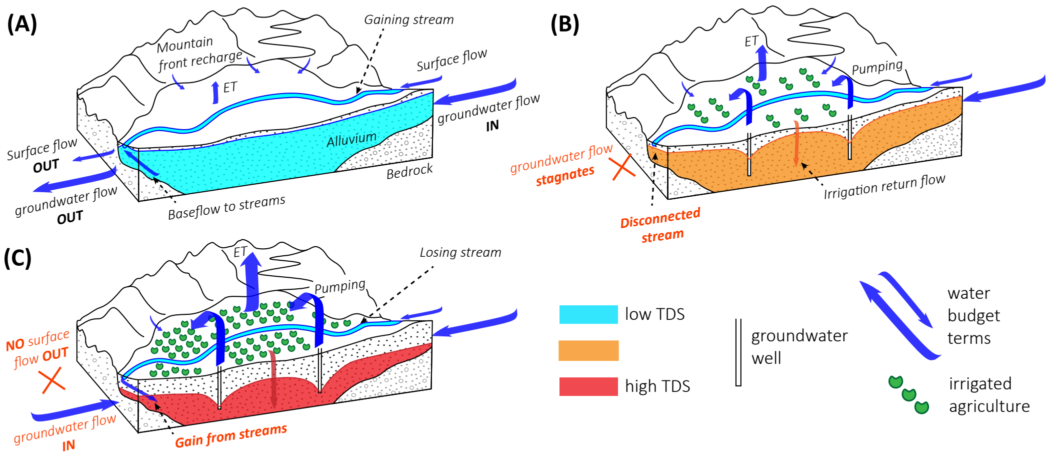

Figure 1. Conceptual model of ABCSAL. (A) Open basin, pre-groundwater development: surface and groundwater systems connect. Groundwater discharges dissolved solids into surface water which exits the basin. Groundwater at this stage is predominantly fresh (e.g., < 1,000 mg/L). (B) Partial basin closure: groundwater pumping causes reduction or elimination of baseflow to streams. Pumped groundwater returns to the basin via irrigation return flow. Dissolved solids begin to accumulate in shallow groundwater. (C) Closed basin: lower groundwater levels cause subsurface inflow to drain adjacent basins. Streams lose to groundwater. Water primarily exits via evapotranspiration (ET), which further concentrates dissolved solids in groundwater. Salts migrate into the production zone of the aquifer, driven by vertical hydraulic gradients from recharge and pumping.

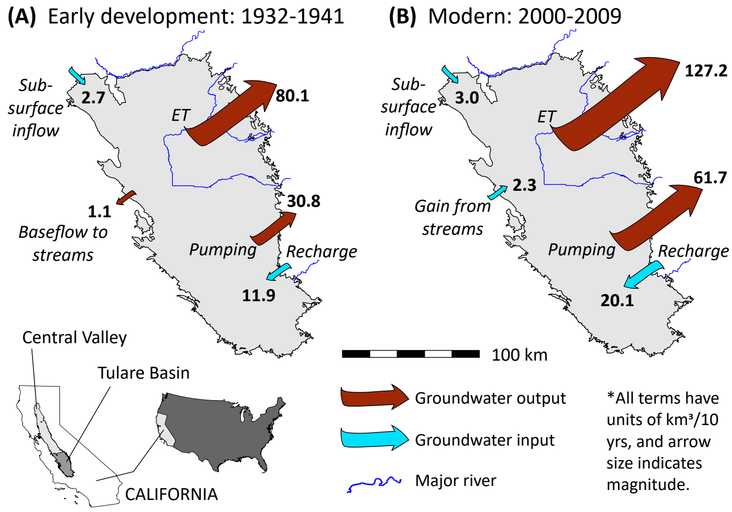

Figure 2. The TLB overlies an agriculturally intensive sedimentary aquifer in California's southern Central Valley. Significant changes are observed in selected decadal hydrologic year water budget terms derived from C2VSim at (A) early-groundwater-development (not to be confused with pre-groundwater-development) 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) or the land surface and rootzone budget (SI Table S1), 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 tributaries are not shown.

