



# The role of interbasin groundwater transfers in geologically complex terranes, demonstrated by the Great Basin in the western United States

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**Abstract** In the Great Basin, USA, bedrock interbasin flow is conceptualized as the mechanism by which large groundwater fluxes flow through multiple basins and intervening mountains. Interbasin flow is propounded based on: (1) water budget imbalances, (2) potential differences between basins, (3) stable isotope evidence, and (4) modeling studies. However, water budgets are too imprecise to discern interbasin transfers and potential differences may exist with or without interbasin fluxes. Potentiometric maps are dependent on conceptual underpinnings, leading to possible false inferences regarding interbasin transfers. Isotopic evidence is prone to non-unique interpretation and may be confounded by the effects of climate change. Structural and stratigraphic considerations in a geologically complex region like the Great Basin should produce compartmentalization, where increasing aquifer size increases the odds of segmentation along a given flow path. Initial conceptual hypotheses should explain flow with local recharge and short flow paths. Where bedrock interbasin flow is suspected, it is most likely controlled by diversion of water into the damage zones of normal faults, where fault cores act as barriers. Large-scale bedrock interbasin flow where fluxes must transect multiple basins, ranges, and faults at high angles should be the conceptual model of last resort.

**Keywords** Groundwater recharge/water budget · Carbonate rocks · Conceptual models · USA · Groundwater flow

## Introduction

The importance of groundwater resources can hardly be overstated, so it follows that a conceptual understanding

of its movement is of great interest and consequence (e.g., Gillespie et al. 2012; Welch et al. 2007; Wilson and Guan 2004). For example, the spatial dimensions of an aquifer greatly affect the availability of water for development as the volume of water in storage scales like the cube of linear aquifer dimensions. Alternatively, simulations of groundwater movement may vary greatly depending on whether a model domain boundary is defined to allow flow across it, and such boundaries are defined by how the system is conceived by a modeler. In essence, a conceptual model largely predetermines the outcome of accompanying quantitative studies of groundwater movement. Interbasin flow is one conceptualization that can affect the subsequent evaluation of a flow system in terms of aquifer dimensions, model boundaries, groundwater residence times, and aquifer sustainability relative to withdrawals.

This report focuses on the so-called Great Basin carbonate and alluvial system (GBCAAS) in western USA (Heilwell and Brooks 2011) where bedrock interbasin flow has been propounded by numerous researchers (discussed in the following) over a vast area (Fig. 1). The Great Basin is the largest contiguous endorheic watershed in North America covering 500,000 km<sup>2</sup>, much of which is underlain by carbonate rock. In this context, one goal of this report is to demonstrate, by a review of previous research in the Great Basin, that large interbasin transfers may be less common than previously thought.

The Great Basin is an unusual hydrogeologic setting. It is a broadly extended region with numerous basins and intervening ranges (Fig. 1). Nonetheless, much of the discussion surrounding bedrock interbasin flow is relevant in a variety of hydrogeological settings where the role of aquifer discontinuities may impact regional flow paths.

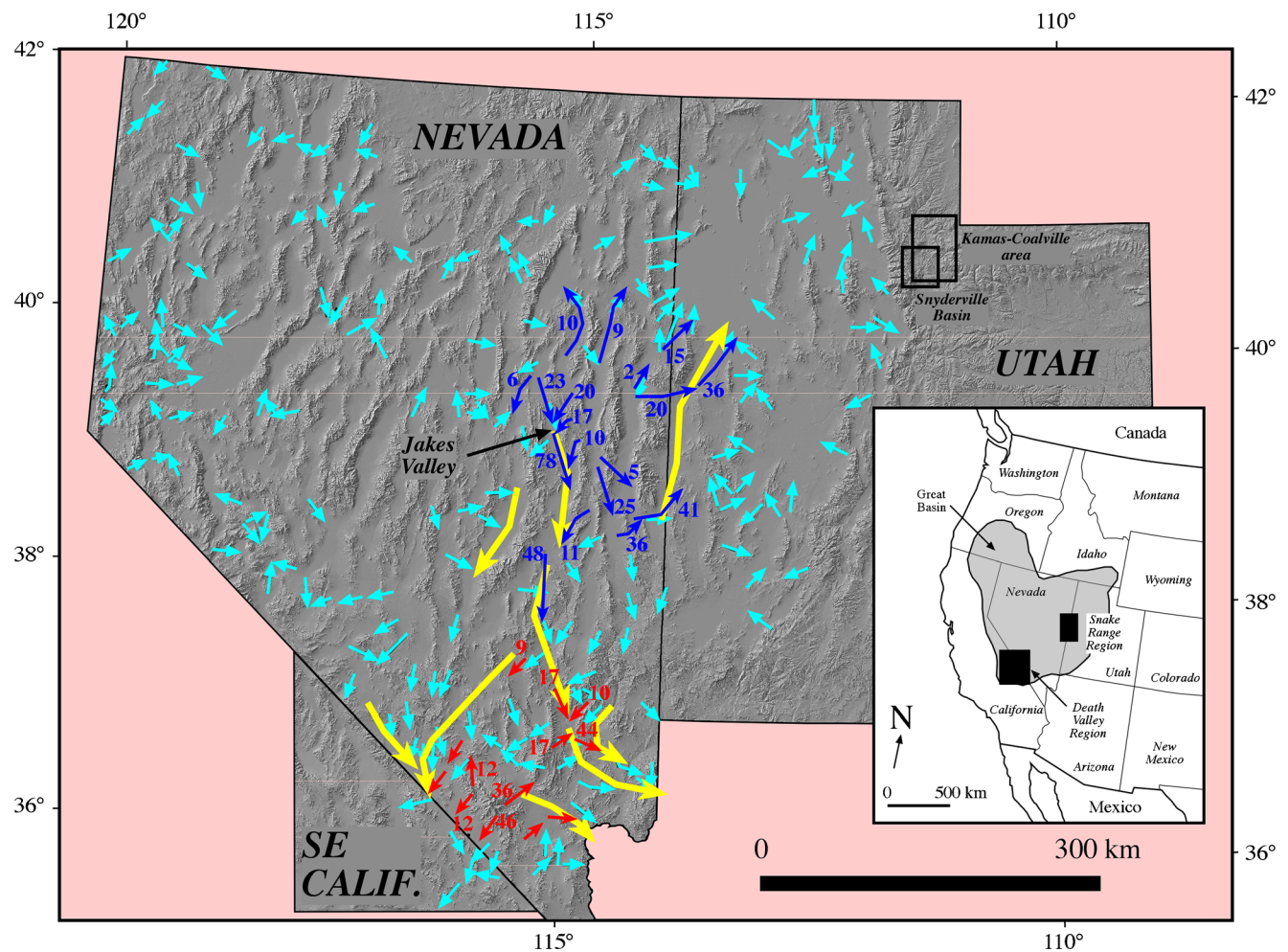
## Purpose

A number of recent studies conducted by the authors and their students (Nelson et al. 2004, 2005; Anderson et al. 2006; Miner et al. 2007; Nelson et al. 2009; Bushman et al. 2010; Gillespie et al. 2012) have called into question the nature and extent of bedrock interbasin flow (hereafter simply ‘interbasin flow’) in the western USA as it has traditionally been envisaged. This has led to a re-assessment

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**Fig. 1** Index map for regional groundwater systems within the western USA discussed in this study. Inset shows the location of the *Great Basin* in the western USA, with locations indicated for the *Death Valley* and *Snake Range* regions discussed in the text. The *larger map* shows the physiography of the *Great Basin* in Utah, Nevada, and southeastern California, with selected interbasin flow paths and fluxes as indicated. *Cyan arrows* indicate posited locations of interbasin transfers from Harrill et al. (1988). *Red arrows* show the flow paths of Thomas et al. (1996), and *dark blue arrows* show flow paths of Welch et al. (2007). Corresponding *colored numbers* indicate proposed fluxes in  $10^6 \text{ m}^3/\text{yr}$ . *Yellow arrows* indicate flow paths from Hershey et al. (2010). *Snyderville Basin* and *Kamas-Coalville area* are discussed in the text

of interbasin flow in this region, and as may be applicable to other regions. In summary, these studies indicate that large-scale interbasin fluxes are less common than traditionally posited, and where they occur, interbasin flow is along tectonically induced, fracture-controlled permeability structures. The purpose of this report is to review the features and processes illustrated by these studies, to contrast previous work, and to reach conclusions regarding how and where interbasin flow operates. The intent is to advance the understanding of bedrock interbasin flow processes through a review of prior studies.

### Hydrogeological setting

The unique geological character of the *Great Basin* warrants a brief review to provide the proper context before presenting a detailed definition of interbasin flow and examining specific flow systems. Detailed hydrogeological framework summaries relevant to interbasin flow in this area are available to the interested reader

(e.g., Welch et al. 2007; Sweetkind et al. 2011a; Heilweil and Brooks 2011; Gillespie et al. 2012).

The geologic history of the *Great Basin* relevant to aquifer formation and geometry can be summarized in three main phases: (1) late-Precambrian to middle-Paleozoic deposition of ~9,200 m of carbonate rock with minor interbedded siliciclastic units deposited along a passive continental margin (e.g., Plume 1996); (2) episodic Devonian to Eocene crustal compression, resulting in regional-scale folding, crustal thickening, metamorphism and emplacement of plutons (e.g., Gans et al. 1985); and (3) Cenozoic extensional faulting accompanied by volcanism and valley-fill sedimentation, producing modern Basin and Range topography (e.g., Welch et al. 2007) (Fig. 1).

Elongate north-south-trending mountain blocks with intervening grabens characterize typical Basin and Range topography. Horst-graben or half-graben sets are usually bounded by major normal faults, with grabens often exhibiting internal drainage and acting as sediment traps for material shed from adjacent ranges. The basins have surface

elevations that range from below sea level to ~2,000 m, whereas ranges may have elevations as high as 4,000 m. Typical Basin and Range topography is illustrated in Fig. 1.

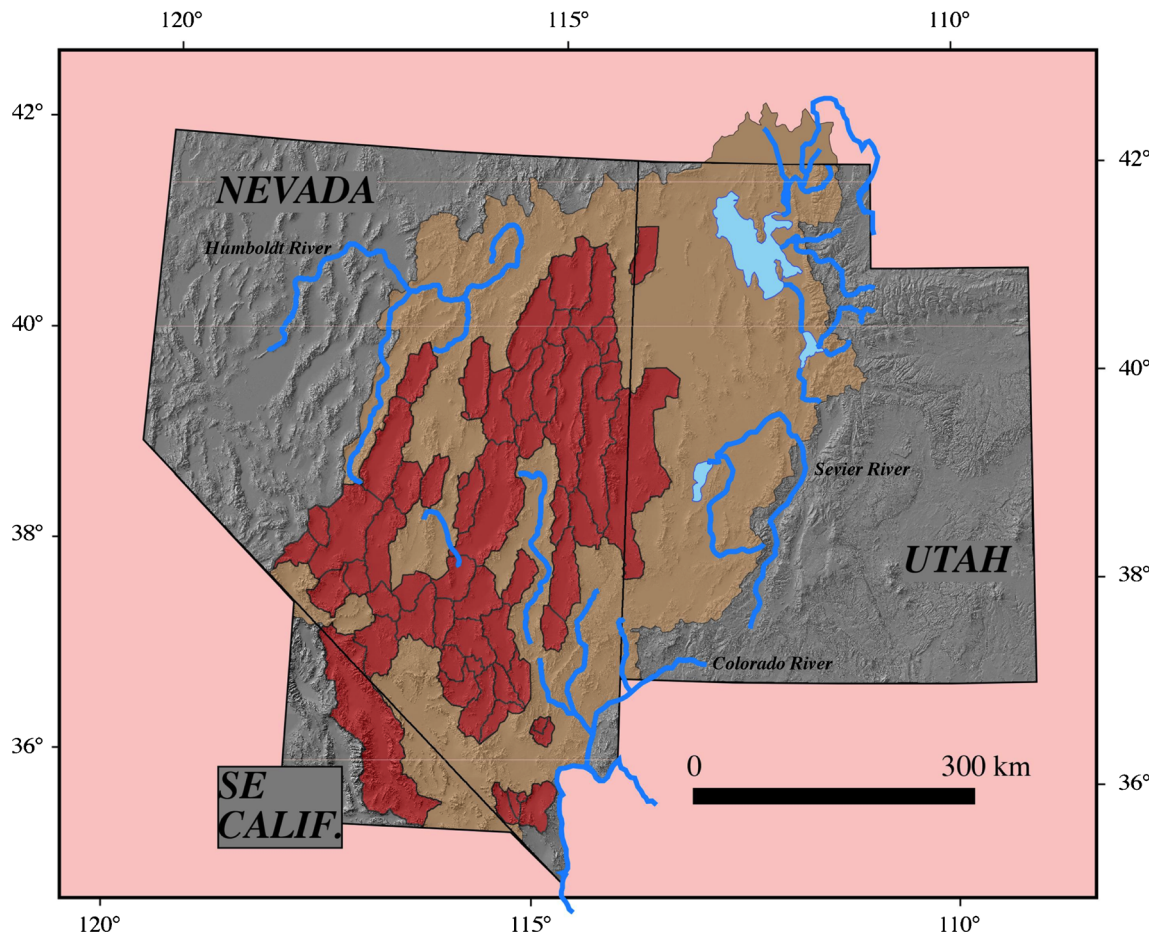
As framework for this study, water budgets and flow system definitions have been employed as reported by Heilweil and Brooks (2011), who divided the GBCAAS into 164 individual topographic basins (hydrographic areas, or HAs) assigned to 17 regional groundwater flow systems. Individual HAs range in size from 31 to 12,038 km<sup>2</sup>, where individual basins have been defined by basin bounding mountains and their surface drainage areas. Many of the HAs are interconnected by alluvial-filled channels, including perennial interconnecting streams and major intermittent drainages, so the focus of this investigation is on interbasin flow within 57 endorheic HAs where interbasin flow must be restricted and forced through mountain blocks (Fig. 2). The remaining 107 HAs were excluded due to the potential for interbasin flow being accommodated by underflow in alluvium.

