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EVALUATING FOCUSED AQUIFER RECHARGE IN THE JORNADA EXPERIMENTAL RANGE, NM, USING CHLORIDE PROFILE ANALYSIS

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Drill rig on the Jornada Experimental Range, NM

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ABSTRACT

Recently published research has suggested that up to 20% of precipitation in a small (~0.05 km^2) piedmont watershed in the Jornada Experimental Range of southern New Mexico becomes deep percolation that leads to aquifer recharge. This finding, based on a watershed mass balance approach measuring precipitation, runoff, evapotranspiration, and soil water storage change through time, challenges a prevalent understanding of Chihuahuan Desert ecohydrology that creosote shrubland prevents recharge. The potential management implications of this finding are consequential. Here we report on research designed to test for the hypothesized focused recharge in small, first-order channels by coring the sediment under two such channels and analyzing the chloride distribution with depth. If abundant recharge occurs, chloride should be flushed through the sediment to the groundwater, but if it does not, we should see an accumulation, or bulge, of chloride at the most frequent depth of infiltration. To test the methodology, we took two cores from flat interfluvial areas far from channels, and one core from a 4-m-wide cobble-bed channel that almost certainly causes focused recharge. In addition to measuring chloride in the cores, we also measured gravimetric water content, soil water potential, and chloride-to-bromide ratios.

The interfluvial cores showed chloride concentrations that rose rapidly within 70 cm of the ground surface and remained high (~ 8,000 mg/kg of pore water) through the remainder of the core, which confirms that chloride has accumulated in these low-infiltration portions of the landscape. Water content was extremely low, with typical matric potential values of −6 MPa and a minimum value of −8.83 MPa, equivalent to −900 m of head. Collection of the large channel core was inhibited by the coarse nature of the channel sediment and most material was lost from the core barrel in retrieval, but the sediment that was obtained had an average chloride concentration of ~100 mg/kg and had a matric potential of ~ 0 MPa, which is effectively saturated.

One channel core, Channel-C, was generally similar to the interfluvial cores, with average chloride concentrations of ~ 8,000 mg/kg of pore water. However, the chloride was displaced downward, beginning to accumulate at 1.0 m depth, and below this the concentration oscillated from 1,000 to 13,000 mg/kg for 1.5 m before stabilizing. Soil water potential also oscillated in unison with the chloride concentration, from ~ 0 to −7 MPa. These oscillations may indicate preferential lateral flow paths, which could produce recharge that bypasses high-chloride sediments in the vadose zone, or they may be an artifact of our sampling method. The other channel core, Channel-A, had ~ 0 MPa water potential and low chloride concentration down to 3 m depth. Below this, chloride concentration rose to ~ 3000 mg/kg and water potential decreased to −2 MPa and both remained at these approximate values to the end of the core at 5 m depth. Core Channel-A was taken at a unique position on the landscape, where a road dammed the channel and caused ponding during runoff events. This road was built 48-77 years before the study, based on time-bracketing air photos. Our interpretation is that while Channel-C is in steady state, with root extraction of soil water in balance with infiltration, Channel-A exhibits a transient wetting front, with plants still expanding to catch up with the newly available water supply. Yet even in this unique location, the augmented infiltration has not removed the chloride accumulation from the underlying soil. Unless lateral preferential flow is active here, this suggests that percolation has not yet infiltrated beyond the root zone even in this recharge hotspot.

These observations are limited to single points on the landscape, and do not preclude the possibility of lateral or preferential flows. However, the extremely negative matric potentials we observed have been associated with limited preferential flow by previous workers. We suggest that the chloride profiles are a more robust indicator of infiltration processes than the watershed mass balance approach. In particular, the location of the evapotranspiration measuring equipment appears to have produced a sampling footprint that excludes the enhanced transpiration along the banks of the small channels within the watershed. We recommend that water planners refrain from incorporating small-catchment focused recharge from Chihuahuan Desert piedmont slopes into their resource assessments until the results of these two studies can be reconciled.

Keywords: Channel infiltration, Aquifer recharge, Chloride profile analysis, Dryland ecohydrology

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1. INTRODUCTION

1.1 Motivation

Groundwater represents 25% of fresh water used in the United States, and in southwestern states like New Mexico, groundwater makes up about 50% of all water used (Maupin et al., 2014). Groundwater is frequently extracted at a faster rate than it is replenished, leading to the depletion of this crucial resource. Populations in the American Southwest are growing, and, due to climate change, runoff and streamflow are declining (Udall and Overpeck, 2017). These two factors make it essential to evaluate the water resources available and to develop a sustainable groundwater extraction plan. For a groundwater extraction plan to be sustainable in the long-term, average yearly water use should not exceed the average amount of water that replenishes the aquifer each year. Therefore, it is essential to determine the rate at which the aquifer is replenished.

Aquifer recharge, the amount of water from precipitation and runoff that percolates down to the water table, is one of the smallest components of the water cycle and can be extremely difficult to quantify. Recharge varies a great deal across the landscape depending on the characteristics of the land surface and subsurface, such as soil thickness, vegetation cover, and hydraulic conductivity. Recharge occurs at a range of scales, from extensive diffuse recharge through the soil to localized focused recharge through ephemeral channel beds, and is difficult to observe and measure. Recharge is therefore often estimated, when assuming a steady state condition or a sufficiently long integration time, as the remainder of the other water balance components:

$$P - Q - ET = R \tag{1}$$

where P is rainfall, Q is runoff or streamflow, ET is evapotranspiration, and R is recharge. Rainfall and runoff are relatively easy to measure directly, using rain gauges and weirs, but evapotranspiration is a difficult quantity to measure. Because rainfall and evapotranspiration (and runoff during certain events) are very large quantities and recharge is a very small quantity, even small percent errors in the estimates of the left-hand side of the water-balance equation can lead to orders-of-magnitude errors in the recharge estimate. Eddy-covariance (EC) towers are capable of generating relatively good estimates of ET, but because one of the assumptions of the EC method is that the terrain is horizontal and uniform (Burba and Anderson, 2010), their accuracy may be questionable in heterogeneous natural environments. For example, the use of flux towers in narrow, arid-region riparian zones is greatly complicated by the advection of hot dry air from the surrounding xeric landscape and the readjustment of the boundary layer to the greater aerodynamic resistance of the riparian trees (e.g., Baldocchi et al., 1988). EC towers also have limited suitability for large-scale ET estimations, because they are limited by the extent of the footprint they sample. Satellite remote sensing is often used for large-scale ET estimates, but points for calibration of ET for remote-sensing estimates are sparse and ET varies greatly over the diurnal cycle, which is not captured by satellite images. In arid areas, the long-term cumulative ET reported by satellite-based estimates is often greater than the amount of precipitation, indicating that this method can overestimate ET in this context (Parrish, 2019). Most remote-sensing methods also have issues with mountainous topography and work best in flat terrain (Aguilar et al., 2010; Senay et al., 2011).

Previous efforts funded by the New Mexico Water Resources Research Institute to better constrain the rate and distribution of groundwater recharge include development of the Python Recharge Assessment for New Mexico Aquifers (PyRANA) model (Ketchum, 2016; Revelle, 2017; Xu, 2018; Parish, 2019). This model tracks soil water in 250 m grid cells covering the entire state at a daily time step. PyRANA uses daily gridded PRISM precipitation data, daily reference evapotranspiration, and soil property data to generate recharge estimates across the entire state. It provides an alternative to large-scale remote sensing ET estimation, producing ET estimates that are limited by both the input of precipitation and the energy budget. A key limitation, however, is that it only tracks moisture in a vertical soil column, and is thereby limited to estimating diffuse recharge. Lateral surface runoff is estimated but not tracked to where it may accumulate in channels and potentially produce focused recharge.

Transmission losses from ephemeral channels can be a large component of groundwater recharge in arid regions (Wainwright et al., 2002; Shanafield and Cook, 2014; McCallum et al., 2014). For example, flume measurements of discharge along the flow path of a flash flood in the Walnut Gulch Experimental Watershed clearly quantified decreasing discharge with distance downstream (Goodrich et al., 2004). But what if focused recharge is happening along smaller channels? Could apparently insignificant rills be contributing to focused recharge?

Recent research using micrometeorological techniques and catchment water budgets to study transmission losses on the piedmont slopes of the semiarid Jornada Experimental Range (JER) suggests that 20% of long-term precipitation becomes 'deep percolation' (Schreiner-McGraw and Vivoni, 2017). To conduct their study, these authors measured the water balance components in a highly instrumented watershed of 0.047 km^2 (4.7 hectares) in the JER. Rainfall and runoff were closely monitored using four tipping-bucket rain gauges and three flumes internal to the watershed, and one flume at the outlet. An eddy-covariance tower, installed upslope of the watershed, was used to monitor ET, and three transects with five profiles each were instrumented with soil dielectric probes installed at 5, 15, 30, 50, and cm depth. In another study, Schreiner-McGraw and others (2016) compared changes in soil moisture determined by this distributed sensor network to the cosmic-ray neutron sensing method and found that the results of each method closely aligned.