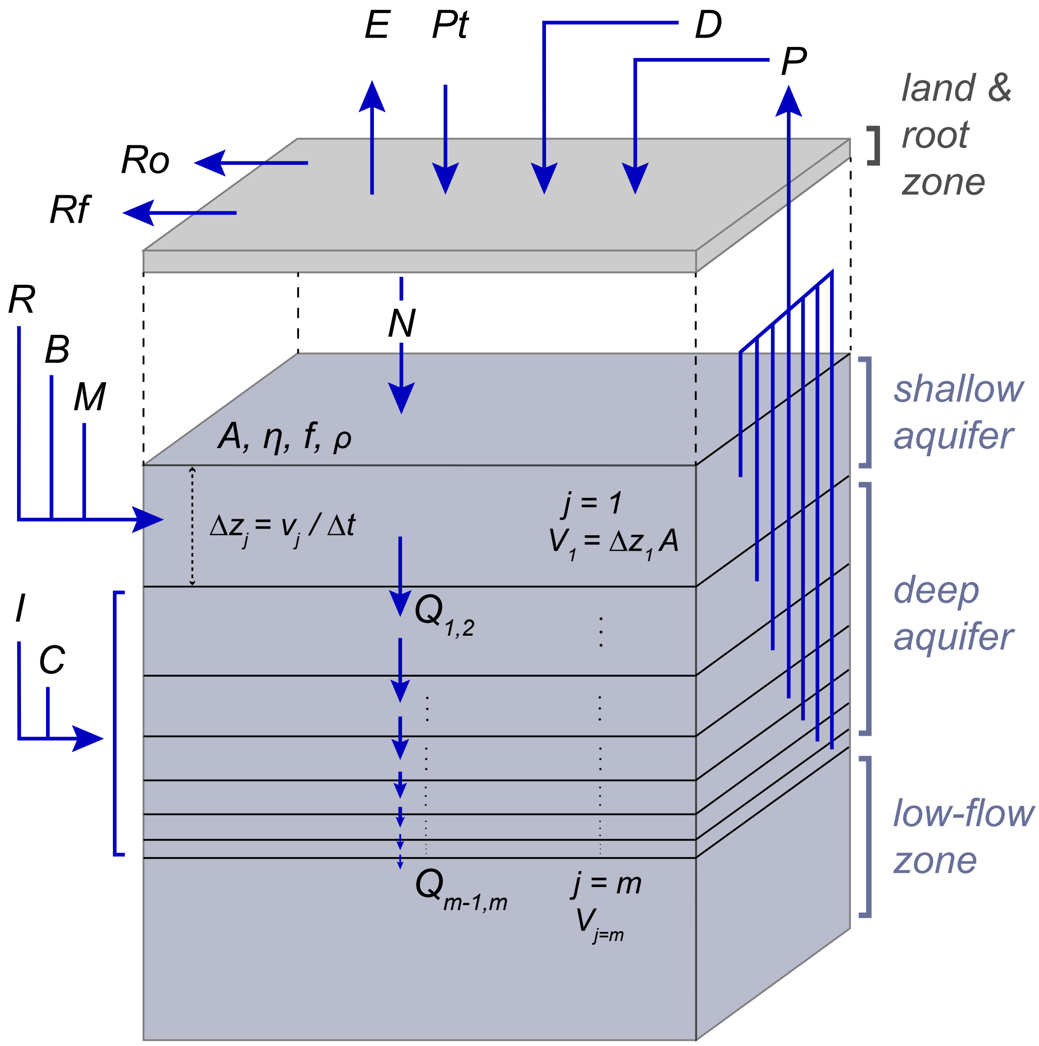

Figure 3. Conceptual land-rootzone model and groundwater mixing cell model with surface area A, porosity η, aquifer fraction f, rock-water interaction coefficient ρ, and m cells. The cell thickness Δz_j is given per equation (4), where linear velocity $v_j = q_j/f\eta$. The cell volume V_j is the total bulk volume of the rock including aquifer and non-aquifer material. The TDS in cell j is calculated by equation (3). The land and root budget (SI Table S1) 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., equation (12). The average annual groundwater and salt budget is reported in Table 1.