### Definition of interbasin flow

We use the term interbasin flow in the same sense as Gillespie et al. (2012) and similar to that of Sweetkind et

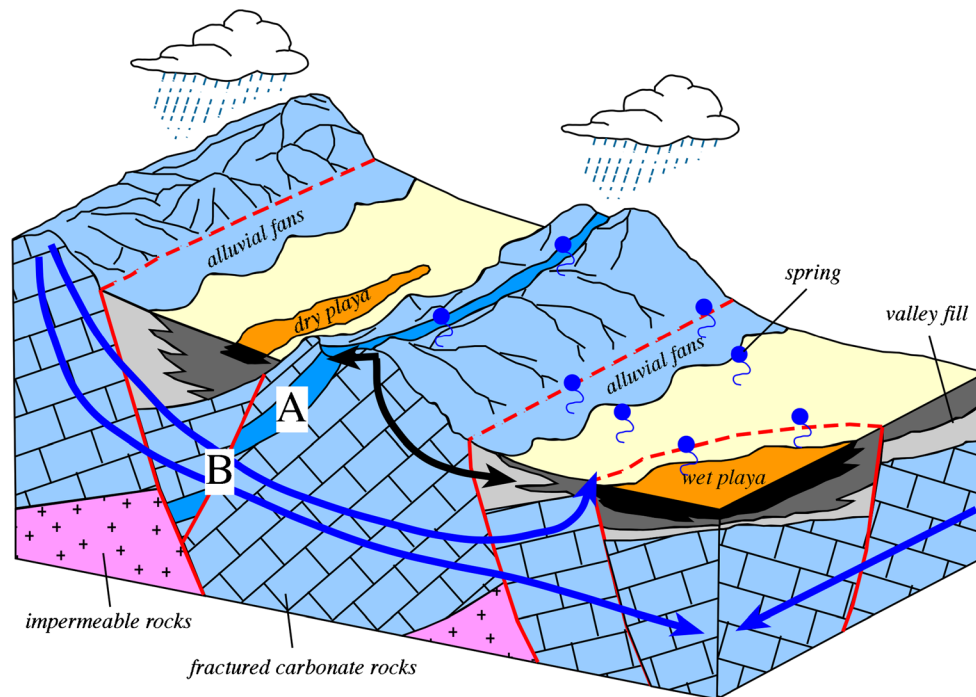
al. (2011b), where it has been applied to groundwater flow throughout much of the GBCAAS (Fig. 2). A conceptual illustration of interbasin flow is provided in Fig. 3. Briefly, it is an extension of the classic studies by Tóth (1963); Freeze and Witherspoon (1967), and others where “regional” flow systems can move water between separate topographic basins, but as discussed here, it applies in particular (although not exclusively) to fractured carbonate rocks.

In the GBCAAS (Fig. 2), this idea has long been adapted to describe the subsurface movement of groundwater in fractured carbonate bedrock between basins and through intervening ranges (Fig. 3). Flow is posited as long as the carbonate rocks are continuous and a potentiometric gradient exists (e.g., Winograd 1962; Eakin and Moore 1964; Eakin 1966; Maxey and Mifflin 1966; Sweetkind et al. 2011b). Harrill and Prudic (1998) suggested that carbonate aquifers have high horizontal hydraulic conductivities caused by fractures and joints that have been widened or sustained by dissolution (Plume 1996). Others have expanded the view of interbasin flow to include transfers between carbonate and non-carbonate rocks (Davisson et al. 1999; Rose and Davisson 2003). Naturally, there is no reason this should not occur as long



**Fig. 2** Map showing the extent of the Great Basin carbonate and alluvial aquifer system (GBCASS), including 57 endorheic basins (*red*). *Brown areas* represent aggregate basin systems that receive or deliver surface water to other basins. Major stream systems are indicated in *dark blue* and lakes in *light blue*





**Fig. 3** Cartoon illustrating conceptualized components of groundwater flow in the western USA similar to the Snake Valley system. Adapted from Welch et al. (2007). *A* and *B* represent potential flow barriers (impermeable bed and fault core, respectively). Blue dots with tails represent springs, whereas black and blue arrows represent local and regional flow paths, respectively. See text for discussion

as their permeability structures are integrated. Overall, past studies have suggested flow paths in the Great Basin that are tens of kilometers up to 300 km in length (e.g., Rose and Davison 2003), in which case underflow may occur through multiple basins and ranges.

### Evidences of interbasin flow

Sweetkind et al. (2011b; p. 52 and references therein) provide an excellent review of past lines of evidence for interbasin flow, including: (1) groundwater budget imbalances or a lack of discharge within a given basin, (2) isotopic studies, (3) structural and potentiometric data, and (4) modeling studies. Although an evaluation was made of most of this evidence, there is no direct addressing of modeling studies. As valuable as models may be, their support of interbasin flow is largely circular. If a model is constructed to permit interbasin flow, interbasin fluxes are nearly guaranteed when the model is run; therefore, numerical models are not addressed further.

It is beyond the scope of this report to re-evaluate each study cited by Sweetkind et al. (2011b). Although reference is made to many of the studies that have invoked interbasin flow, the number of papers and reports is so numerous it is not practical or beneficial to discuss them all. Interbasin flow truly has been, in many respects, the default paradigm in the GBCAAS in the Great Basin of the USA. Rather, discussion is based on specific locations where the authors have worked and where a re-evaluation of available evidence leads to alternative, and in the authors' view, better conclusions regarding the flow systems. In some cases, the evidence against interbasin

flow comprises another category of data than that presented in its favor. For example, water budget evidence may be inferred to be imprecise on the basis of isotopic studies.

Interbasin transfers, like all subsurface flows, depend on the existence of sufficient aquifer permeability combined with differences in water potentials (Sweetkind et al. 2011b). The aquifer system must also lack flow barriers such as intervening impermeable interbeds in bedrock or fault cores, and in the case of carbonate rocks, fractures that have not been sealed by secondary calcite or other minerals. In this context, the issue of scale obviously becomes critical. At sufficiently large spatial dimensions, the probability that potential, permeability, and a lack of flow barriers are all present without aquifer segmentation becomes increasingly remote. Logic suggests that underflow between adjacent basins separated by a single mountain range is much more likely than between basins that are separated by multiple basins and ranges.

### Magnitudes of proposed interbasin fluxes

If interbasin flow were purported to be a trivial fraction of total groundwater fluxes, its importance might be diminished, perhaps to the point it could be ignored. For example, Heilweil and Brooks (2011) recently contended that "most" groundwater in the GBCAAS was considered to discharge in valleys adjacent to mountainous recharge areas, and their water budgets did not quantitatively estimate interbasin transfers. However, this conclusion is a great departure from prior studies.



Interbasin transfers have traditionally been assigned a large portion of the flux in the GBCAAS area, and although several lines of evidence have been presented to suggest flux between basins through carbonate bedrock, groundwater budget imbalances are a primary evidence. Examples of such imbalances and resulting interbasin fluxes are examined here to provide context for discussions of some specific HAs that follow.

Prudic et al. (1995) conducted a quantitative assessment of fluxes in a large portion of the Great Basin using finite difference methods. They created a two-layer model where the deeper layer was intended to simulate deep flows in carbonate rocks. A summary of their conclusions regarding deep, and therefore largely interbasinal flow, are summarized in Table 1. Other estimates are included for comparison. The interested reader may refer to the primary sources for details concerning the geographic definition of each system as well as details surrounding the development of budget estimates.

According to Prudic et al. (1995), deep flow as a fraction of the total flux varies by a factor of ~20 in the Great Basin. In some regions, nearly 50 % of the total interbasin flux is discharged from regional springs. In areas such as Death Valley that estimate is much lower (~12 %), but absolute fluxes are still large ( $27 \times 10^6$  m<sup>3</sup>/yr) for such an arid region. Not only are large quantities of water involved, most of this flux discharges at Ash Meadows, Nevada, well known as a critical habitat for several endemic endangered plant and animal species.

The water budgets of Welch et al. (2007; Table 1) require special comment. The percentages reflect net

interbasin transfers relative to total recharge estimates for a series of adjacent topographic basins near the Snake Valley area. One such basin, Jakes Valley (Fig. 1), is envisaged to receive interbasin fluxes from basins up gradient and also transport water to basins down gradient. The negative percentage suggests that Jakes Valley is a net exporter of water via interbasin flow and the remainder must be balanced by local recharge. However, similar to the results of Prudic et al. (1995), up to 50 % of the net flux through these desert valleys is attributed to interbasin transfers (Table 1).

On a larger scale, an overview of groundwater budgets in the 57 endorheic GBCAAS HAs (Fig. 2; Heilweil and Brooks 2011) has been provided where published interbasin inflow and outflow estimates have been tabulated. The interbasin inflows and outflows as a percentage of average annual groundwater recharge are shown in Fig. 4. Heilweil and Brooks (2011; their auxiliary 3 E and 3 M, respectively) provide recharge estimates as well inflow and outflow data.

About 50 % of endorheic basins have water budgets where interbasin inflow exceeds 20 % of the basin groundwater recharge and about 67 % had water budget interbasin outflow greater than 20 % of the basin groundwater recharge. Surprisingly, 25 % of the basins had published mean interbasin water budget inflows and outflows that were greater than the total basin groundwater recharge.

Such large calculated water budget imbalances for so many individual basins can only be attributed to the cumulative effect of up-gradient water budget imbalances where it was necessary to transfer calculated up-gradient excess groundwater recharge to down-gradient basins. The extent of this pass-through groundwater was evaluated for the 29 closed basins where both interbasin inflow and outflow were calculated (Fig. 5). Total water budget pass-through imbalances are as great as  $4 \times 10^7$  m<sup>3</sup>/yr.