The key finding in these studies is an estimate that over the last 100 years, 48 mm/year of percolation has occurred, resulting in a ratio of percolation to rainfall of 0.19, and a basin-outlet streamflow-to-rainfall ratio of 0.02 (Schreiner-McGraw and Vivoni, 2017). To determine stream bed percolation P_{SB}, expressed as depth of water per unit time averaged over the entire watershed, they used Shanafield and Cook's (2014) equation:

$$P_{SB} = \frac{1}{A_t} Z_c W_c L_c \frac{\Delta\theta_c}{\Delta t} \qquad (2)$$

where Z_c is the depth of sediment in the channel, θ_c is the depth-averaged channel soil moisture, W_c is the average channel width, L_c is the total channel length, and A_t is the watershed area. The watershed water balance was expressed as:

$$Z_r \frac{\Delta\theta}{\Delta t} = P - ET - Q - R \qquad (3)$$

where Z_r is the soil depth, defined by the rooting depth of the vegetation, $\Delta\theta/\Delta t$ is the change in volumetric soil moisture content in the rooting zone over time, P is precipitation falling as rain, ET is evapotranspiration, Q is streamflow out the watershed outlet, and R is deep percolation below Z_r, which is equivalent to aquifer recharge. Schreiner-McGraw and Vivoni (2017) evaluated the first four terms of Eq 3 for their study watershed at a monthly timestep for 71 months, and obtained a cumulative deep percolation for the entire 71-month study period. They found that percolation was greater than ET during the summer monsoon season from July – September, but they found no deep percolation during periods without runoff.

The shrub-vegetated rangelands of the Chihuahua Dessert, where the study took place, were not previously thought to contribute to recharge. Water infiltrating into the beds of rills and small channels (less than a meter in width) was assumed to become transpiration for the plants lining the channels before it was able to percolate beyond the root zone to the water table. But there has been very little research done on focused recharge in streams of this size to confirm this inference. If the findings of Schreiner-McGraw and Vivoni (2017) are going to influence water planning and management decisions, we suggest it would be prudent to first confirm the results before abandoning the existing understanding of these small piedmont catchments. Hence, the goal of this study is to test independently the finding of deep percolation originating from small ~1–10 hectare watersheds in the JER by analyzing the chloride profiles of cores drilled from such channels.

1.2 Ecohydrology Background

Desert shrubs such as creosote (*Larrea trindentata*) and mesquite (*Prosopis glandulosa*) are well adapted to arid environments, and capable of withstanding long periods of drought and generating very high matric potentials in order to make use of any infiltrated water (Sandvig and Phillips, 2006). The more water stressed these plants become, the deeper their tap roots extend. At the Desert Botanical Laboratory in Tucson Arizona, Spalding (1904) found roots of a small creosote bush exposed to 3 m depth, and from the size of the roots he estimated the roots extended to a depth of 5.5 m.

Previous arid rangeland infiltration studies have used environmental tracers to study the role of vegetation on vadose zone dynamics. For example, Sandvig and Phillips (2006) extracted soil cores beneath areas dominated by ponderosa pine, juniper, grassland, and creosote. The creosote cores were taken from the Sevilleta National Wildlife Preserve, north of Socorro, NM, roughly 190 km north of the JER, which is the field site in the Schreiner-McGraw study as well as this report. Despite the distance separating these two sites, they are both in the Chihuahuan desert and have a similar climate. Sandvig and Phillips (2006) found that there has been no downward movement of water past the root zone of creosote bushes over the past 20 kyr, and their data actually supported a small net upward flux of water. These cores, however, were from flat interfluvial sections of the landscape, not from small channels where water can accumulate. In a lysimeter study, Gee and others (1994) found that the presence of plants eliminated deep drainage. A lysimeter 6 m in depth and 2.44 m in diameter was filled with Berino loamy fine sands from the JER. They found that if no vegetation was present the annual potential recharge was 25% of the annual precipitation, but when desert plants were added to the lysimeter, no deep drainage occurred. During the study period from 1984 – 1992 the average rainfall was 338 mm/yr, significantly higher than the average rainfall of 234 mm/yr during the time period of the Schreiner-McGraw study.

Large areas of the Chihuahuan desert have experienced a vegetation change from grasses to shrubs over the last 100 years, leading to increased erosion. Wainwright and colleagues (2002) studied the relationship between hydrology and vegetation on the piedmont slopes of the JER by looking at the discontinuous rills that have ‘beads’ or ‘splays’, areas of deposition and loss of channelization that separate segments of continuous rills. Many of these splays have been re-eroded and are now part of continuous segments that may in time be disconnected again. Splays allow water to infiltrate and accumulate sediment as well as nutrients, so they are a crucial part of the ecosystem. The aim of the Wainwright study was to test whether vegetation was more prolific in and near these beads, and, if vegetation was able to become established, whether that would lead to an increase in slope that would eventually lead to re-incision of a channel. They conducted carbon and nitrogen isotope analysis of creosote bushes both in and out of beads, and measured rainfall and runoff both before and after bead formation. They found that the beads experienced net infiltration in all but the three biggest of the 20 flow events studied. The proportion of land covered by vegetation inside and outside the beads was found to be 50.0% and 26.3% respectively, and they found muhley grass much more prevalent in the beads than outside. The results of the carbon isotope study indicate that creosote bushes inside of the beads are much less water stressed than those outside of the beads, but this was in conflict with the xylem-pressure-potential data that showed creosote bushes in the beads were more water stressed than those outside of the beads at the height of the dry season. From this conflicting evidence they postulated that beads experience enhanced infiltration during wet periods, which leads to the growth of more vegetation, but during a drought these areas become more stressed because of the increased vegetation. This is supported by the xylem pressure potentials that were found to be higher in creosote inside the bead than outside of the bead prior to monsoon storms.

Shrubs have replaced grasses in much of the Chihuahuan Desert. Grasslands are now shrublands with higher erosion rates and higher runoff generation. Grass species (*Muhlenbergia porten* and *Bouteloua eriopoda*) are now sparse, with the densest remaining populations located in topographic areas that accumulate water and nutrients (Wainwright et al., 2002). Drought resistant shrubs like creosote and mesquite now dominate the landscape. These shrubs grow slowly and are capable of sending their roots to great depths. Mesquite roots have been reported growing to depths of 53 m (Phillips, 1963). Root depths are highly variable and plants typically do not send down roots to depths that water percolation does not reach. High osmotic potentials in the roots that can overcome high matric potentials in the soil allow plants to extract moisture from soils with very little available water.

Biological activity has been documented at great depths in the JER. Freckman and Virginia (1989) studied the occurrence of nematodes (invertebrates that consume plant roots, microfauna, and microflora) in deep desert soils in the JER. They found that nematode and root densities decreased with depth, but they were both found as deep as 12 m on the playas, 6 m in arroyos, 3 m in dunes, and 2 m in the grassland and creosote sites. The study was more focused on nematode occurrence than rooting depth, but from this study the relationship between rooting depth and infiltration depth is apparent. Roots are deeper underneath arroyos than on the vegetated interfluves and hillslopes because of the increased infiltration under channels, and biological activity occurs at the greatest depth in the playas where recharge to the water table is documented (Snyder et al., 2006).

In summary, plants that persist in the Chihuahuan Desert are adapted to arid environments and make use of the available water. Shrubs have deep root systems which are even deeper under arroyos (Freckman and Virginia, 1989); however, little research exists on what becomes of the infiltration beneath small channels. If desert plants use all of the infiltration, we expect to find high chloride concentrations at the depths to which water typically percolates. If water percolates past the root zone of the plants that line the channel, the concentration of chloride will be reduced by the transport of dissolved chloride to the water table. If the wetting front is uniform, as in many unsaturated situations, the chloride is expected to be constant with depth. The presence of ponded water in the channel, however, and the resulting potential for saturated and preferential flow, greatly expands the range of profiles that might be observed. If preferential or lateral flow paths, which may be promoted by impervious caliche layers as well as ponding, develop, then the vertical chloride profile at a single location may show concentration increases and decreases with depth associated with two-dimensional flow paths.

1.3 Introduction to Chloride-Mass-Balance Method

A highly reliable way to determine whether or not recharge is occurring is by using the chloride-mass-balance (CMB) method. Atmospheric chloride, originating from sea salt, accumulates in the soil and soil moisture over time through wet and dry deposition (Phillips, 1994). While chloride is readily dissolved in water, it does not adsorb to soil particles (except under very low pH), and so it travels down through the soil with percolating precipitation. Because it is not volatile, it remains in the soil profile after the water that carried it has been transpired by plants. In the absence of recharge, chloride concentration in the soil increases as water delivers chloride from the surface (and the atmosphere) and the water is subsequently transpired (Phillips, 1994). In arid environments with negligible recharge, the concentration of chloride will become elevated near the base of, and just below, the rootzone. If a recharge event does occur, it will flush the accumulated chloride down to the water table (Phillips, 1994).

The time it takes for a given mass of chloride (Cl^-) to be accumulated within a given depth of a soil profile can be determined by dividing the Cl^- contained in the soil-water by the Cl^- deposition rate:

$$t_z = \int_0^z \frac{\theta C_{Cl}}{D_{Cl}} dz \qquad (4)$$

where t_z is the transport time of the soil-water chloride to depth z, θ is the volumetric water content, C_{Cl} is the chloride concentration that varies with depth, and D_{Cl} is the combined wet and dry deposition rate of chloride.

For the chloride-mass-balance method to be accurate, it is necessary that all the Cl^- originate from atmospheric deposition and not from mineral grains within the soil. Analysis of the bromide concentration enables evaluation of this assumption. Bromide is a conservative tracer that is derived from atmospheric sources and behaves much the same way as chloride (Davis et al., 1998), but it is more reactive than chloride and more susceptible to biochemical and geochemical cycling (Gerritse and George, 1988). Plants require halide elements for osmotic regulation and metabolic processes, and they tend to take up more bromide than chloride. This may cause the ratio of Cl^- to Br^- to be somewhat higher in the root zone than in precipitation (Gerritse and George, 1988). The Cl^- to Br^- ratio in the atmosphere usually ranges from 80 to 160, whereas the Cl^- to Br^- ratio for sedimentary sources is at least an order of magnitude higher (Davis et al., 1998). The ratio of Cl^-/Br^- in precipitation ranges from 50-180 (Davis et al. 1998). Chloride is usually 40 – 8000 times more abundant than bromine in natural waters, therefore the ratio of Cl^-/Br^- is drastically changed by small variations in bromine (Davis et al. 1998). If chloride concentration in soils are low then the bromide content is usually very low or undetectable. The highest Cl^-/Br^- ratios are found in evaporites. Halite, for example, can have ratios exceeding 100,000, and water affected by the dissolution of halite can have ratios ranging from 1,000 – 10,000 (Davis et al. 1998). Any ratio above 800 could signify that Cl is being added to the soil water by dissolution of evaporites.