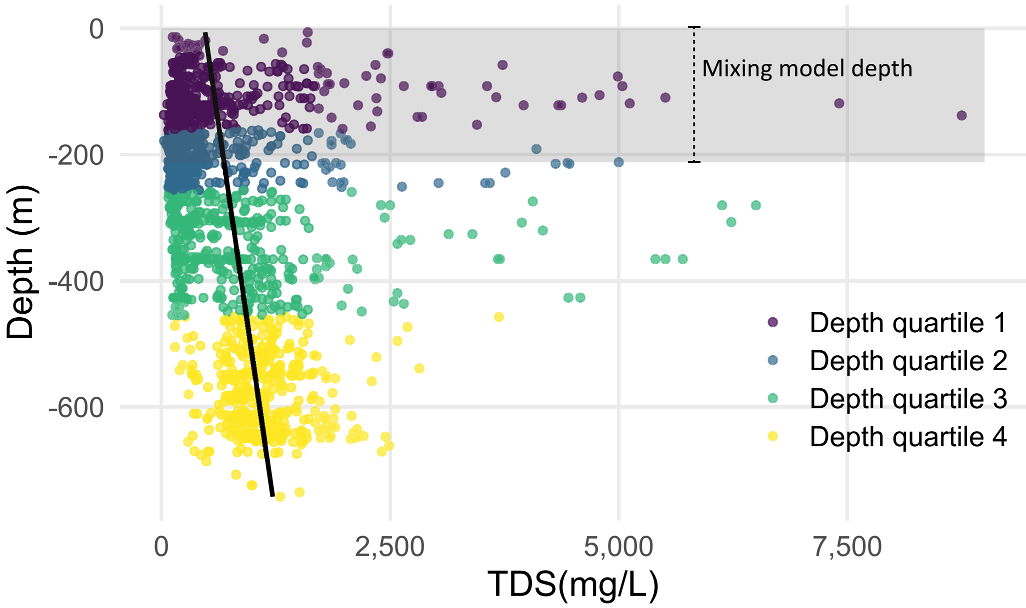

Figure 4. Pre-1960 groundwater quality generally decreases with depth, reaching an average concentration of 1,000 mg/L at 526 m deep. The initial TDS-depth concentration at $t = 0$ is approximated by a linear model, shown as a black line. The transparent, grey rectangle shows the depth of the mixing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