**Table 1** Summary of proposed deep flows in selected regional flow systems within the Great Basin, where regional springs represent the principal source of the discharge of deep waters. See Heilweil and Brooks (2011; Plate 1) for locations

| Flow system                      | Deep (interbasin) flow (%) <sup>d</sup> | Total interbasin flow fluxes (10 <sup>6</sup> m <sup>3</sup> /yr) |
|----------------------------------|---|---|
| Colorado River <sup>a</sup>      | 47                                      | 120   |
| Death Valley <sup>a</sup>        | 12                                      | 27  |
| Bonneville <sup>a</sup>          | 7                                       | 79  |
| Humboldt River <sup>a</sup>      | 2                                       | 5   |
| Railroad Valley <sup>a</sup>     | 18                                      | 30  |
| Death Valley <sup>b</sup>        | 13                                      | 18  |
| Butte Valley <sup>c</sup>        | -24                                     | 30  |
| Cave Valley <sup>c</sup>         | -9                                      | 11  |
| Jakes Valley <sup>c</sup>        | -14                                     | 17  |
| Lake Valley <sup>c</sup>         | -9                                      | 11  |
| Little Smoky Valley <sup>c</sup> | 0                                       | 0   |
| Long Valley <sup>c</sup>         | -24                                     | 30  |
| Newark Valley <sup>c</sup>       | 5                                       | 6   |
| Snake Valley <sup>c</sup>        | 49                                      | 25  |
| Spring Valley <sup>c</sup>       | 33                                      | 41  |
| Steptoe Valley <sup>c</sup>      | -29                                     | 36  |
| Tippett Valley <sup>c</sup>      | -10                                     | 12  |
| White River Valley <sup>c</sup>  | 33                                      | 41  |
| Butte Valley <sup>c</sup>        | -24                                     | 30  |

<sup>a</sup> Prudic et al. (1995)

<sup>b</sup> Belcher et al. (2002)

<sup>c</sup> Welch et al. (2007)

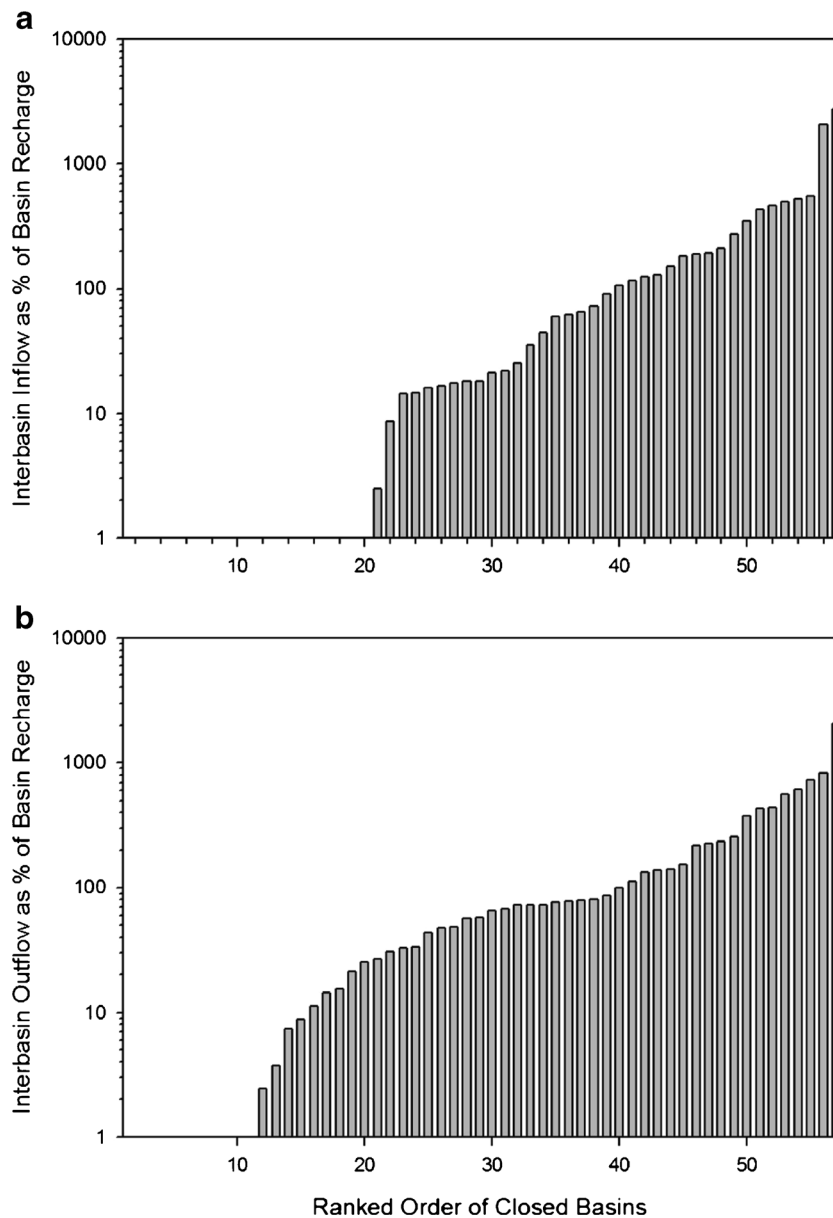
<sup>d</sup> Positive values indicate inflow into the basin, negative values indicate flow out of the basin

## Discussion

### Example comparison of a water budget for the Snake Range region

In addition to the preceding discussion, interbasin flow evidences from water budgets are presented for the entire Great Basin. Here, water budgets for an area of east-central Nevada and west-central Utah (Fig. 6) were examined in greater detail. This area has been intensively investigated over the last decade owing to a proposal to export large volumes of groundwater to Las Vegas, NV to support development (e.g., Welch et al. 2007; Gardner et al. 2011; Heilweil and Brooks 2011; Gillespie et al. 2012).

Gillespie et al. (2012) and Heilweil and Brooks (2011) provide excellent reviews of prior recharge estimates to Snake Valley via interbasin flow (Fig. 6). Rush and Kazmi (1965) suggested  $4.9 \times 10^6$  m<sup>3</sup>/yr of inflow from southern Spring Valley to Snake Valley and  $24.1 \times 10^6$  m<sup>3</sup>/yr from Pine and Wah Wah valleys (Hood and Rush 1965). Nichols (2000) estimated  $17.3 \times 10^6$  m<sup>3</sup>/year of interbasin



**Fig. 4** Reported interbasin groundwater flow in 57 closed basins in the Great Basin, USA (see text for sources): **a** inflow and **b** outflow. Data are plotted as a percentage of total basin recharge

flow enters Snake Valley from Spring Valley north and south of the Snake Range. Welch et al. (2007) provided total inflow estimates to Snake Valley of  $60.4 \times 10^6 \text{ m}^3/\text{year}$ , accounting for 30–55 % of the total recharge to Snake Valley.  $40.7 \times 10^6 \text{ m}^3/\text{year}$  were allocated to a flow path at the southern end of the Snake Range, and  $19.7 \times 10^6 \text{ m}^3/\text{year}$  to the north with most of the intervening range interior considered to be a barrier to flow (Fig. 6). These estimates of interbasin transfers from Spring to Snake Valley vary by a factor of 12.

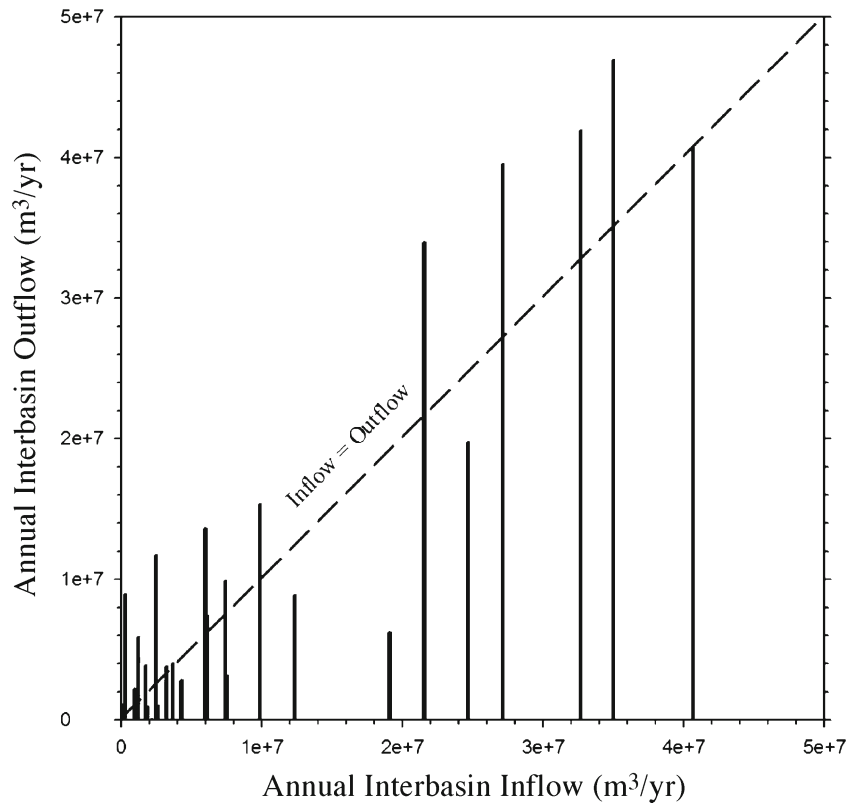
Recently, Sweetkind et al. (2011b) concluded that the potential exists for interbasin flow from Spring to Snake Valley at the southern end of the Snake Range (Fig. 6), but noted that water-budget data presented by Masbruch et al. (2011) do not require it. Masbruch et al. (2011) stated that their budget estimates have uncertainties of about 50 %, which is similar to the upper range of interbasin fluxes described in the preceding section.

which is similar to the upper range of interbasin fluxes described in the preceding section.

#### Summary

There is little doubt that water budget estimates are a valuable tool in the investigation of aquifer systems and their sustainability as water resources. However, in a large and structurally complex region, where recharge and discharge are difficult to estimate with great accuracy and precision, they are often going to be too coarse an instrument to stand alone as compelling evidence for or against interbasin transfers.

Recognizing the imprecision of water-budget estimates, Gillespie et al. (2012) turned to solute, stable isotope, and radioisotope evidence to assess the interbasin flow paths proposed by Welch et al. (2007; Fig. 6). They concluded



**Fig. 5** Comparison of published annual interbasin inflow and outflow values for 29 closed basins where both inflow and outflow were required for water-budget balances. Each line represents the calculated inflow and outflow values for individual basins

that the large posited interbasin flux at the southern end of the Snake Range was unlikely to be present based largely on  $^{14}\text{C}$  and other evidences. Southern Spring Valley waters have apparent residence times on the order of 1 ka, whereas down-gradient waters in Snake Valley are much older, exhibiting residence times that are pre-Holocene in some instances (Fig. 6). A large interbasin flux from Spring Valley at this location should have displaced these pre-Holocene waters.

Systematic differences in well depths could explain the age gradients in Fig. 6, where shallow wells intersect Holocene water and deeper wells tap Pleistocene aquifers. Gillespie et al. (2012; data source for Fig. 6), however, noted that except for a subset of samples for Snake Valley obtained by the Utah Geological Survey, well depths are poorly documented, with few if any wells completed in carbonate rock. Yet, prior studies (Welch et al. 2007; Wilson 2007) maintained that basin fill and alluvial aquifers are in communication such that alluvial aquifer elevations reflect that of underlying carbonate bedrock. If this conceptualization of the system is correct, then the gradient in ages should not exist because upward fluxes into unconsolidated material in Snake Valley should not be so much older than Spring Valley given the large posited interbasin fluxes.

### Potentiometric maps

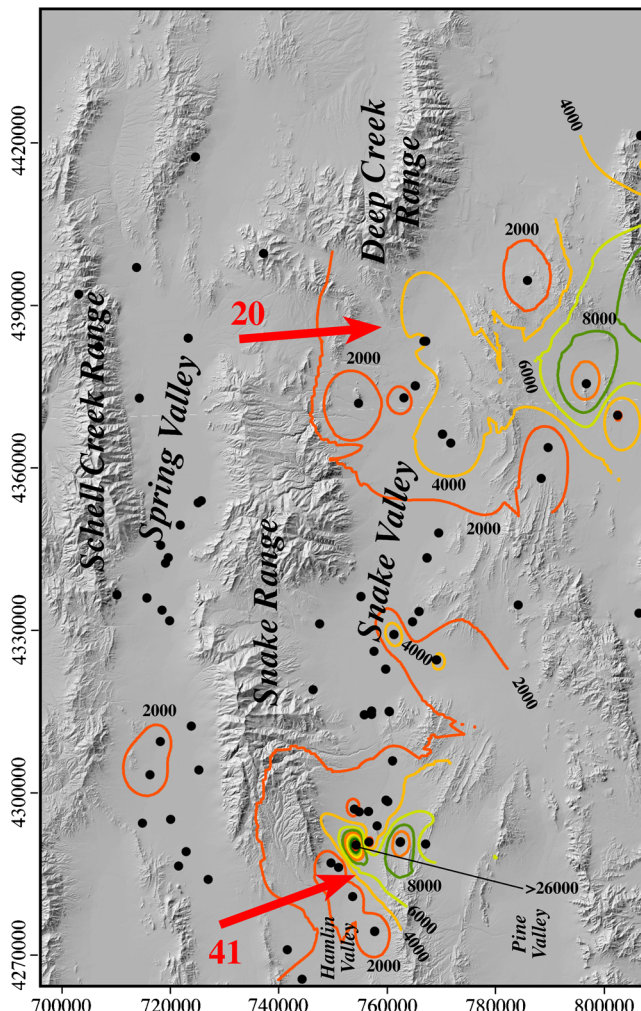
No two studies can be expected to produce identical potentiometric surfaces based on the data that were used

(or available), as well as how the system was conceptualized. Potentiometric contour maps are, however, intended to be models of the real world, and like all good models, should simplify the system while retaining its essential characteristics and permitting predictions. The potentiometric contour maps discussed in the following were constructed at different scales for different purposes, so due caution is warranted in their use. The primary objective of this study is not to point out flaws but rather to compare them, which is very instructive in the context of using fluid potentials to evaluate interbasin flow where it is assumed that flow is generally perpendicular to contours (e.g., Sweetkind et al. 2011b).