The study reported here used CMB to look for evidence of focused recharge in channels on the piedmont slopes in an effort to corroborate the findings in Schreiner-McGraw and Vivoni (2017).

2. METHODS

2.1 Study Site Description

The study took place on the Jornada Experimental Range (JER), which was established in 1912 and has been a site for research and experimentation ever since (Havstad, 1996). Located about 20 km north of Las Cruces, New Mexico within the Chihuahuan Desert, the range is flanked on the east by the San Andres Mountains. The study watersheds drain the western piedmont slopes of these mountains (Figure 1). The core of the San Andres Mountains is made up of Precambrian and Tertiary igneous intrusive bodies (King and Hawley, 1975). Covering the Precambrian rocks are Paleozoic marine rocks, predominately limestone and dolomite, deposited in a shallow marine environment. The Tertiary sedimentary rocks consist of conglomerate, sandstone, and shale. The limestone, calcareous sandstone, and siltstone that outcrop in the San Andres Mountains break down into gravel or fine-sand-to-silt sized particles that are then deposited on the piedmont slopes. The piedmont-slope alluvium has a wide variation in textures and consists mainly of coalescent alluvial-fan surfaces with minor post-Santa-Fe-Group deposition in the middle-to-late Quaternary. The alluvium on the piedmont slope is derived from the San Andres Mountains and can be more than 1200 m thick, and the depth to groundwater ranges from 90 to 105 m (King and Hawley, 1975). The ephemeral channels carry runoff and sediment generated in the mountains onto the basin floor that was created by opening of the Rio Grande rift (Monger et al., 2006). There is very little organic matter in the soils of the piedmont slopes and little change in texture between the surface soil and subsoil. Calcareous dust has leached downward and deposited at the depth to which rainfall typically penetrates, most often between 5-75 cm. At this depth is a caliche layer (pedogenic calcite) that is frequently exposed in channels. Monger and others (2006) classified the soils of the piedmont slopes into four units: a clay or adobe soil with lime cementation; reddish aeolian sands called the Goldenberg sands that overlie alluvial strata; Jornada clay loam consisting of a range of textures and degrees of pedogenic carbonate; and gravelly soils that contain coarse rock detritus that is cemented into conglomerates by pedogenic calcite in some places.

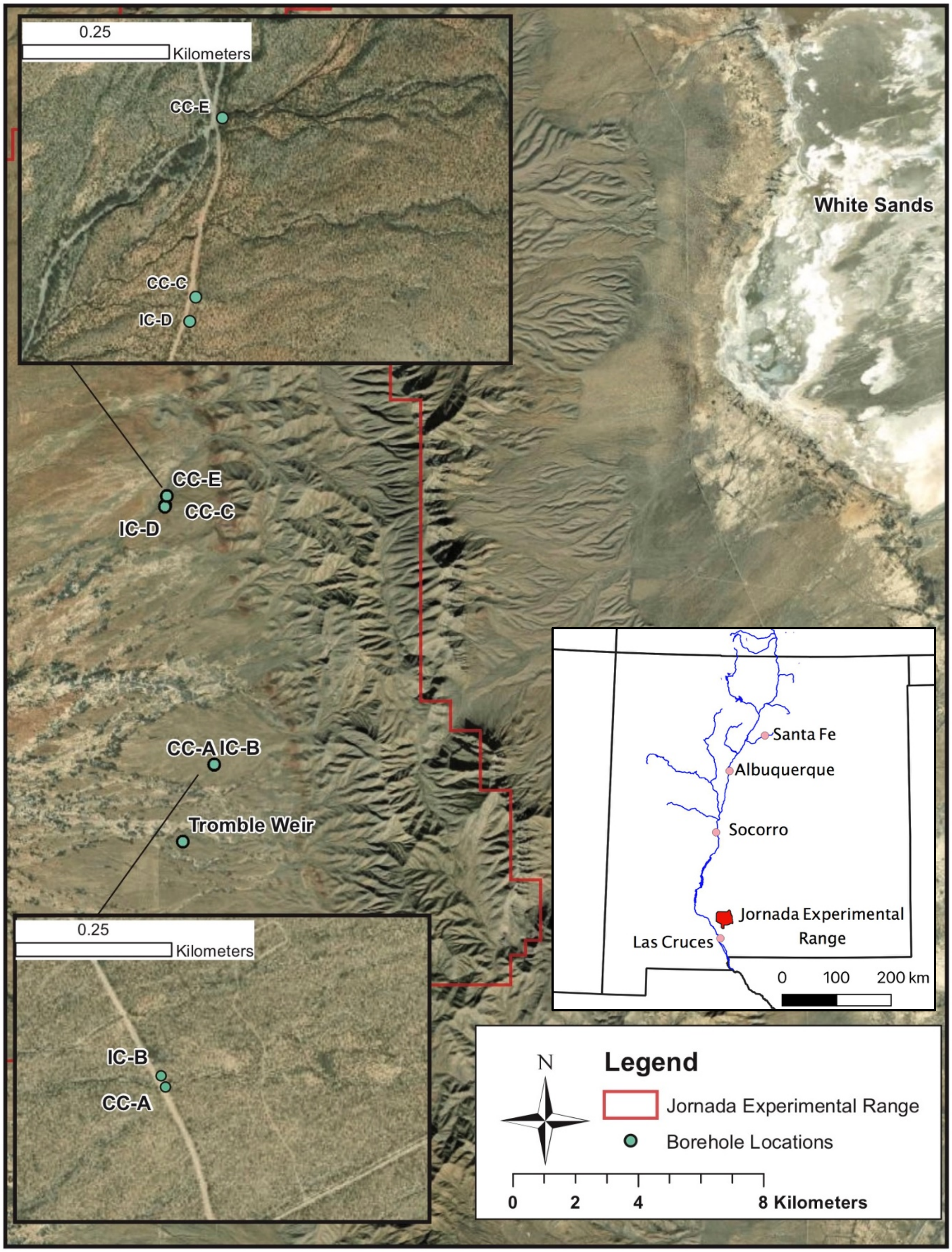

Figure 1. A satellite image of the study area, with air photo insets of the two coring areas (located at 32.6039°N 106.5984°W and 32.6665°N 106.6125°W). The point labeled Tromble Weir (located at 32.5853°N 106.6075°W) is the outlet of the watershed studied by Schreiner-McGraw and Vivoni (2017). The location of the Jornada Experimental Range within New Mexico is shown in the inset map on right.

Average rainfall in the JER is around 245 mm a year, with a standard deviation of 85 mm for 1915-1995 (Wainwright, 2006). Rainfall peaks between July and October, usually in August, due to monsoon storms that are derived from moisture coming off the Gulf of Mexico (Wainwright, 2006). During the time period of the Schreiner-McGraw study, from 2010-2016, the annual average rainfall was 257 mm and 70% of the precipitation occurred during July-September (monsoon season), which is slightly higher than the long-term average annual monsoon rainfall contribution of 50%. Winter rainfall is characterized by longer, lower intensity storms than the brief, intense monsoon rains. Temperatures are at their lowest in January (mean value 3.8°C) and highest in July (26°C) (Wainwright, 2006).

2.2 Drilling and Sampling Methods

In order to compare our results to the findings in the Schreiner-McGraw and Vivoni (2017) study, we chose to extract cores from channels that were similar to the channel at the outlet of their watershed, which is known as the Tromble weir and watershed, after an early researcher who conducted work at that site (Tromble, 1988). Two channels (A and C) were selected because their channels overlie the same soil type, have similar channel morphology, are surrounded by the same vegetation community, and are approximately the same size as the 0.047 km^2 Tromble watershed. Due to the small relief of these watersheds, adequately detailed topography is not captured by any existing topographic maps, so a drone equipped with a camera was flown over the watersheds to create orthophotos and digital elevation models (DEMs) using the structure-from-motion technique. The watershed areas were calculated from these DEMs. Watershed A was estimated to have an area of 0.08 km^2; however, this is a slight underestimate of the watershed area because the drone flight did not quite cover the entire watershed (Figure 2). Watershed C was estimated to have an area of 0.032 km^2 (Figure 3). Watershed A is 2.25 km north west of the Tromble watershed and Watershed C is a further 7.0 km north of Watershed A (Figure 1).

A hollow-stem-auger drill rig was hired to drill five soil cores, collecting core in 2-foot intervals with a split-spoon sampler. Two of the cores were taken from channels (Channel cores A and C), two from flat, interfluve locations near the drainage divides of the same watersheds (Interfluvial cores B and D), and one from a larger, cobble-bed channel draining the crest of the San Andres Mountains, and under which it is assumed focused recharge is occurring (Channel core E).