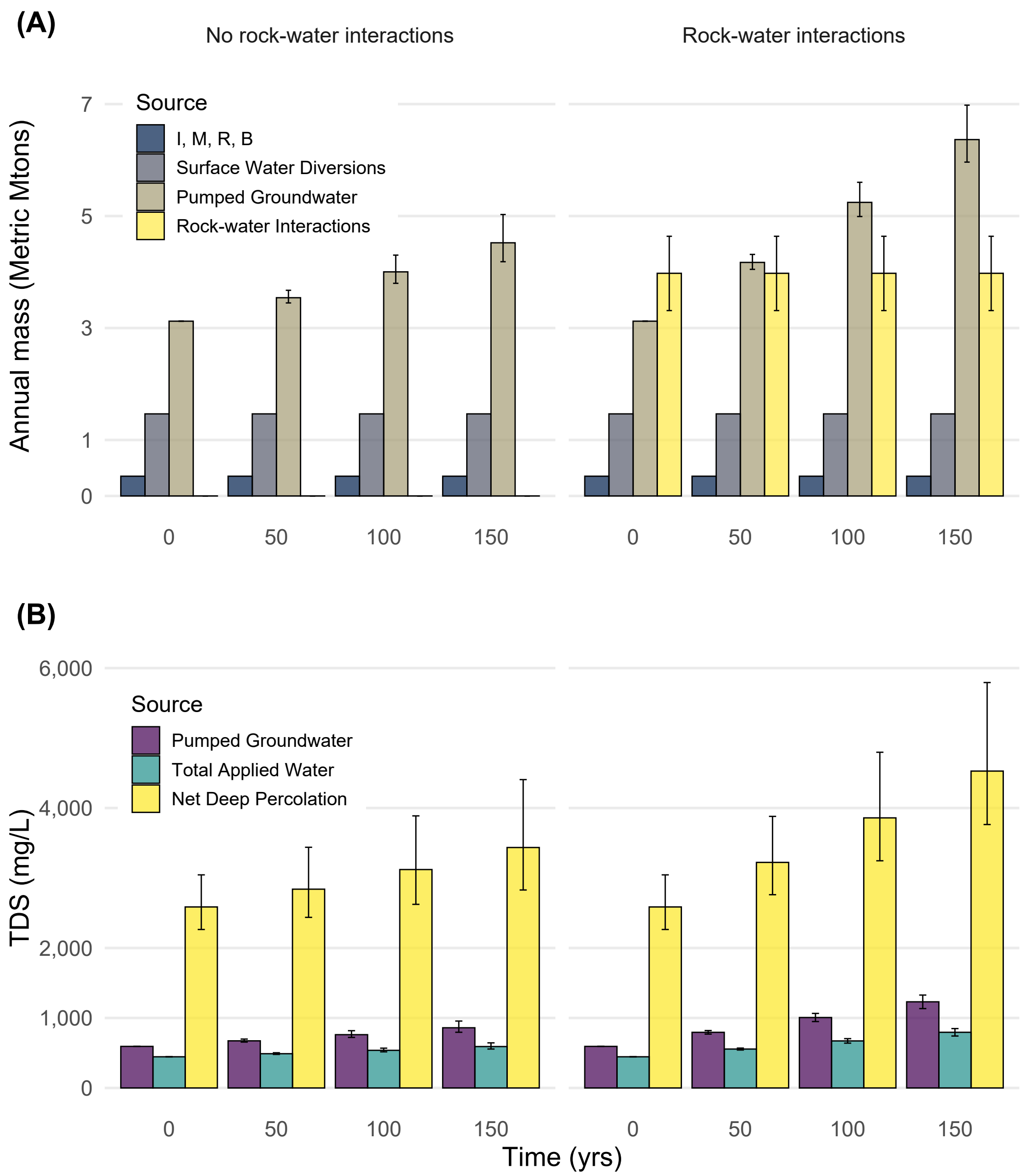

Figure 5. Annual mass flux and TDS of selected budget terms. The height of each column represents the 1,000 scenario ensemble median result, and the width of error bars, if present, represent the interquartile range (IQR) of the ensemble distribution. (A) Annual mass flux varies by source. Pumped groundwater contributes more TDS 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*. Evapotranspiration concentrates total applied water, which enters the groundwater system as *net deep percolation*. Over time in a closed basin system, groundwater salinates, which in turn increases the concentration of total applied water and net deep percolation.

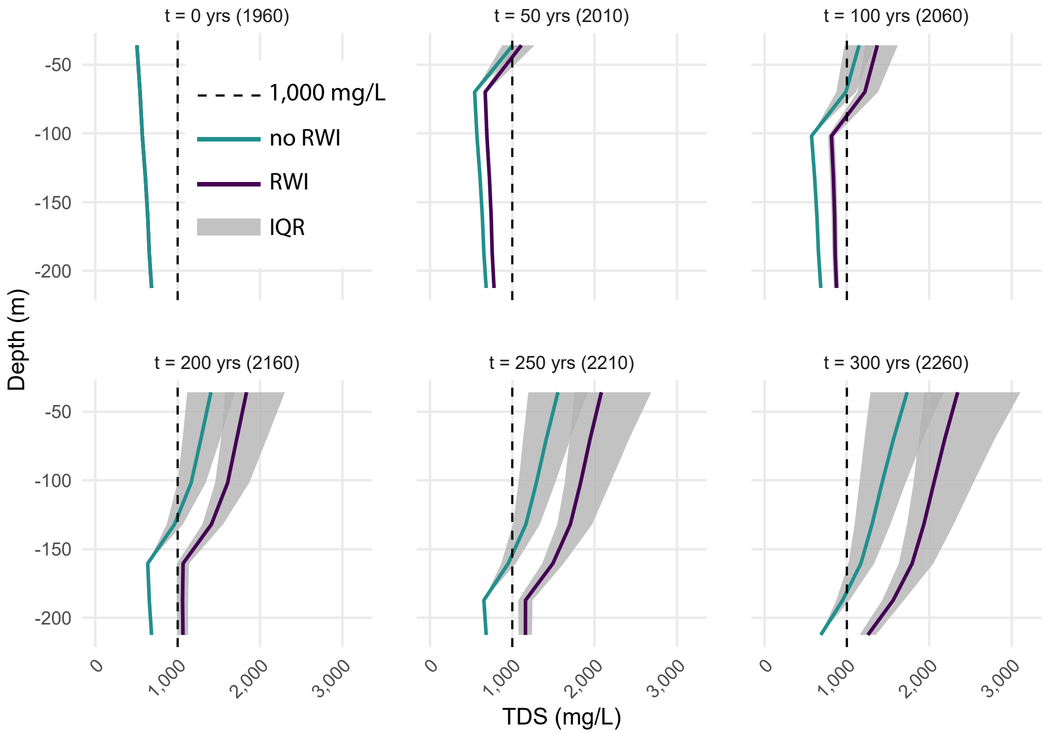

Figure 6. Progression of groundwater salinization ensemble results for two scenarios (with and without rock-water interactions). RWI stands for rock-water interactions. The blue and purple lines show the ensemble median concentration for the two scenarios, and the interquartile range (IQR) of the ensemble simulations is shown as a grey shaded area. Complete statistics are provided in SI Table S7.

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

	Source	Q (km^3/yr)	C (mg/L)	m $(Metric\,Mtons)$
Historical budget	R	2.451	32.5	8.027E-02
	B	0.236	32.5	7.475E-03
	C	0.572	32.5	1.852E-02
	I	0.011	32.5	3.250E-04
	P	-6.761	*	*
	N	1.883	*	*
	RWI	-	-	*
	ΔS	-1.608		
Alternate budget	R	2.451	32.5	8.027E-02
	B	0.236	32.5	7.475E-03
	C_{alt}	0	-	-
	M	0.678	32.5	2.204E-02
	I	0.011	32.5	3.250E-04
	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 of salt. 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 equation (12), and introducing recharge M.

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