### Snake Range area

Recent studies have produced four sets of potentiometric contours for the Spring and Snake Valley area (Fig. 7). Welch et al. (2007) produced separate contours for: (1) valley fill, and (2) regional bedrock aquifers (Fig. 7a). Heilweil and Brooks (2011) produced a map of much of the Great Basin with a single set of contours (Fig. 7b), whereas a map by Gardner et al. (2011) focused on the area of Spring and Snake valleys, also employing a single contour set (Fig. 7c). Gardner et al. (2011) who had access to drilling data that were unavailable to Welch et al. (2007), concluded that there is only a single aquifer system on the basis of similar water levels in nested wells completed in valley fill and bedrock.





**Fig. 6** Digital topography of the Snake Valley area. Colored lines represent contours of Pearson model ages (Pearson and Hanshaw 1970) for Snake Valley region waters where control points (springs and wells) are indicated by black dots. Red arrows indicate the location and flux ( $10^6 \text{ m}^3/\text{yr}$ ) of proposed interbasin flow paths by Welch et al. (2007). Contour labels are in years and the contour interval is 2,000 years. Data sources are from Gillespie et al. (2012). UTM coordinates, zones 11 and 12 (NAD83) are in meters

Differences between inferred elevations between regional- and basin-fill aquifer potentials of Welch et al. (2007) in eastern and northeastern Spring Valley are on the order of 100–125 m, are approximately zero in southern Snake Valley, and from 30 to 60 m different in northern Snake Valley (Fig. 7a), with the regional aquifer having higher potentials. The construction of separate potential maps infers that although there may be upward leakage from the regional system into valley-fill aquifers, they are largely separate systems or were conceived as such.

Although Welch et al. (2007) invoked interbasin transfers to Snake Valley at discrete locations due to suspected flow barriers underneath much of the Snake Range (Fig. 6), if contours were considered in isolation, interbasin flow would be envisaged from a recharge area in the Schell Creek Range (groundwater mound, Fig. 7a) and to flow through the Snake, Confusion, and Fish

**Fig. 7** Index map for the Snake Valley area and potentiometric contours from a Welch et al. (2007) where valley-fill aquifer elevations are in blue and regional carbonate aquifer in black. b Heilweil and Brooks (2011) potentiometric contours (black) with control points (star = well, triangle = spring, circle = stream). Red contours in a and b represent mean annual precipitation (1981–2010) in mm/yr (PRISM 2012). c Potentiometric contours (blue) from Gardner et al. (2011). For a, b and c, hydrostratigraphy is based upon geology by Ludington et al. (2005). Contour elevations, originally published in feet, are reported in meters. UTM coordinates, zones 11 and 12 (NAD83) in meters

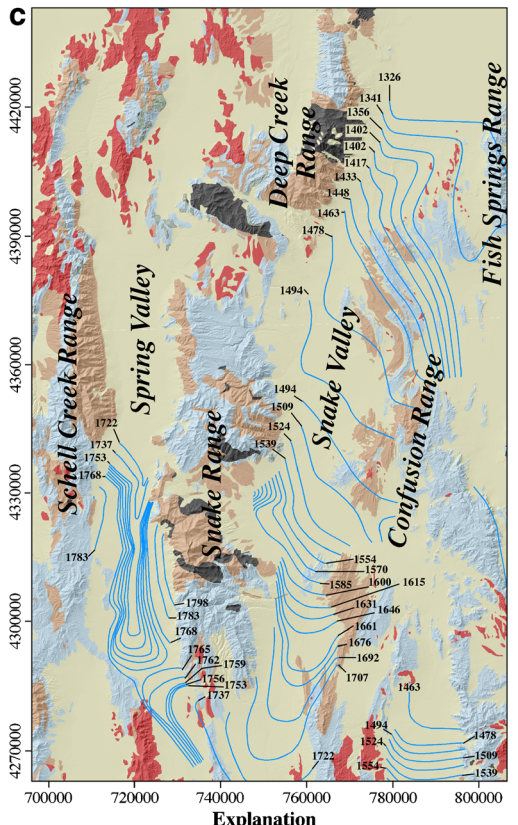
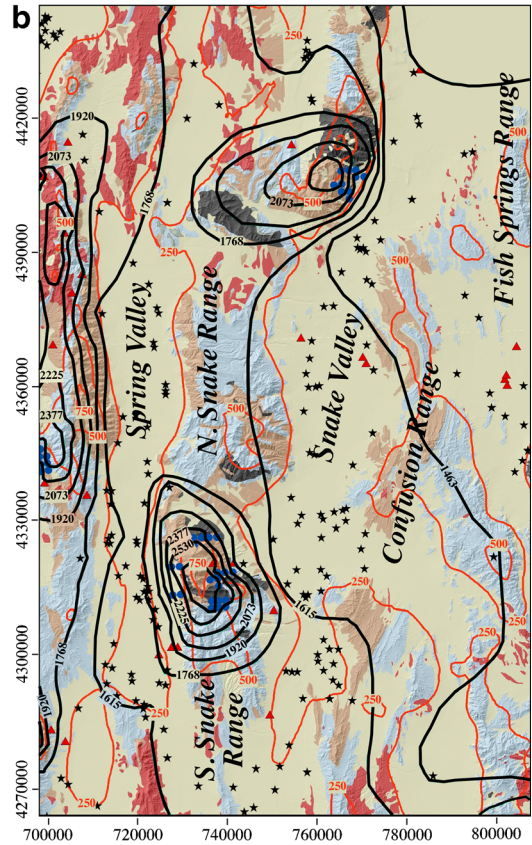
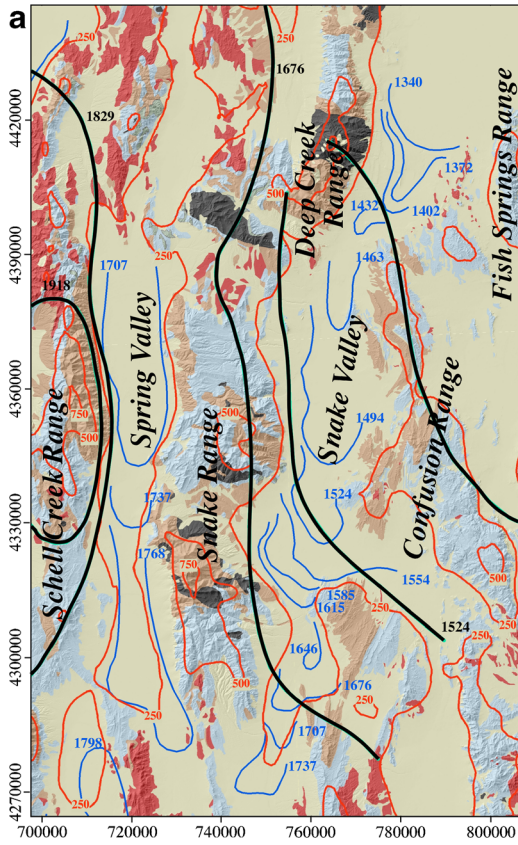
Springs ranges. The authors have found it odd that a groundwater mound is present beneath the Schell Creek Range, but not present in the somewhat high and broad northern Snake Range, which should be an effective orographic moisture trap and locus of groundwater recharge. Isohyet maps indicate that this should be the case (Prudic et al. 1995, p. D8; PRISM 2012). As noted in Fig. 7a, there are broad regions of elevated precipitation in the Snake Range.

In contrast, Heilweil and Brooks (2011; Fig. 7b) show recharge mounds beneath the Schell Creek, southern Snake, and Deep Creek ranges. However, there is no mound beneath the northern Snake Range. This also seems odd, and inspection of their map data (Fig. 7b) indicates that this is an artifact due to the choice of control points.

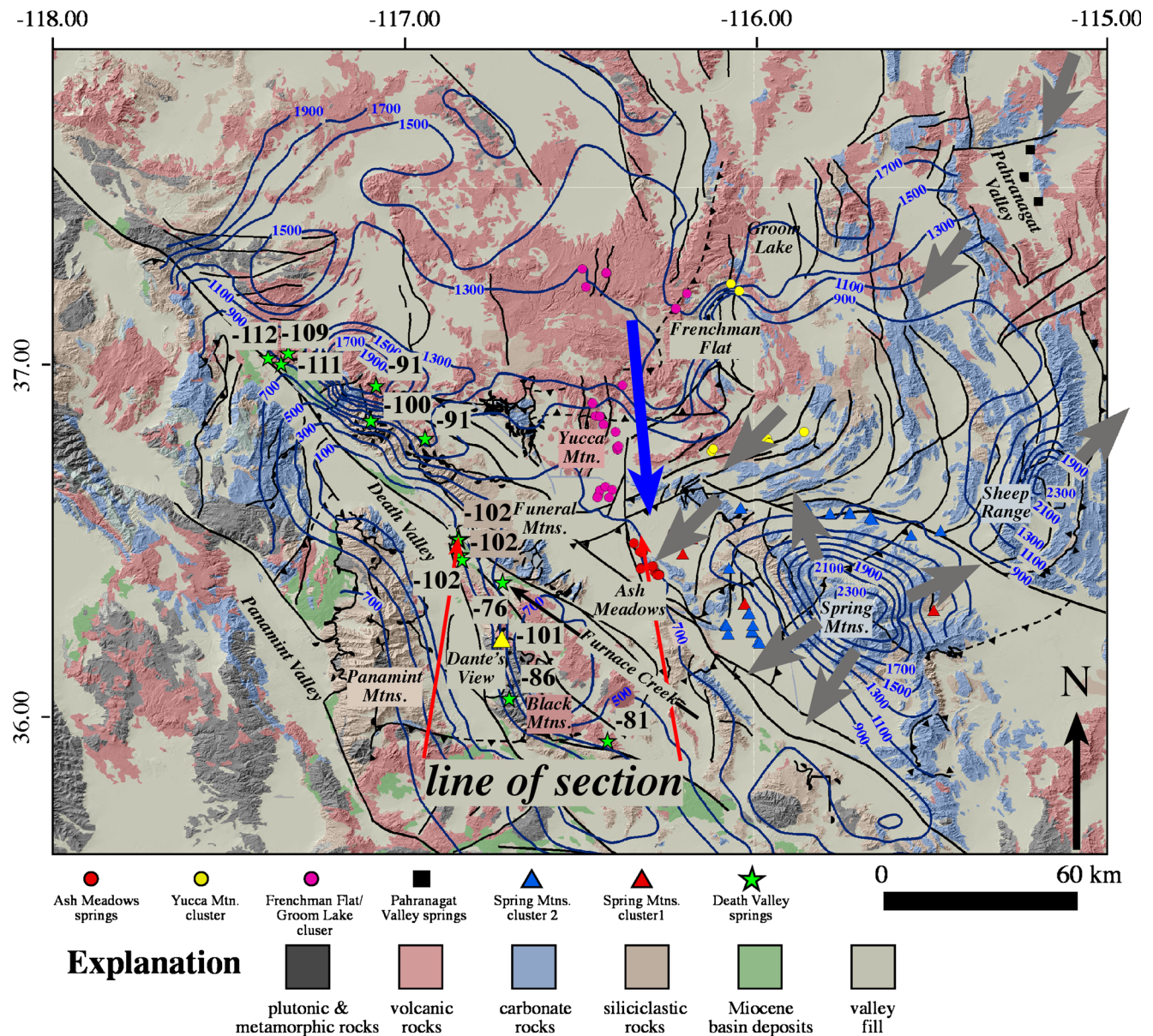
Most of the control points for high-elevation contours centered over the southern Snake, Schell, and Deep Creek ranges are streams and springs, some of which have elevations  $>2,500 \text{ m}$ . In the adjacent valleys, subsurface water levels in wells comprise the control on the potentiometric surface. In other words, if a single connected aquifer system exists between high elevations in the southern Snake Range and beneath Snake and Spring valleys, there are  $>1,000 \text{ m}$  of head over relatively short horizontal distances (Fig. 7b).