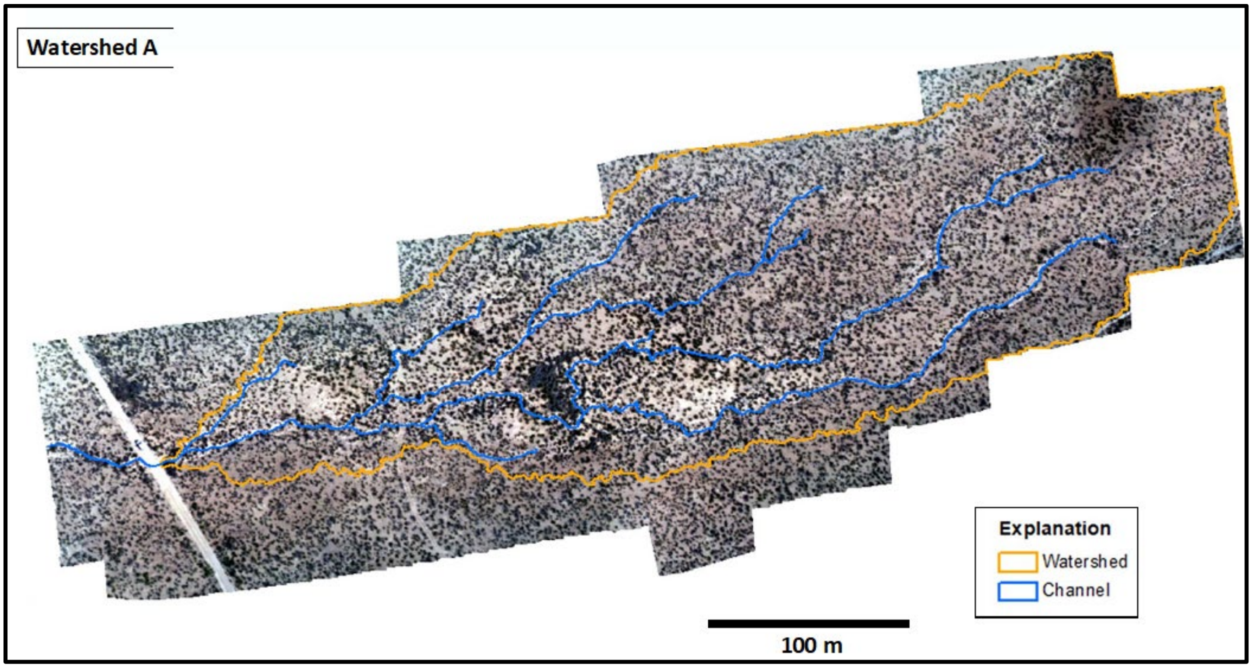

Figure 2. Orthophoto of watershed draining to coring site A. Watershed boundary based on DEM derived using the structure-from-motion technique. Flight did not fully cover the watershed area.

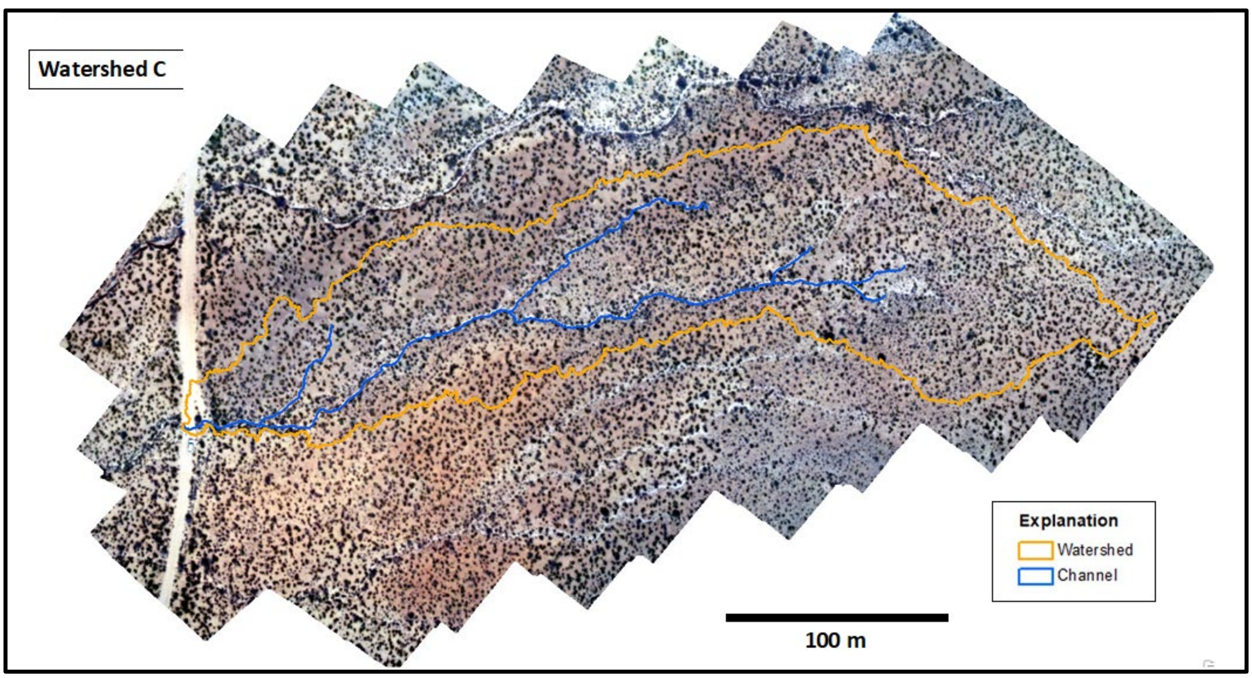

Figure 3. Orthophoto of watershed draining to coring site C. Watershed boundary based on DEM derived using the structure-from-motion technique.

The drilling procedure started with a split-spoon sample barrel being threaded through a hollow stem auger attached to a truck-mounted drill rig. The outer diameter of the split-spoon sampler was 2 inches, while the inner diameter of the auger was ~ 3 inches, meaning it fit easily but with the potential for a ring of sediment to exist between the sample barrel and the auger stem. A hydraulic hammer then advanced the split-spoon core barrel ahead two feet beyond the original position. The number of blows for each half-foot increment was recorded, and if the barrel did not advance half of a foot in 50 blows, the push was ended. The split-spoon sampler was then pulled from the borehole, opened, and sediment samples were extracted. The hollow stem auger was advanced to match the depth of the sample. A new split-spoon barrel was attached to the hammer, inserted into the auger stem, and pushed to the prior depth, which coincided with the new auger position. This new barrel was then advanced two feet ahead of the auger, and the process was repeated until the final depth was reached or the barrel could not be advanced.

The initial plan was to drill each core to a depth of five meters, but only three of the cores were successfully drilled to depth. The core from the large channel, Channel-E, was drilled to a depth of 3.81 m before hitting a boulder, but only 1.5 m of core was recovered because the sediment was loosely consolidated and much of the core fell out as it was being retrieved. This made it impossible to know the true depth of the returned samples, and because of this, profiles of the Channel-E samples are not included in this report. The soil moisture and chloride concentrations are only useful for qualitative interpretation. Both of the small-channel cores were drilled to five meters, as was one of the interfluvial cores (Interfluve-B), but the second interfluvial core, Interfluve-D, was obstructed by something the drill rig could not penetrate at a depth of two meters.

After each barrel was recovered and opened, the soil was removed and placed into plastic jars and capped with lids These were then placed into plastic bags and packed in ice in a cooler to minimize evaporation of soil moisture from the samples. Soil samples were taken every ten centimeters, with the first sample containing the sediment from the top (0 cm) to 10 cm depth, and the next sample from 10 to 20 cm. After the first 2 m, 10-cm-length soil samples were collected every 20 cm, for example, a sample from 200 - 210 cm was collected and the next sample was from 220 - 230 cm. The process of hammering in the split-spoon sampler caused the soil to compact. For example, advancing the bore hole two feet (~60 cm) may have only returned 50 cm of compacted sediment in the sampler. Therefore, the initial and final borehole depth was noted for each core barrel that was retrieved, and later the sediment thicknesses were converted to borehole depth in order to correct for compaction.

2.3 Water Potential

Water potential is influenced mainly by the solute content of the water (osmotic potential) and the capillary state of the soil-water system. In this study the solute concentrations were too low to affect the water potential significantly, making capillary state the main influence. The water potential of a saturated soil is zero and the potential becomes more negative as the soil dries. In a state of vertical hydrostatic equilibrium, the negative water potential at a point in the vadose zone will equal the magnitude of the positive height above the water table. Water potential can be expressed either in terms of pressure units or units of head (i.e., length). In this report we employ the pressure unit MPa for water potential.

We measured soil-water potential in each sample using a WP4 Soil Water Potential Meter (Decagon Inc.). The WP4 measures water potential by determining the relative humidity of the air above a sample in a sealed chamber using the chilled mirror method, in which relative humidity is determined by cooling a tiny mirror until dew starts to form. To take these measurements, we quickly removed approximately 2.5 cm^3 of soil from each sample jar, put it into one of the small containers provided with the WP4, capped the container, and then put it on top of the WP4 for 15 minutes so that the temperature could equilibrate with that of the machine. After removing the cap, we inserted the sample into the chamber and waited until the WP4 signaled to remove it (usually ~2-5 minutes). The reading from the instrument was recorded in a spreadsheet.

2.4 Gravimetric Water Content

We measured gravimetric water content following the procedure laid out in Methods of Soil Analysis, Part 1 (Klute and Page, 1986). The general procedure is to weigh a soil sample, dry it, and then reweigh the sample. The mass difference is attributed to water loss, and is divided by the dry soil mass to yield the water content of the soil in grams of water per gram of soil.

We dried the soil samples (entire sample) in an oven for 24 hours at 42.8°C (109°F) following Klute and Page (1986). To ensure samples were not losing mass past the 24-hour mark, we dried some samples for 48 hours, weighing them at hour 24 and hour 48. No difference in mass was observed between these times, confirming that 24 hours was sufficient to dry the sample completely. After drying, the soil samples were reweighed and then sieved to separate the coarse material from the material less than 2 mm in diameter (fines). Water loss was attributed only to the fine soil fraction because soil moisture is predominantly retained in the clay-to-sand fraction, not in the pebbles and cobbles. The gravimetric water content was determined by dividing the water loss by the mass of the fines.