There is little doubt that such high-elevation spring and stream waters are related to recharge that replenishes valley fill and bedrock aquifers in this area via losing streams and mountain-front recharge (Gillespie et al. 2012 and references therein). However, high-elevation springs and streams likely reflect local perched systems in the active zone (Mayo et al. 2003) that may not connect to deep bedrock and valley-fill aquifer systems across range interiors.

Davisson et al. (1999) present a clear discussion of the issue of the separation of local, mountain and regional flow systems in terms of stable isotope evidence. In the Spring Mountains (Fig. 8) of southern Nevada,  $\delta^{18}\text{O}$  values of springs and streams exhibit expected depletions at increasing elevation, consistent with local flow systems that discharge close to their recharge elevations. By contrast, the nearby Sheep Range (Fig. 8) has high-elevation waters that are enriched relative to that from wells in adjacent valleys. In this case, depleted valley waters were suggested to have moved southward from more northerly recharge areas, with mountain springs representing local systems. In both cases, oxygen and







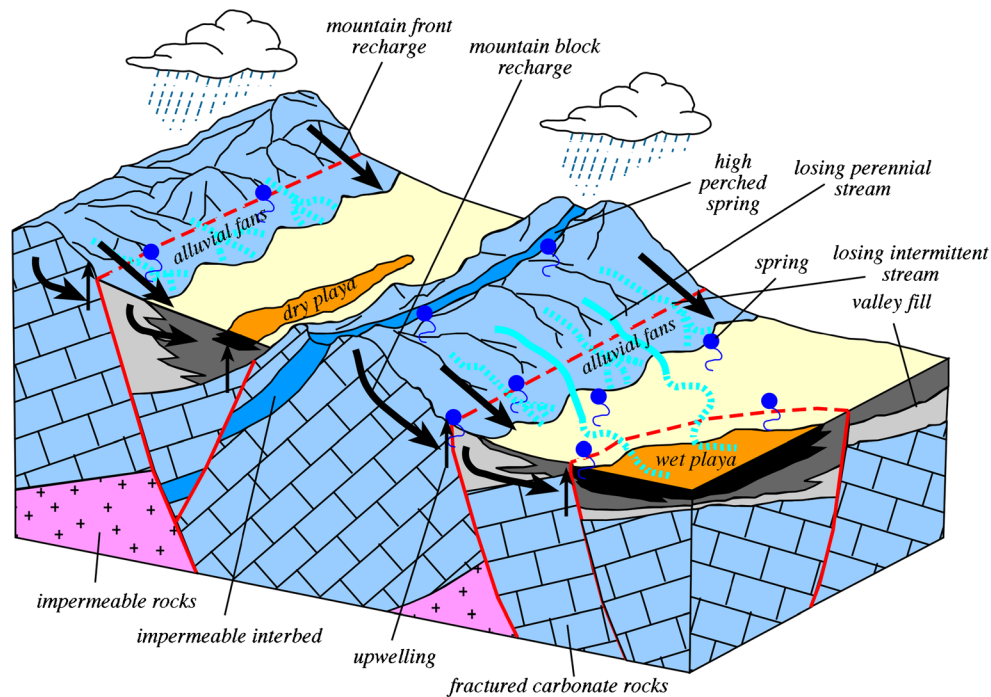
**Fig. 8** Simplified hydrogeologic and fault map with potentiometric contours (blue lines; adapted from Ludington et al. 2005; Potter et al. 2002) for the Death Valley and Ash Meadows regions. Bold grey arrows show the posited flow paths of Thomas et al. (1996) for Ash Meadows waters. Bold blue arrow shows the flow path proposed by Bushman et al. (2010). Geometric symbols refer to spring and well locations and groups discussed in the text. Numbers adjacent to green stars are  $\delta D$  values (%) for selected springs and wells in the Death Valley area. Faults are indicated as black lines

hydrogen isotope data suggest high-elevation waters are perched systems unrelated to regional aquifers and their inclusion in generating potentiometric contours may lead to groundwater mounds that do not exist in regional flow systems.

There are other potential problems with Fig. 7b. If one accepts a groundwater mound beneath the southern Snake Range, there should be one beneath the northern Snake Range as well. Overlapping groundwater mounds could create a groundwater divide preventing interbasin flow between Spring and Snake valleys, although separate mounds should form beneath the northern and southern parts of the range due to the topographic divide that separates them.

Gardner et al. (2011) produced a potentiometric map with fine-scale contours based upon the results of a drilling program dedicated to a better definition of the flow system prior to proposed groundwater withdrawals. They also did not separate valley fill and regional carbonate aquifer systems, nor did they attempt to contour groundwater elevations within the Snake or other ranges (Fig. 7c). Their contours imply that water flows away from the range fronts toward valley centers and then along valley axes. The only location where interbasin flow seems possible is between Spring and Snake Valleys south of the southern Snake Range where the large flux was proposed by Welch et al. (2007; Fig. 6). However, as discussed in the previous, residence times (Gillespie et al.





**Fig. 9** Cartoon illustrating conceptualized components of local recharge through mountain blocks, mountain fronts, and losses from perennial and intermittent streams. See text for discussion

2012) seem to preclude this flow path, or at least large fluxes along this path, and Masbruch et al. (2011) shied away from invoking large interbasin transfers at this locality.

Gillespie et al. (2012) present a detailed view of the Spring and Snake Valley systems and concluded that recharge to Spring and Snake valleys is sustained by mountain front and mountain block recharge as well as stream losses in and near adjacent high mountains (Fig. 9). High-elevation perched systems are rejected recharge, representing only a fraction of the infiltration at high elevation. Thus, Snake Valley aquifers are replenished without a significant component of interbasin flow from Spring Valley. Given that Spring Valley is bounded by both the Snake and Schell Creek ranges, both effective moisture traps, residence times of groundwater in this valley are shorter than those in Snake Valley where some waters are clearly pre-Holocene and the valley is bounded by a single large range. As a result, Snake Valley is more vulnerable to excessive water withdrawal than Spring Valley. The interested reader is referred to Gillespie et al. (2012) for more detail.

#### Death Valley area

It is noted that published potentiometric contour maps near Furnace Creek are problematic, and similar to the Snake Valley area (Fig. 7), exhibit very different patterns between published studies (Fig. 10). For example, there is little or no well control for contours through the interior of the ranges that bound the east side of Death Valley. It is possible to draw contours through the Funeral Range based on elevation differences between water levels in

wells and springs on both sides of the range, but this is not direct evidence of a hydraulic connection. The authors disagree that these contours provide direct evidence of underflow as contended by Belcher et al. (2009).

Similar to the Snake Range region, one set of contours (blue) in Fig. 10 shows groundwater mounds beneath portions of the mountains on the east side of Death Valley, whereas the other set (red) does not. In this regard, one or both of these sets of contours misrepresents the natural system.

The mismatch between the two sets of contours (Fig. 10) has been quantified in order to ascertain how much apparent variation in potential can exist as a function of generating contour maps by different researchers working in the same area. Reasonably consistent maps should have contours that seldom cross, and when they do, the differences in inferred potentials should be small, especially on average. In Fig. 10a, contours cross more than 100 times, and the average difference (red contour elevations minus blue) is  $\sim -325 \pm 320$  (1 SD) m, with a maximum absolute difference of  $>1,000$  m. The negative sign of the mean is largely a function of red contours crossing the blue contours of groundwater mounds.

In examining contour elevations in Panamint Valley, through the Panamint Mountains, and into Death Valley (Fig. 10), underflow beneath the core of the range is inferred but highly unlikely given the nature of bedrock, and the authors are not aware of any studies that propose it. This raises the question of the value of contouring water potentials through range interiors that lack adequate data. The high and extensive Panamint Range should be a locus of recharge as isohyets indicate that high elevations

should receive between 20 and 40 cm/yr precipitation (Prudic et al. 1995, p. D8). In fact, the Panamint Range is nearly as high as the Spring Mountains, where isohyets indicate precipitation rates at >50 cm/yr (Prudic et al. 1995, p. D8), so it is possible that precipitation (and recharge) in the Panamint Range is underestimated.

### Summary

Regarding basin-to-basin transfers, potentiometric maps as guides for interbasin flow have a number of potential pitfalls, depending on how the flow system is conceptualized, which in turn influences those data to be included to generate contours. Potential is a necessary, but insufficient condition to demonstrate interbasin transfers. A bottle of water on a table will have greater potential than another on the floor, although there is no flow between them. It is important not to confound potential with kinetic energy.

Regarding groundwater mounds, the authors wish to be clear that in this study the analysis of them beneath mountain ranges is centered on the criteria used (or omitted) when constructing contours. As discussed in the following, it is questioned whether the very construction of groundwater mounds is warranted in such situations due to the nature of candidate control points that are available in geologically complex interiors.

In the authors' view, groundwater mounds beneath large mountain ranges may simply be artifacts of the types of data used in contouring rather than representing real integrated flow systems. If such mounds represent local or ephemeral features, which they probably often do, such high elevation streams and springs may lead apparent groundwater divides that are identified as barriers to interbasin flow where none actually exist. These high elevation perched systems may lead to recharge of deep, regional aquifers via losing streams, mountain block and mountain-front recharge (e.g., Wilson and Guan 2004; Fig. 10). However, this does not mean they are connected.

A simple example of the perils of linking shallow perched flow systems and deep aquifers can be seen in central Oahu, Hawaii, USA. Nelson et al. (2013; and references therein) noted that in central Oahu, the top of the Ghyben-Herzberg lens as documented from numerous deep wells is generally <30 m above sea level (asl), whereas the ground surface is >250 m asl. The numerous perennial streams that flow across this region are clearly supported by shallow perched systems and it would be a gross error to use stream elevations in this region to define the top of the regional aquifer.

Thus, improper use of control points may lead to fictitious groundwater mounds, which would appear to impede interbasin flow. It is noted that this conclusion is ironic given the authors' position that interbasin flow is generally uncommon in this region; however, it is clear that caution should be applied to both the construction and interpretation of potentiometric contours as they relate to interbasin fluxes.