2.5 Ion Chromatography

Chloride content was measured by ion chromatography in the analytical chemistry lab at the New Mexico Bureau of Geology. For most samples – those from cores B, C, D, and E – we put 50 grams of the sieved soil (finer than 2 mm) into leaching bottles and added 100 g of deionized water (18 MΩ resistivity). Initial test samples had indicated that solute concentrations in core Channel-A were lower than in the other cores. To ensure that solutes were above the detection limits of the ion chromatograph, we added only 75 g of 18 MΩ water to the soil samples from Channel-A. The bottles containing the soil and water were shaken for one hour using a shaker device with rotating arms, after which we put them into a centrifuge to settle the sediment to the bottom of the bottle. We then decanted the solution from the top of the bottle and vacuum filtered it using 47 µm nylon filters. We passed this filtered liquid through Pall Nylon 0.22 µm filters in a syringe before transferring it into vials for the ion chromatograph (Dionex ICS-5000). The chloride concentration was reported in Cl^- mg/H_2O kg. We calculated the mass of chloride contained in 50 g of soil as the chloride concentration detected by the ion chromatograph multiplied by the 0.1 kg of water added (or 0.075 kg for Channel-A). The pore water present in 50 g of soil was determined by multiplying the gravimetric water content by 50 g. Then the concentration of chloride in the pore water was determined by dividing the mass of chloride present in 50 g of soil (in units of mg) by the mass of pore water present in 50 g of soil (in units of kg). The chloride concentration is reported in mg/kg of pore water, based on the equation:

$$C_{Cl_p} = \frac{C_{Cl_l} M_w}{\theta_g M_s} \qquad (5)$$

where $C_{Cl\,p}$ is the chloride concentration in the soil pore water measured in Cl^- mg/H_2O kg, $C_{Cl\,l}$ is the chloride concentration in the leach solution, M_w is the mass of water used in the leach solution (typically 100 g), θ_g is the gravimetric water content of the soil measured in H_2O g/dry soil g, and M_s is the mass of soil that was leached (always 50 g).

2.6 Age of Accumulation

Soil-water chloride values can be used to determine the accumulation time of chloride in the soil water using Eq 4. In this study a D_{Cl} value of 60 mg m^{-2} yr^{-1} was used. The determination of D_{Cl} is explained in detail in Appendix A of Sandvig (2005).

3. RESULTS

3.1 Interfluvial Core B (Interfluve-B)

Water potential values for the interfluvial core Interfluve-B have a narrower range than either of the channel cores (Figure 4). Values range from −8.83 to −3.04 MPa, but there is no clear trend present in the water potential data. These tensions are equivalent to −900 to −310 m of head, and represent exceedingly dry soil. However, water content in the first 60 cm is higher than in the rest of the column (Figure 5). This indicates that despite nine days passing since the last rain prior to drilling, infiltration was still contained within the top 60 cm. The total range in gravimetric water content is from 0.01 to 0.06, a relatively narrow range, which indicates that, as expected, very little infiltration is happening, and markedly less than observed in the channel cores.

The chloride bulge in Interfluve-B starts below 61 cm (Figure 6), which is also where the gravimetric water content dips. The highest concentration is 14,000 mg/kg in the sample from 98 cm, and the second highest concentration is 11,600 mg/kg in the sample from 363 cm.

All Cl^-/ Br^- ratios are too low to be from sedimentary sources, and within the range typically associated with atmospheric sources (Figure 7). The three samples just below the surface have very low Cl^-/ Br^- ratios. This could be due to the near-surface soil being higher in organics, because organic rich environments preferentially retain Br^- over Cl^- (Davis et al., 1998).

3.2 Channel Core A (Channel-A)

The core in channel A was drilled in a unique position on the landscape. It was bored into the channel just upstream of a road crossing, which was necessitated by drill rig access considerations (Figure 8). On a trip to the site two days after a rain event, we noted that water pools where channel A intersects the road, which is the location of core Channel-A. By examining historic orthophotos available from the U.S. Geological Survey (USGS) website, the time of construction of the road was constrained to sometime between 1942 and 1971.

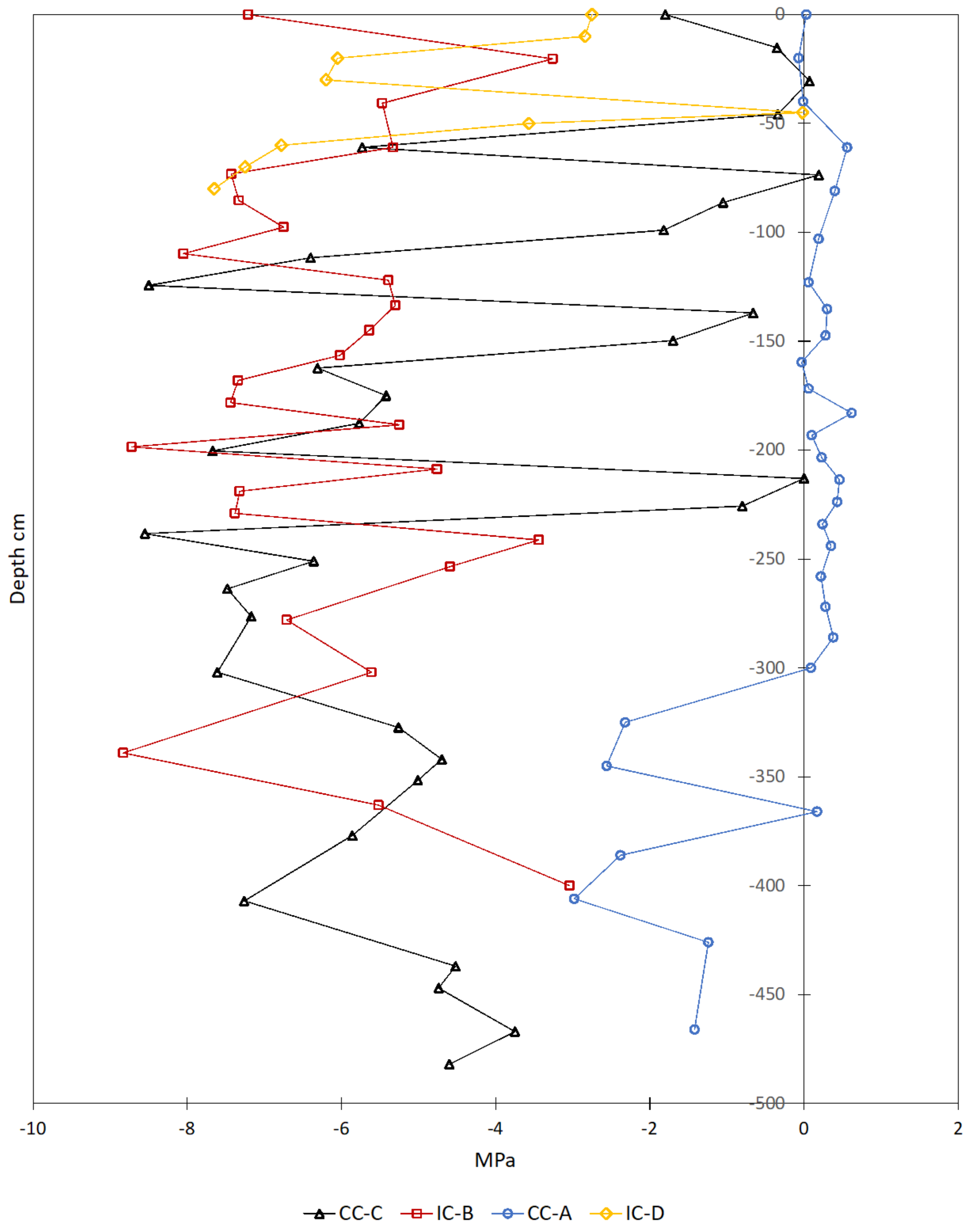

Figure 4. Soil water potential in MPa plotted against depth for the four cores. CC- indicates channel core, and IC- indicates interfluvial core.