**Fig. 10** Simplified hydrogeologic maps (adapted from Ludington et al. 2005) for the immediate Death Valley and Ash Meadows regions. **a** Illustrates a regional scale, whereas **b** focuses on the Ash Meadows–Furnace Creek region. *Blue arrow* illustrates a generalized flow path from Ash Meadows to Furnace Creek, as well as an approximate line of section in Fig. 13. Faults (*black lines, bold* for major structures) and *red* potentiometric contours are adapted from Belcher et al. (2009) with *blue* contours from Potter et al. (2002). *Numbers* adjacent to *green stars* are  $\delta D$  values (‰) for selected springs and wells in the Death Valley area. Only faults in the immediate vicinity of Furnace Creek are shown. Symbols as for Fig. 8

### Isotopic evidence

#### Death Valley region

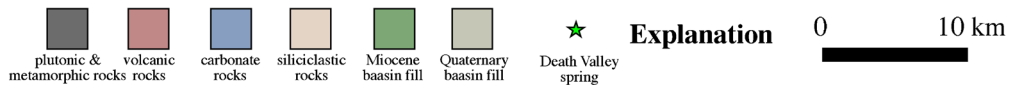
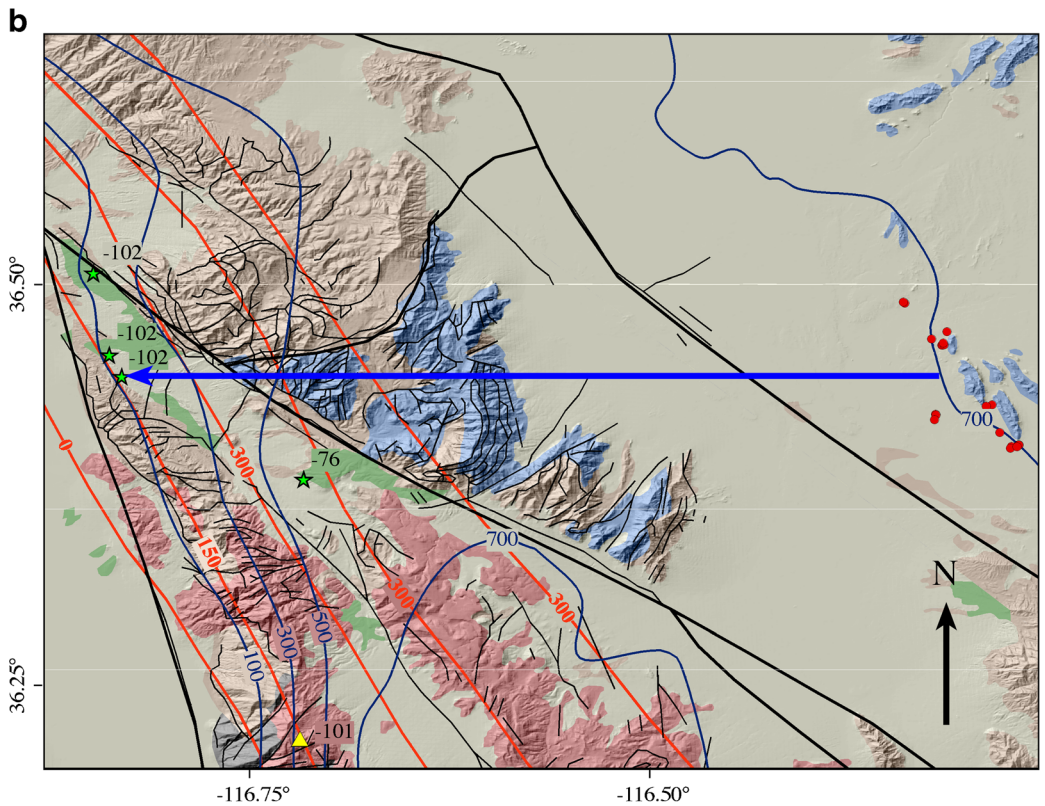
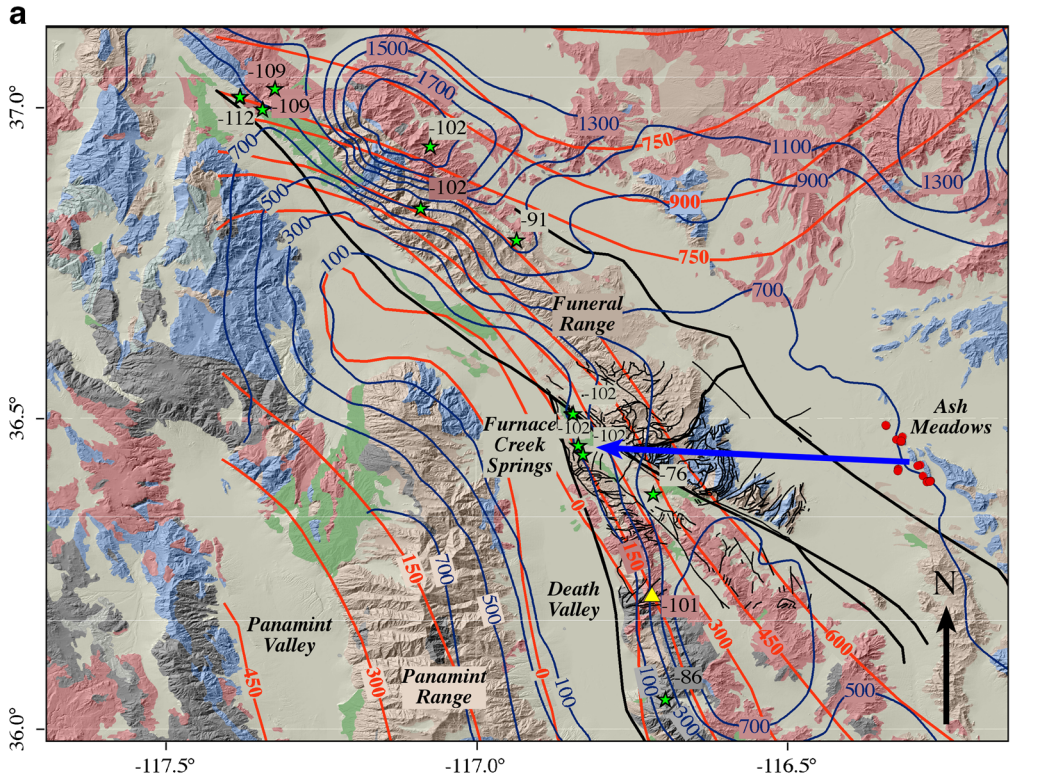
Early and many subsequent studies invoking large-scale bedrock interbasin flow within the Great Basin were often based on isotopic and other geochemical evidences (e.g., Winograd and Friedman 1972; Winograd and Pearson 1976; Thomas et al. 1996; Davisson et al. 1999; Rose and Davisson 2003), with some studies suggesting that water transects numerous basins and ranges with transport distances within carbonate rocks up to hundreds of kilometers from recharge to discharge locations (Fig. 1). As the interbasin flow from an isotopic perspective is evaluated, this study focuses on flow paths leading to discharge at Ash Meadows, Nevada, and Furnace Creek California (USA; Figs. 8 and 10).

Employing hydrogen isotope data ( $\delta D$  values), Winograd and Friedman (1972) estimated that ~35 % of the ~38,000 L/min discharging at Ash Meadows (Winograd and Pearson 1976; Bushman et al. 2010) arrives by interbasin flow, with the remainder arriving from the Spring Mountains, also by interbasin flow albeit, along a much shorter flow path (Fig. 8). More recently, Thomas et al. (1996) reached similar mixing proportions, employing carbon isotope data and inverse solute models as additional constraints. Other studies suggest water sources other than these may contribute to Ash Meadows (Davisson et al. 1999; Rose and Davisson 2003).

As noted by Winograd and Friedman (1972), the Ash Meadows and Furnace Creek areas are hot and very arid. Although the Funeral Range adjacent to the Furnace Creek springs is plausible as a recharge area (Fig. 10), Ash Meadows is bounded to the east by a low set of hills that could not possibly act as a recharge area for the observed discharges. Thus, isotopic evidence is consistent with interbasin flow because of: (1) great aridity and high regional potential evapotranspiration relative to rainfall, (2) very small local precipitation catchments, and (3) depleted  $\delta^{18}O$  and  $\delta D$  values indicative of recharge at higher elevation or more northerly (i.e., cooler) regions.

Ash Meadows and Furnace Creek waters have  $\delta^{18}O$  values that range from –13 to –14 ‰ (Anderson et al. 2006; Bushman et al. 2010 and references therein). Using the model of Bowen and Wilkinson (2002), precipitation weighted  $\delta^{18}O$  values of rainfall on summit regions of the hills adjacent to Ash Meadows should be on the order of –6 to –7 ‰, and –8 to –9 ‰ in the Funeral Mountains. This seems to require that both Ash Meadows and







Furnace Creek waters were recharged at higher elevation, under a cooler climate, as a result of cold-weather precipitation, or elements of all three.

The Bowen and Wilkinson (2002) model is an excellent starting point for setting expectations regarding the isotopic composition of recharge water. However, it may break down in cases where volume-weighted means are not appropriate. Throughout much of the arid and mountainous Great Basin of the USA, most warm-weather precipitation may be returned to the atmosphere through evapotranspiration rather than producing significant recharge (e.g., Winograd et al. 1998). Winter rains, snowfall at high elevation, and spring snowmelt likely produce most groundwater recharge. For example, the stable isotopic values of groundwaters in the Snake and Spring Valley region ( $\delta D$  generally  $< -100$  ‰ and  $\delta^{18}O$  generally  $< -13$  ‰; Gillespie et al. 2012) are only consistent with cold weather precipitation (IAEA 2001).

Water discharging at Furnace Creek has a  $\delta D$  value of  $\sim -102$  ‰ (Anderson et al. 2006). Friedman et al. (1992) report that modern winter (mid-Oct. through mid-April) precipitation at nearby Dante's View (1,575 m elevation) on the crest of the Black Mountains had a nearly identical mean value of  $-101$  ‰ (Fig. 10a) over a 7-year period. Anderson et al. (2006) report, however, that low-flux mountain springs in the Death Valley region, which may act as proxies for modern recharge by integrating winter precipitation, tend to be more enriched (Fig. 10a). Thus, the evidence is mixed as to whether Furnace Creek waters could have been recharged locally under the current climate.

Recharge under a cooler climate (i.e., last glacial maximum or Younger Dryas), however, could permit recharge to have occurred near Furnace Creek. Winograd et al. (1997), by contrast, assert that Ash Meadows waters were recharged in Holocene time. If this is true, then it is necessary to invoke higher-elevation or more northerly recharge localities that are remote to Ash Meadows and Furnace Creek. An inherent difficulty in establishing flow paths by examining remote candidate recharge locations with suitable isotopic compositions is that the candidate areas are not necessarily unique. If one looks far enough or in multiple directions, it will often be possible to find multiple candidate waters of suitable  $\delta^{18}O$  and  $\delta D$  values.

Ash Meadows, for example, may have its isotopic composition matched by more than one single candidate recharge area, or its isotopic composition could be matched by subsurface mixing of multiple sources. Thus, other lines of evidence should be brought to bear to discriminate between candidate sources. In the case of Ash Meadows, possible sources and flow paths have been assumed to lie within carbonate rocks and geochemical signatures imprinted by them, which is the essence of the arguments made by Winograd and Pearson (1976) and Thomas et al. (1996).

As discussed in the preceding, Winograd and Friedman (1972) noted that  $\delta D$  values at Ash Meadows are matched by a mixture of 65 % water derived from the Spring Mountains and 35 % from Pahranaagat Valley (Fig. 8).

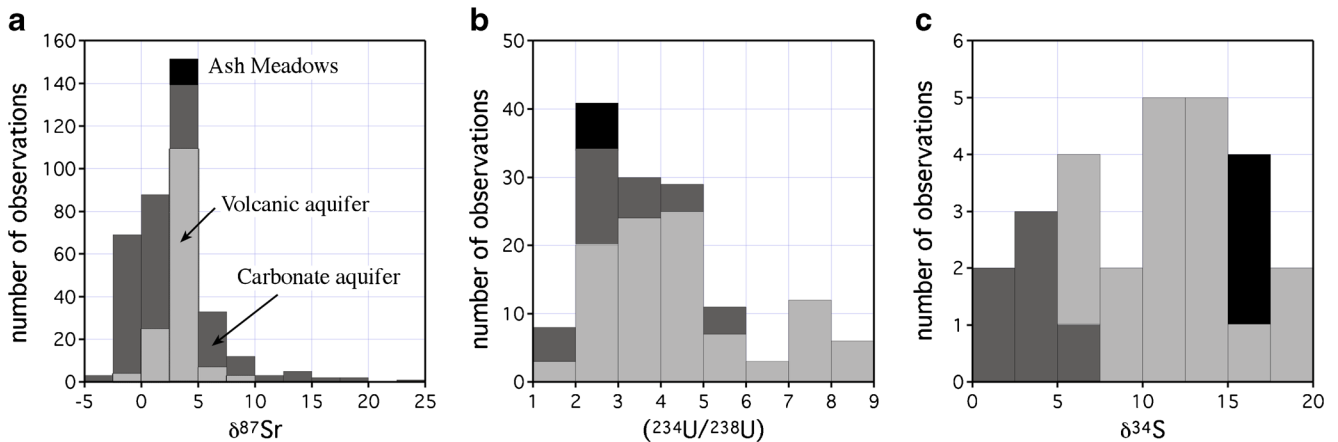
Components of more depleted water from Pahranaagat Valley are required as waters from Spring Mountains are too enriched to match Ash Meadows. In contrast, a parsimonious interpretation would involve a single, relatively nearby source without mixing.