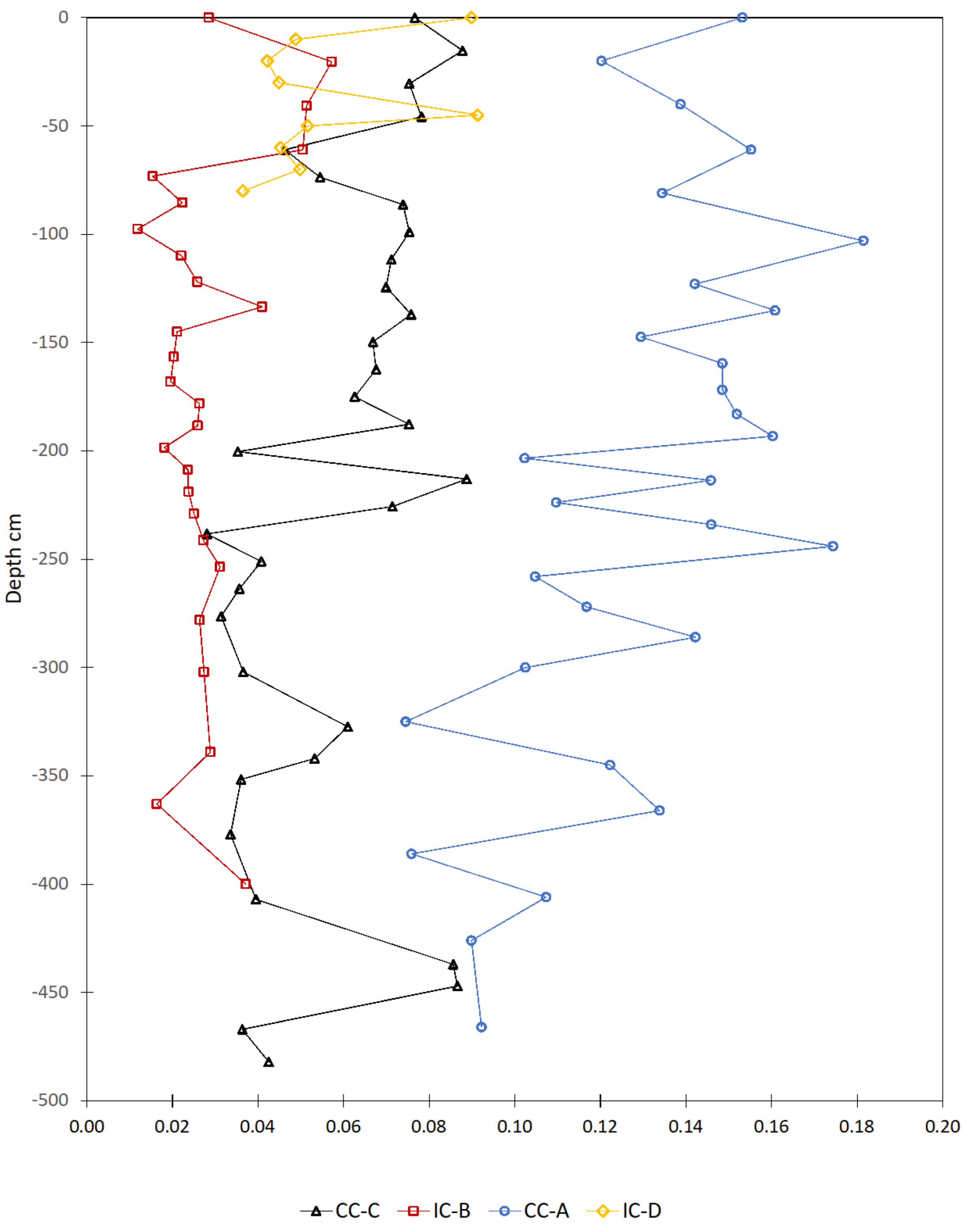

Figure 5. Gravimetric water content (unitless) plotted against depth for the four cores. CC- indicates channel core, and IC- indicated interfluvial core.

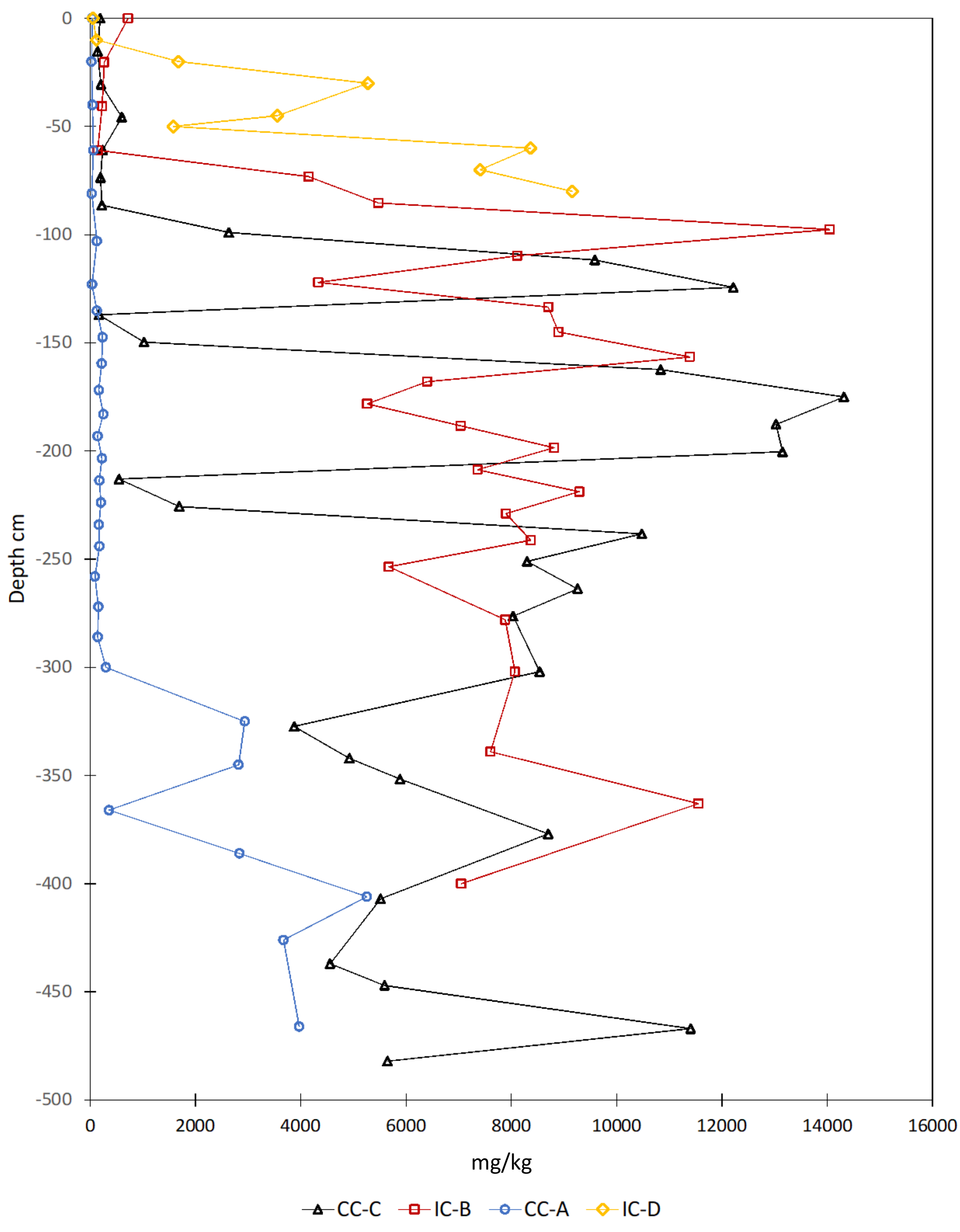

Figure 6. Chloride concentration in mg/kg of pore water plotted against depth for the four cores. CC- indicates channel core, and IC- indicated interfluvial core.

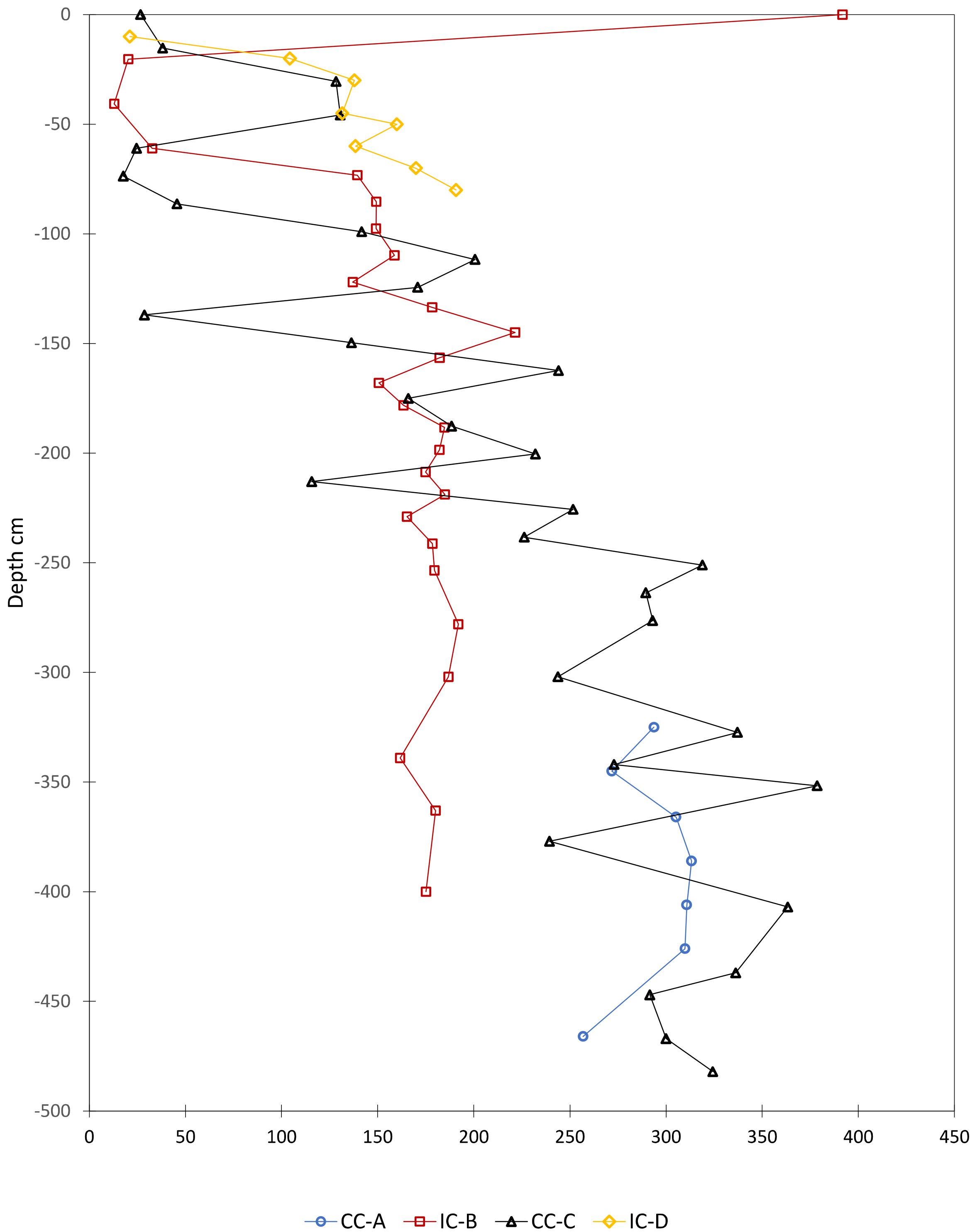

Figure 7. Ratio of chloride to bromide plotted against depth for the four cores. CC- indicates channel core, and IC- indicated interfluvial core.