Bushman et al. (2010) compared  $\delta^{18}O$  and  $\delta D$  values for waters of various hydrochemical facies north of Ash Meadows. They first conducted cluster analysis based on solute chemistry of waters in order to establish endmember facies for inverse mass balance modeling. The  $\delta^{18}O$  and  $\delta D$  values for these clusters were then compared to Ash Meadows. Two clusters, one from the Yucca Mountain area and one from the Frenchman Flat/Groom Lake region north of Ash Meadows (Fig. 8) are excellent isotopic matches. The probabilities that they are an isotopic match to Ash meadows are 0.43 and 0.85, respectively.

Inverse mass-balance models are useful tools in constraining flow paths where a reasonable set of mineral dissolution or precipitation reactions must be found. However, such models do not establish flow paths, but they can infer that such paths are plausible. Alternatively, they can rule out certain paths if there is no reasonable evolutionary path for the solutes. Like Thomas et al. (1996), Bushman et al. (2010) were able to show that a reasonable set of mass balance reactions exist for both the Yucca Mountain and Frenchman Flat/Groom Lake clusters, in addition to their isotopic concordance.

Thus, the mass balance models of Bushman et al. (2010) explain the chemical evolution of waters from volcanic aquifers north of Ash Meadows to the carbonate aquifer of Ash Meadows proper, rather than being confined within a carbonate sequence. Their conceptualization of the system was also broad enough to test waters outside of carbonate lithologies. Sr-, U-, and especially S-isotopic data are consistent with flows originating from volcanic aquifers to the north (Bushman et al. 2010). In fact, available  $\delta^{34}S$  values seem to rule out flow being confined entirely within a carbonate aquifer (Fig. 11). Thus, two important conclusions can be drawn. First, conceptualizing and then modeling flow paths to only lie within carbonate rocks is circular. Second, the re-interpretation of the origin of Ash Meadows waters by Bushman et al. (2010) is more straightforward as no mixing is invoked and extremely remote mixing endmembers are not required. This argument is strengthened when structural considerations (discussed in the following) are brought to bear.

Winograd and Friedman (1972) and subsequent authors noted correctly that one must be able to assume that  $\delta^{18}O$  and  $\delta D$  values remain essentially constant between recharge and discharge areas, or between observation points along flow paths. Large changes in the composition of precipitation due to climate variation could mask the genetic relationships among waters in a given system. Winograd et al. (1997) concluded that this was met in the Ash Meadows region as they estimate groundwater residence times of  $\sim 2,000$ – $3,000$  years (i.e., late Holocene), based on unpublished ages on dissolved organic



**Fig. 11** Histograms of **a** strontium (Sr), **b** uranium (U) and **c** sulphur (S) isotopes from available data for volcanic and carbonate aquifers compared to Ash Meadows, where  $\delta^{87}\text{Sr} = \left( \frac{^{87}\text{Sr}/^{86}\text{Sr}_{\text{measured}} - 0.7092}{0.7092} \right)$ ,  $(^{234}\text{U}/^{238}\text{U})$  is the activity ratio, and  $\delta^{34}\text{S}$  is standard notation. The isotopic composition of U and Sr overlaps with both aquifer systems. S isotopes, however, are consistent with Ash Meadows waters being derived from volcanic sources. Modified from Bushman et al. (2010)

carbon as well as  $\delta^{18}\text{O}$  stratigraphy of vein calcite from Devils Hole, which is part of the Ash Meadows system.

Anderson et al. (2006) calculated  $^{14}\text{C}$  ages from dissolved inorganic carbon for springs and wells of Ash Meadows and surrounding regions that are generally several thousand to a few tens of thousands of years in age. This includes two of the major Ash Meadows springs, which yielded ages from 13,000 to 25,000 years as calculated from the Pearson and Fontes models (Pearson and Hanshaw 1970; Fontes and Garnier 1979). Using data in Winograd and Pearson (1976), Pearson model ages have been calculated that range from 7,200 to 26,200 years for 14 samples from 8 Ash Meadows springs. Two young ages (7,200 and 8,000) represent a single spring (Crystal Pool), whereas all 12 remaining apparent ages exceed 15,500 years.

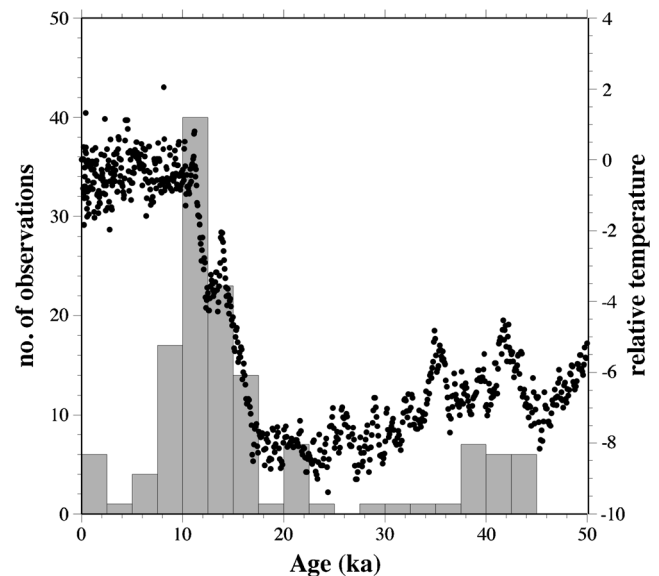
The divergent residence times inferred by Winograd et al. (1997) and Anderson et al. (2006), including the recalculated ages in this study, cannot all be meaningful estimates of the residence times of the same waters. Anderson et al. (2006) provided what the authors of this report view to be a compelling argument that Ash Meadows waters are pre-Holocene in age. In essence, they showed that the water in Devils Hole is in isotopic equilibrium with vein calcite (data from Landwehr et al. 1997) that reflects full or nearly full glacial conditions. In other words, the water in Devils Hole today was recharged during a pre-Holocene cold climate state, consistent with evidence from  $^{14}\text{C}$  ages on dissolved inorganic carbon.

The view of Anderson et al. (2006) is supported by regional studies of paleo-spring deposits in the region. Figure 12 compares a compiled histogram of 137 published  $^{14}\text{C}$  ages of spring deposits against the Vostok ice core temperature record as a reference. A very large fraction of these ages indicates that springs systems in the Death Valley region were active during the last glacial maximum and especially during the transition to Holocene climate. This view is consistent with the notion that large

regional springs in the Death Valley area are fed by groundwaters recharged during cooler climates in pre-Holocene time.

#### Summary

The difference in estimated residence times of water discharging from the Ash Meadows system has obvious implications for the use of  $\delta^{18}\text{O}$  and  $\delta\text{D}$  values to delineate flow paths. If the residence time of all water along a flow path is confined to the Holocene Epoch, then a relatively stable climate will have imposed only minor



**Fig. 12** Histogram (gray bars) of  $^{14}\text{C}$  ages of published spring deposits in the Death Valley region (Quade et al. 1995, 1998, 2003; Paces et al. 1996) with the Vostok temperature record (black dots; Petit et al. 2001) as a baseline comparison for climate state as a function of time. Relative temperature represents deviations compared to modern mean temperature

variations on the isotopic composition of recharge waters, making these stable isotopes useful tracers. However, if the waters at a discharge point reflect the influence of a different climate than modern recharge, it may be difficult to trace water up gradient toward its source, in which case the use of  $\delta^{18}\text{O}$  and  $\delta\text{D}$  values becomes more difficult to employ. The authors' view is that water discharging at Ash Meadows has been transported from the north (Bushman et al. 2010). However, this interpretation can only be correct if both source and Ash Meadows waters were recharged in pre-Holocene time.

### **Structural and stratigraphic controls**

In addition to potential differences, an integrated fracture network is required to permit bedrock interbasin flow, especially in carbonate rocks which have little primary permeability. It is suggested that there are three mechanisms by which such integrated permeability structure could be modified: (1) fracture self sealing by vein calcite or other mineral deposition, or widening of fractures by solution weathering, (2) aquifer segmentation by impermeable fault cores or capture of flow into permeable damage zones, and (3) segmentation by impermeable siliciclastic interbeds intercalated within carbonate rocks. The question then becomes the spatial dimensions over which integrated fracture permeability can exist in fractured carbonate rocks without flow being impeded by one of these mechanisms.

#### *Solution weathering and vein formation*

Casual inspection of outcrops of carbonate rock in the Death Valley region shows that many fractures are weathered and widened, with a tendency for karstification within certain beds. Winograd and Thordarson (1975) noted solution weathering is pronounced near the surface and that many fracture systems in the Death Valley region are transmissive. However, they also noted in recovered well cores that about 95 % of fractures are sealed due to secondary carbonate minerals, quartz, clay and sometimes iron/manganese oxide. Thus, only a small subset of fractures transmits water. Surprisingly, some recent regional syntheses of large-scale carbonate aquifer systems pay little attention to vein calcite or other fracture filling as a control on flow (e.g., Welch et al. 2007; Sweetkind et al. 2011a, b).

In addition to solution weathering, the plugging of fractures with calcite in carbonate rock indicates that aquifer systems in times past have been oversaturated with carbonate minerals and may presently be oversaturated. Oversaturation does not necessarily result in precipitation due to kinetic inhibitions on crystallization, but oversaturated waters are not likely to produce solution weathering.

We have re-examined the data presented by Bushman et al. (2010) and note that their Spring Mountain and Pahranaagat Valley hydrochemical facies, sources of Ash Meadows water invoked by Winograd and Thordarson

(1975) and Thomas et al. (1996), are all oversaturated with respect to calcite, aragonite, and dolomite, as are Ash Meadows waters themselves. If anything, there will be a tendency to seal fractures along such flow paths unless extension can keep pace with calcite precipitation (Riggs et al. 1994).

Bushman et al. (2010) presented a thorough discussion of solute evidence of interbasin flow, noting that the hydrochemical facies near Yucca Mountain (Yucca Mountain and Frenchman Flat/Groom Lake, Fig. 8), their preferred sources for Ash Meadows, are undersaturated with respect to all three carbonate phases. Thus, as flow is transferred from volcanic to fractured carbonate rocks, there may be a tendency to maintain or widen fracture conduits until saturation is reached along the flow path. Additionally, Bushman et al. (2010) noted that north-south fracture flow should be favored in this region due to active east-west extension, which may tend to counteract the plugging of fractures by carbonate minerals. Ultimately, it is impossible to know the state of solution weathering or fracture sealing in deep aquifers. As noted by Winograd and Thordarson (1975), outcrop studies are poor analogs in this regard, coring can never sample enough of the subsurface, and pumping tests at a large enough spatial scale to test interbasin flow are impractical. Nonetheless, examining saturation states can reveal the tendency of systems to plug or widen fractures.

#### *Stratigraphic flow barriers*

Impermeable bedrock commonly creates confined conditions. However, in tilted rocks, which underlie most ranges in the Great Basin, impermeable beds can act as dams to flow, especially in regions where the dip of beds forms a physical barrier. Recent studies have highlighted the role that such impermeable interbeds probably play in interbasin flow (Anderson et al. 2006; Miner et al. 2007; Gillespie et al. 2012).

Sweetkind et al. (2011a) provide a synthesis of the hydrogeologic framework for carbonate aquifers of the Great Basin, USA. In both the Snake Range and Ash Meadows regions, a Cambrian through Devonian lower carbonate aquifer has been proposed, overlain by a Mississippian shale aquitard, overlain by a Pennsylvanian and Permian upper carbonate aquifer. Although there is little doubt that the Mississippian shales are impermeable, there may be a tendency to overlook low permeability strata within both carbonate aquifers. Tables 2 and 3 summarize evidence for a significant number of such possible barriers, many of which may hold the potential to restrict flow. As noted by Hershey et al. (2010), clastic rocks in the eastern Great Basin contain minor carbonate sequences, whereas carbonate rocks contain a clastic component, albeit minor.