Figure 8. Drill rig activity at core site Channel-A. The road slightly dams the channel at this location, meaning the core is being extracted from a local depression that collects water during runoff.

The first 200 cm of core samples from Channel-A were noticeably wet in the field. The day after extraction, all the samples from the first 300 cm had condensation visible on the inside of the sample jars when they were removed from the ice-chest in which they were retrieved from the field. The water-potentiometer was not designed for wet soils and has a range of −0.1 to −300 MPa (Decagon Devices, 2000). Pure water has a value of 0 MPa and any positive water potential values in Channel-A are errors caused by exceeding the calibration range of the instrument. Below 300 cm the water potential declines to approximately −2 MPa, indicating that infiltration is unlikely to be happening past 300 cm. The first 300 cm of core Channel-A contain very little chloride but at 300 cm the concentration increases from 300 mg/kg to 2900 mg/kg. The concentration increase aligns with the reduction in water potential at 300 cm, confirming that water has not recently infiltrated past 300 cm within this vertical profile.

The damming effect began sometime between 1942 and 1971 when the road was constructed, and the repeated ponding of water appears to have led to increased infiltration, pushing the chloride bulge farther into the soil (Figure 6). Given the > 50-year-long influence of this ponding effect, it is possible that the chloride bulge started at a higher position prior to the construction of the road, more similar to the bulge found in core Channel-C. It should be noted, though, that channel A has a larger drainage area than channel C (>0.08 km^2 vs. 0.032 km^2). A core extracted farther up channel A away from the road would have been more representative of the entire channel. The Cl^- peak of 5260 mg/kg at 406 cm likely does not represent the true peak of the chloride bulge, because only the top of the bulge was captured in core Channel-A. Our sample drilling stopped at 4.7 m depth, but the bulge appears to extend deeper, apparently due in part to the road ponding.

Bromide concentrations in the first 300 cm of soil core were below 0.10 mg/kg, outside of the calibration of the ion-chromatograph, and so Cl^-/ Br^- ratios are not reported for those samples (Figure 7). The last 166 cm of core had Cl^-/ Br^- ratios that are much too low to be from sedimentary sources, and slightly higher than typical ratios for precipitation. This could be because of the preferential uptake of Br^- over Cl^- of plants (Gerritse and George, 1988).

3.3 Channel Core C (Channel-C)

Water potential in core Channel-C ranged from 0.19 MPa (an erroneous value due to the limits of the water-potentiometer) to −8.55 MPa (Figure 4). Between 61 cm and 226 cm, water potential values alternate between high and low, and after 226 cm depth values remain low, although the last 100 cm of core have higher water potential than the 100 cm from above. The oscillations in water potential correspond with similar oscillations in chloride content, with potentials close to 0 MPa associated with low chloride concentration (Figures 4 and 6). These variations may be caused by any of the following: material falling from above when the core barrel was removed for sample extraction, heterogeneity of the soil texture, or because of a preferential lateral flow path that intersects the core at this location. Overall, matric potential and water content are lower than in Channel-A, a difference that likely is due to the increased infiltration of ponded water in Channel-A. If so, this suggests that the Channel-C core may be more representative of typical sub-channel hydraulic conditions in the study area.

3.4 Interfluvial Core D (Interfluve-D)

Between 10 cm and 20 cm, the concentration of Cl^- changes from 118 to 1674 mg/kg, and then the Cl^- concentration continues to increase to the bottom of the core at 80 cm. The gravimetric water content declines with depth as does the water potential. Very little infiltration appears to be occurring in this location. The core was only drilled less than 1 m depth because an object was encountered that the drill rig could not penetrate.

3.5 Channel Core E (Channel-E)

A core from a large arroyo just north of Channel-C was extracted, but because the sediment was not consolidated, most of the core was not recovered. The hole dug was 350 cm deep, but only 100 cm of core was recovered from the coring barrels. It is impossible to know the true depth of each sample; however, the chloride content in all samples was below 410 mg/kg and the 11 samples had an average Cl^- concentration of 115 mg/kg. All bromide values were below the reporting limit, the average water content was 0.13, and water potential values could not be determined because the samples were all too wet for the water potential meter.

4. DISCUSSION

Drilling took place on October 21, 2019. Multiple rain events were recorded on the two nearest rain gauges operated by the Jornada Experimental Range staff within the preceding month (gauge tank PK and gauge tank CT). One gauge recorded a precipitation event of 7.1 mm on October 9, 2019, and the other recorded 4.3 mm of rainfall that day. A more notable rain event occurred on October 4 (10.9 mm and 5.6 mm respectively) and continued into October 5 (23.9 mm and 24.9 mm respectively). These precipitation events are likely the reason that every core has a higher gravimetric water content near the surface than at depth. Hydraulic lift, where deeply rooted plants like shrubs bring up water from depth and then exude that water into upper, drier, soil layers at night when transpiration stops (Caldwell et al., 1998), could conceivably also help to maintain water near the surface. The redistribution of water by deeply rooted plants provides water for shallow rooted plants, and inhibits percolation to the water table.

None of the soil cores were dug deep enough to capture the entire chloride bulge; therefore, it is only possible to determine the lower boundary of the age of accumulation. Equation 4 was developed for flat areas, not channel beds. Using this equation for channel beds results in an older age of accumulation than the true age because the deposition rate of chloride, D_{Cl} in Eq. 3, is dependent on the depth of precipitation, and while a channel experiences as much rainfall as the surrounding land surface, it also experiences run-on. The additional water delivered to the channel brings with it additional chloride, making the deposition term used for flat areas a lower bound of the true value in channels. The chloride concentration of runoff is probably similar to the concentration of rainfall, but we do not have data to confirm this assumption. To determine more accurately the accumulation time in the channel cores, it would be necessary to determine a modified D_{Cl} based on the amount of runoff, or more specifically infiltration, a channel receives.

One notable weakness of this study is that a soil core only represents a point measurement. If preferential flow is occurring, there is a possibility that a soil core could misrepresent the chloride concentration profile of an area. This is less of a problem for unsaturated flow. A trench study of unsaturated infiltration in the Jornada Range using a drip irrigation system (Wierenga et al., 1991) found no evidence of preferential flow through the soil on the piedmont slopes. Wierenga and others (1991) specifically looked for signs of water flowing through root cracks or soil layering affecting water movement, but could find evidence of neither. They found that the overall location of the wetting front was well predicted by a single layer, one-dimensional model, despite the spatial variations in the soil and changes in the hydraulic conductivity. These observations may have been influenced by the nature of the water application to their study site. Water was applied to the 4 m x 9 m area through sixty irrigation lines each containing 13 emitters, for 17 minutes a day for an average flux of 1.82 cm/day for 86 days. This low intensity wetting is dissimilar to runoff in channels that may cause saturated preferential flow. Preferential flow is most likely to occur when there is ponding (Hendrickx and Flury, 2001). Scanlon and others (1997) found that preferential unsaturated flow beneath large surface fissures in the Chihuahuan Desert in Trans-Pecos Texas extended to 10-20 m depth, based on the results of a water potential and chloride analysis. The degree to which preferential flow occurred depended on ponding and topographic variations within the fissure. Hence, the ponded flow in our study channels is likely to cause more preferential subsurface flow than found by Wierenga and others (1991).

The chloride profile from core Channel-C has two sharp drops in chloride that coincide with increased water content at 137 cm depth (samples C100 and C110) and 213 cm depth (samples C160 and C170). One possible explanation for these features is two-dimensional preferential flow, for example along the top of an impervious caliche layer. If true, this would limit the interpretability of these cores for understanding recharge.

An alternative explanation is that these discursions represent sediment from higher in the profile that fell into the core hole between pushes. Both of these segments of low Cl^- concentration occur at the tops of retrieved core barrels (Figure 9). The four samples in question (C100, C110, C160, and C170) were red in color and friable in texture, while the rest of the samples from their 2-foot-long barrels were white and compacted. These observations do not preclude the samples from being distinctive sedimentary layers that were acting as conduits for lateral flow, but their position in the cores would be highly coincidental. The nature of our coring methods makes such out-of-sequence samples more likely. The hollow stem of the auger was ~3 inches in diameter, while the slit-spoon barrels were 2 inches. The anulus of material between them could be displaced during advancement of the auger, which took place when the core barrel was removed.

If this dropping in of higher material was common, it would introduce noise to our chloride and moisture profiles. We carefully reviewed the data, and did not observe other sections of core where a pattern of deviations from the trend occurred so clearly at the tops of the core barrel samples. But this raises concerns about using this precise sampling method in the future.

Figure 9. Photos of the first four core barrels extracted from Channel C. The depths used to label each core are the uncorrected depths, which do not match perfectly with Figures 4-7. The black braces indicate portions of C100-160 and C160-230 that have low chloride and may be sediment fallen from above.

Not many studies on chloride concentrations of small channels exist, but Scanlon (1991) conducted a study of chloride profiles in the Hueco Bolson, and found chloride bulges that peaked between 1.3 and 4.6 m depth. She found the highest Cl^- concentration in a small ephemeral stream (0.6 m of vertical change in topography at banks). The chloride concentration gradient was up to 12,000 g m^{-3} m^{-1}. This steep gradient supports the idea that plants were able to utilize all infiltrated water at that location. She also discussed how very low water potentials limit preferential flow because water is adsorbed onto grain surfaces and is not free to move along larger openings or root channels, except possibly during extreme storms.