#### *Structural barriers*

Faults may act as conduits or barriers to flow, especially where fine-grained fault gouge comprises a relatively



**Table 2** Summary of potential stratigraphic flow barriers within carbonate aquifer rocks in the Ash Meadows region

| Unit                    | Stratigraphy: Ash Meadows–Death Valley region | Comment   |
|-------------------------|---|---|
| Upper carbonate aquifer | Tippipah Limestone                            | –   |
|                         | Siliciclastic aquitard                        | –   |
| Lower carbonate aquifer | Chainman Shale                                | –   |
|                         | Sultan Limestone                              | –   |
|                         | Simonson Dolomite                             | –   |
|                         | Sevy Dolomite                                 | –   |
|                         | Eureka Quartzite                              | ~100-m middle shaley unit (Hunt and Mabey 1966) |
|                         | Pogonip Group                                 | ~85-m middle shaley unit (Hunt and Mabey 1966)  |
|                         | Nopah Formation                               | ~15-m basal shale (Hunt and Mabey 1966)         |
|                         | Bonanza King Formation                        | –   |

impermeable core with enhanced permeability in the fractured rock of adjacent damage zones (e.g., Caine et al. 1996). Other studies indicate that fault cores may be barriers to transverse flow (e.g., Evans et al. 1999; Lachmar et al. 2002; Garringer et al. 2004), including faults in limestones with small offsets (Geraud et al. 2006; Micarelli et al. 2006). Workers in the Neogene clastic sedimentary rocks of the Santa Fe Group, New Mexico, USA, have repeatedly concluded that normal faults comprise barriers to transverse groundwater flow, thereby compartmentalizing aquifer systems (e.g., Minor and Hudson 2000; Caine et al. 2002; Caine and Minor 2005). Similarly, Nelson et al. (2009) indicated that the fault core and impermeable hanging wall strata of the Hurricane fault, a major normal structure at the eastern edge of the Great Basin, impeded flow across the fault as evidenced by the discharge of CO<sub>2</sub> and thermal waters dammed by this structure. In contrast, deep circulation

parallel to the fault, including deep flow within basement rock, is accommodated by the footwall damage zone.

Miner et al. (2007) discussed the relationship between fracture aperture and hydraulic gradient (Fig. 13). From their analysis, two fractures are considered in this study: (1) one perpendicular to the least principal stress vector ( $\sigma_3$ ), and (2) a second parallel to  $\sigma_3$ . Fracture 1 is expected to sustain about the same flux as fracture 2 even if fracture 1 has only twice the aperture and just 10 % of the hydraulic gradient of fracture 2. In most cases, fractures perpendicular to  $\sigma_3$  are expected to accommodate greater fluxes than those parallel to it. When one considers faults oriented perpendicular to  $\sigma_3$ , generally north–south in an actively extending region like the Great Basin, their damage zones are expected to strongly divert flow parallel to active or otherwise favorably oriented faults. In this sense, fault cores do not necessarily need to contain fault gouge to act as a barrier, as damaged bedrock adjacent to faults may act as hydraulic capture zones.

In this context, Bushman et al. (2010) preferred flow from volcanic rocks beneath the Yucca Mountain region to sustain flow at Ash Meadows, largely along the damage zone of existing faults that are in extension. Despite the potential pitfalls of contour maps, the flow path of Bushman et al. (2010) is perpendicular to regional contours (Fig. 8). Faunt (1997) and Ferrill et al. (1999) reached similar conclusions regarding the anisotropy of permeability related to faults in this area. The view of Bushman et al. (2010) is not only consistent with stable isotope values and chemical evolution of groundwater, but the north–south grain of active normal faults may provide preferred pathways for waters to arrive at Ash Meadows from interbasin regions not far removed (Fig. 8).

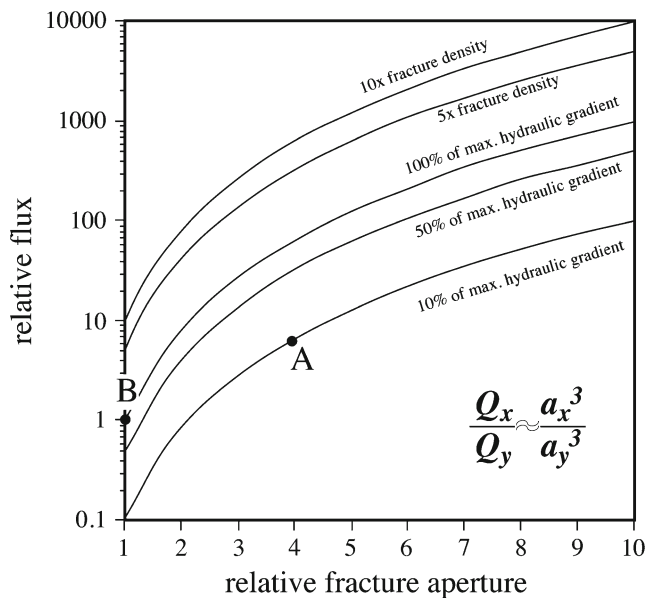
#### Combined structural and stratigraphic barriers

Anderson et al. (2006) evaluated the potential for interbasin flow across the southern Funeral Mountains to the Furnace Creek area of Death Valley as proposed by many researchers (e.g., Eakin 1966; Mifflin 1968; Winograd and Friedman 1972; Winograd and Thordarson 1975; Harrill et al. 1988; Dettinger 1989; Thomas et al.

**Table 3** Summary of potential stratigraphic flow barriers within carbonate aquifer rocks in the Snake Range Region

| Unit                               | Stratigraphy: Snake Range region    | Comment   |
|------------------------------------|-------------------------------------|---|
| Upper carbonate aquifer            | Arcturus Group                      | Contains siliciclastic interbeds (Hintze and Kowallis 2009) |
|                                    | Ely Limestone                       | Contains siliciclastic interbeds (Hintze and Kowallis 2009) |
| Siliclastic aquitard               | Manning Canyon Shale                | –   |
| Lower carbonate aquifer            | Joana Limestone                     | –   |
|                                    | Pilot Shale                         | ~150 m thick (Hintze and Kowallis 2009)                     |
|                                    | Guilmette Formation                 | –   |
|                                    | Sevy Dolomite                       | –   |
|                                    | Laketown Dolomite                   | –   |
|                                    | Eureka Quartzite                    | –   |
|                                    | Pogonip Group                       | Contains shale interbeds (Hintze and Kowallis 2009)         |
|                                    | Notch Peak Formation                | –   |
|                                    | Dunderberg Shale <sup>a</sup>       | ~90 m thick (Hintze and Kowallis 2009)                      |
|                                    | Lincoln Peak Formation <sup>a</sup> | –   |
| Pole Canyon Limestone <sup>a</sup> | –                                   |   |

<sup>a</sup> From Hintze and Kowallis (2009). Omitted in the hydrostratigraphic columns of Sweetkind et al. (2011a)



**Fig. 13** Relative flux in fractures ( $Q_x$ ,  $Q_y$ ) as a function of their apertures ( $a_x^3$ ,  $a_y^3$ ), contoured as a percentage of the maximum hydraulic gradient, based upon the well-known ‘cubic law’ (Bear 1972) where all other variables are equal. Numbers on contours represent the percentage of maximum hydraulic gradient imposed. For example, the lowest curve indicates that a fracture that experiences only 10 % of the maximum hydraulic gradient (point A, for example) may transmit fluxes approaching or even exceeding that of a fracture experiencing the full hydraulic gradient (point B) as long as it has a larger relative aperture

1996; Kirk and Campana 1990; Steinkampf and Werrell 2001; Belcher et al. 2009).

In the context of the control on flow by both fault cores and impermeable strata, interbasin flow seems unlikely in this area. Figure 14 is a cross section across the southern Funeral Range. In order to transit from Ash Meadows to Furnace Creek, water would have to flow transverse to ~10 to 12 cores of major faults, some of which Faunt et al. (2010) specifically identify as potential flow barriers. It is relatively simple to illustrate the improbability of this occurring. If there is just a 50 % chance that a particular fault core is a transverse flow barrier, the probability that flow would be permitted across all these major faults is  $\sim 0.5^{10}$ – $0.5^{12}$ , or 0.1–0.02 %. Additionally, impermeable interbeds in the southern Funeral Mountains, that dip counter to the proposed flow direction, may provide stratigraphic barriers that have been repeated by faulting (Fig. 14).

In map view (Fig. 10b), carbonate rocks in the southern Funeral Range are discontinuous, having been tectonically intermingled with lithologies often considered to be aquitards. There are also many, many more mapped faults than those shown in the idealized cross section (Fig. 14), which suggests, however, that carbonate rocks may be continuous in the subsurface, at least in some areas. Some of these faults are perpendicular to potentiometric contours, especially in the extreme southeastern corner of the Funeral Range (Fig. 10b), and may be favorable to fault-parallel flow. However, there are other areas of the

southern Funeral Range where numerous faults in carbonate rocks are unfavorably oriented. In summary, the authors’ view is that special pleading must be invoked to conceptualize an integrated fracture network through carbonate rock and intercalated impermeable siliciclastic interbeds and faults.

#### Analogue studies

In northeastern Utah, just outside the Great Basin to the northeast lie two areas that have been extensively studied in terms of aquifer partitioning by faults and impermeable interbeds—the Snyderville Basin and the Kamas-Coalville areas (Fig. 1; Ashland et al. 2001; Hurlow 2002). These “back basin” case studies provide a particularly good comparison to the Great Basin. Much of the stratigraphy is similar, except that the thickness of time-equivalent units are thinner in the Snyderville basin and Kamas-Coalville area than in the Great Basin (Hintze and Kowallis 2009), and the Snyderville Basin and Kamas-Coalville area have a greater proportion of siliciclastic rock than does the Great Basin. The tectonic setting and history is also similar, including Neogene extension, except that total fault offsets are much smaller. As such, these areas are excellent small-scale analogues for the Great Basin.

Hurlow (2002) and Ashland et al. (2001) conclude that groundwater in fractured limestone and sandstone is stratigraphically compartmentalized by impermeable interbeds. Low porosity units comprise barriers to flow perpendicular to bedding with some fractured shales having permeabilities nearly as low as intact rock. Faults, on the other hand, generally act as barriers to perpendicular flow but as conduits to fault-parallel flow. As a result, individual groundwater compartments are separated from one another by a combination of faults and bedding.

The hydrogeology of these marginal back basins offer a very different conceptual model of flow for areas that might be expected to act as smaller-scale models of the nearby Great Basin. Particularly compelling are observations provided in Ashland et al. (2001; and references therein). In addition to field observations, pumping tests showed a lack of interaction between compartments. This area also includes the major historical Park City mining district. The history of the construction of thousands of meters of drifts and shafts, and combined water inflows, is consistent with aquifer segmentation by both stratigraphic and structural elements (Ashland et al. 2001).

#### Summary

Where multiple faults and impermeable interbeds must be crossed at a high angle, the potential for interbasin flow is greatly diminished. Highly faulted terranes like the Great Basin should exhibit segmented bedrock aquifer systems. By contrast, according to the Bushman et al. (2010) model, interbasin transfers of waters to Ash Meadows were guided by flow along favorably oriented fracture networks, consistent with isotopic evidence and solute evolution.



fracture heterogeneities must cancel out such that on some scale the carbonate rock is an equivalent porous medium.

- Fractures with aperture must exist. They have not been sealed by secondary minerals nor are they closed due to lithostatic loads.
- Impermeable interbeds must be capable of sustaining and retaining open fractures or are bypassed by a tortuous flow system.
- Faults, especially those with favorable orientation with respect to the stress field, may divert water into their damage zones, favoring fault parallel flow. Fault cores may be flow barriers, but damage zones containing fractures with aperture will tend to capture flow. Thus, interbasin flow is most likely to occur where flow is channeled by fracture systems with aperture.

In summary, bedrock interbasin flow may occur wherever potential and integrated permeability permit. However, as the spatial dimensions of an aquifer system increase, the probability of an integrated flow network being present in bedrock declines. In the absence of compelling evidence to the contrary, the complex geology of regions like the Great Basin of the USA suggests that the first working hypothesis adopted should investigate local recharge through the mechanisms of mountain front recharge, mountain block recharge, and losses from both intermittent and perennial streams (Fig. 9). Even at local scales, the influence of climate variation may be needed to explain isotopic compositions and long apparent residence times. This is a parsimonious approach relative to first assuming interbasin flow.

Where interbasin flow occurs on a large scale, like Ash Meadows, fault-produced anisotropy in bedrock fractures is likely to control flow directions. Interbasin flow paths that cross multiple faults and mountain ranges at high angles should probably be the conceptual model of last resort.

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