Tracking vadose zone water fluxes in arid environments with deep water tables is quite complex. The relationships between moisture content, water potential, and hydraulic conductivity are not linear, and this results in large uncertainties in flux calculations. The added complexity of flow in arid environments makes Darcy's Law less useful because of the liquid and vapor transport that occurs in response to water potential and temperature gradients (Scanlon, 1991). Many reviews of methods used to track fluxes expound on the difficulty in calculating fluxes in arid environments because of the uncertainty in ET estimates (Allison et al., 1994; De Vries and Simmers, 2002; Gee and Hillel, 1988; Scanlon et al., 2002). In the rare cases where recharge is happening, it is a very small part of the total water balance, meaning any small errors in the other components result in very large errors in the recharge estimate. Chemical tracer studies provide the clearest evidence of recharge, but, because a concentration profile is more or less a point measurement, they can miss evidence of recharge in environments where preferential flow and lateral flow is a mechanism for significant recharge. However, preferential flow is a more prevalent process in wet regions. In arid areas where soils are very dry, preferential flow is less likely (Scanlon, 1991). Scanlon (1991) found that water potentials in the Hueco Bolson near El Paso, Texas range from -2 to -16 MPa and because of this, most water is adsorbed onto grain surfaces at depth and not free to move along larger openings or root channels except possibly during extreme storms. In the piedmont slopes of the JER, the water table is 90-105 m below the surface (King and Hawley, 1975), and the water potentials in the cores were as low as −8.83 MPa, which is −900 m of head. These lines of evidence suggest that preferential flow is not generally a dominant process in our study area. But the possibility remains that focused flow in small channels is a localized exception to this generalization.

We initially expected the Cl^-/Br^- ratio values to be more consistent and closer to the ratios found in precipitation, but only because we did not account for the preferential uptake of bromide by plants. Gerrtise and George (1988) found that Cl^-/Br^- values in rootzones are elevated because of the uptake of Br^- by plants, and that decomposing plants release bromide at the surface thereby decreasing the Cl^-/Br^- ratio. The general cycle is for plants to take up Br^- at depth and then redeposit it onto the surface when they die and decompose. This helps explain why the bromide ratios are low at the surface and increase with depth, except in Interfluve-B, where ratios are low at the surface and then constant at depth. This exception is probably due to the limited infiltration in the interfluve, leading to less uptake of Br^- at depth allowing for more consistent ratios.

The water balance study in the Tromble Weir catchment that inspired this research (Schreiner-McGraw et al., 2016; Schreiner-McGraw et al. 2017) made many efforts to quantify error and even assumed a 100% error in the discharge estimate in order to come up with a conservative estimate of percolation. To estimate the error associated with ET, they applied a method by Twine and others (2000) using the Bowen ratio, the ratio of sensible heat flux to latent heat flux ($B = H/\lambda ET$), and found a closure error of 19%. By forcing closure, they achieved a higher ET estimate. They also considered a scenario where the streamflow had an error of 100% in order to achieve a conservative percolation estimate, and still came up with a positive value (recharge) for deep percolation. However, we suggest that it is still possible that the ET was underestimated, because the footprint of the EC tower only captured a small percentage (~ 10%) of the total watershed, and this was in the headwaters, where run-on would be lowest (Figure 10). Within the watershed are clusters of dense vegetation, especially alongside the small channels, which are readily identified from air photos and observed in person. The EC footprint does not capture these areas of highest productivity, and thus might be missing a significant portion of the ET. The EC tower was also placed in a flat area at the top of the basin, likely because EC towers work best when there are no large irregularities in topography. However, the radiation on a flat surface does not accurately reflect the radiation on hillslopes facing different directions. Therefore, it would be informative to compare the effective radiation flux within the watershed to the radiation in the EC footprint. Finally, we suggest that the rooting depth considered may have been too shallow. Percolation below the soil moisture sensors remained at depths accessible to plants, making transpiration an alternative pathway to percolation down to the water table.

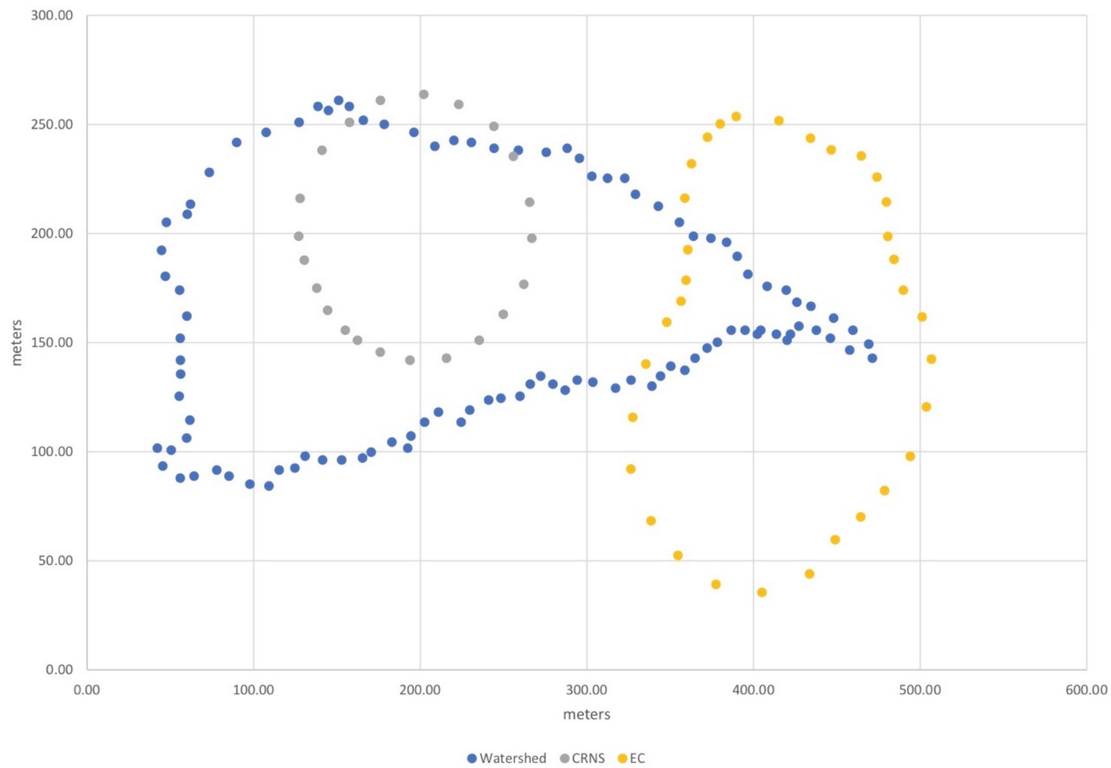

Figure 10. Instrument footprints at the Tromble Weir watershed. Blue dots outline the watershed (headwaters in the east), gray dots outline the estimated footprint of the cosmic ray neutron sensor used to estimate soil moisture, and the yellow dots outline the estimated footprint of the eddy covariance tower.

Schreiner-McGraw and Vivoni (2017) noted a lysimeter study in the JER (Levitt et al., 1999) that found no deep percolation in soils with natural vegetation, and a review on vadose-zone techniques for estimating groundwater recharge in arid and semiarid regions (Allison et al., 1994) that found that physical approaches like water balance and Darcy Flux measurements are the least successful methods of estimating recharge, and that environmental tracers are the most successful.

An inspection of aerial photos of the entire study area (western piedmont slopes of the San Andres Mountains) clearly reveals vegetation that is denser around depressions and channels. There are both more and larger plants along channels than on hilltops. Therefore, despite the increased infiltration that occurs in channels, it is reasonable to conclude that the increased infiltration leads to increased transpiration, and not to recharge.

5. CONCLUSION

Our analysis of chloride profiles extracted from sediment cores from below two small channels on the piedmont slopes of the San Andres Mountains suggests that infiltration leading to aquifer recharge is unlikely in this setting, but uncertainties regarding saturated preferential flow and issues that arose during sampling limit firm conclusions. One channel was locally dammed by a road crossing at the sample site, leading to enhanced infiltration and limited representativeness of the broader channel network. However, even at this site with apparently much higher infiltration, a chloride bulge remains within 4.7 m of the land surface, qualitatively indicating that percolating water does not flush chloride to the water table. In the other channel sampled, two short intervals of high moisture content and low chloride content within the profile complicate interpretation. These could indicate preferential lateral flow, which can produce recharge even in the presence of a sampled chloride profile bulge. But the locations of these low-chloride intervals at the tops of individual core barrels of returned sample suggest that they may be sediment from higher in the profile that fell into the hole between samples, highlighting a potential weakness and source of noise in our sampling method. It is clear that there is increased infiltration beneath channels relative to the interfluvial portions of the landscape, but the relatively dense plants lining the channels may transpire this infiltrated water before it reaches the water table. These two channel boreholes give a limited but informative picture of the unsaturated hydrology under channels. One promising way to explore more deeply the question of channel infiltration would be to conduct another similar experiment but with numerous cores extracted from each channel, and extending to slightly greater depths.

Given this preliminary lack of evidence for small channel percolation, especially when some evidence would be expected if it were a significant source of groundwater recharge, we recommend that water planners should not begin to make plans that rely on piedmont small-channel recharge as a source of water in the future. Over-estimating such recharge could lead to water users making financial decisions that rely on a source of water that is not actually available, and more work is needed to better estimate recharge rates. This study does not support the hypothesis that small channels are contributing to groundwater recharge; however, it does not eliminate the possibility that preferential lateral flows bypass portions of the vadose zone. Localized hotspots of infiltration and recharge may also exist, although at the one location of enhanced infiltration that we sampled, recharge would still require that preferential lateral flows bypass the chloride we sampled. In order to better understand the fate of percolation in these small channels, we recommend: (1) systematic sampling away from anthropogenic infrastructure such as roads; and (2) deeper coring, beyond the 5-m-depths analyzed in this research. Deeper drilling and a greater number of cores would incur significant research cost. In the absence of further evidence, the two cores analyzed in this report suggest that societally meaningful recharge is unlikely beneath these types of channels.

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