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Abstract

We summarize the literature relevant to the hydrogeology of the Española Basin, in Northern New Mexico, and present new analyses of water level and streamflow data to define regional flow patterns and to quantify pre-development fluxes into and out of the regional aquifer. At the basin-scale, groundwater flow is generally topographically driven, with strong upward hydraulic gradients present near the regional discharge zone, the Rio Grande. Although many complex geologic features exist within the basin, available water level data are generally insufficient to demonstrate their hydrologic significance. We estimate that the long-term, average rate of recharge to the entire basin is approximately 3.3 – 4.0 m^3/s. Most of this water discharges to various stream reaches throughout the basin; approximately 0.40 m^3/s discharged historically to the Rio Grande in the southern portion of the basin where population density is now the highest. In this southern region, approximately 0.42 m^3/s is currently being withdrawn by municipal wellfields. Groundwater withdrawals, therefore, are relatively small in relation to total aquifer recharge at the basin scale but are large in relation to baseflow discharge in this region. Therefore, impacts to surface flow may be significant

Introduction

Like many basins in the semi-arid southwestern U.S., the Española Basin is experiencing rapid population growth and is heavily dependent on groundwater resources for water supply, Estimating the impact of groundwater withdrawals on quantities such as aquifer storage and surface water flow is necessary for water resource planning, and requires a basin-wide understanding of aquifer characteristics and groundwater flow. In particular, aquifer recharge is important to estimate, not only to evaluate "sustainable" water supply but also to constrain groundwater flow models which, in turn, can play important roles in evaluating various water use scenarios. Although the relevance of recharge to estimation of "sustainable yield" has been questioned (Bredehoeft, 1997), recharge rates remain a critical factor in determination of aquifer and surface water responses to long-term pumping.

The purpose of this paper is twofold: first, to summarize much of the relevant literature concerning the hydrogeology of the basin and previous estimates of aquifer discharge and recharge; and second, to present new analyses which provide estimates of recharge rates at the basin-scale. In order to accomplish this, we focus on estimatation of long-term average fluxes which approximate "pre-development" (or steady-state conditions) in the basin. For comprehensive descriptions of post-development stresses to the aquifer and to the surface water flows throughout the basin, we refer the reader to Duke Engineering and Services (2000) and Brian and Wilson (1997).

setting

The Española basin is located in northern New Mexico, USA, It is one of a series of basins located within the Rio Grade Rift zone, shown in ***Figure*** 1, a tectonic feature that extends from northern Colorado south into Mexico. Elevations within the basin range from more than 3,800 m along peaks in the surrounding mountain ranges to about 1,700 m at the basin surface-water outlet. Vegetation within the basin varies with elevation, with ponderosa pine, spruce and fir, aspen, and alpine grasses found with increasing elevations above 2,300 m and pinyon pine, juniper, and grasses dominant with decreasing elevations below abount 2,300 m (Spiegel and Baldwin, 1963, p. 17). The largest cities in the basin are Santa Fe, Española, and Los Alamos; numerous Indian pueblos are also present. Total population of the area is 130,000 of which about half lives in the City of Santa Fe. Some portions of the basin are

in agricultural use; Landsat images indicate that the irrigated areas are mostly located in the northern part of the basin along the Pojoaque River, the Santa Cruz River, the Rio Chama, and the Rio Grande, and in the southern part of the basin along the Rio Grande south of Cochiti Reservoir (Duke Engineering & Services 2000, fig. 3-3). Estimates of groundwater use compiled by Duke Engineering Services (2000, table 5-6) for an area slightly smaller than the basin totaled 0.68 m^3/s. Roughly 20 percent of this amount or 0.13 m^3/s is returned to the aquifer by seepage from irrigation or domestic and municipal wastewater.

The Española basin and surrounding areas receive annual total precipitation ranging from 18 to 86 cm/yr. As shown in *Figure* 2, precipitation is strongly elevation dependent The largest streams in the basin are the Rio Chama and Rio Grande. Median monthly flow, calculated using USGS average monthly flow data for the past 80 years, is 26.0 m^3/s along the Rio Grande (at Otowi Bridge) and 10.0 m^3/s along the Rio Chama (at Chamita). (Note: 1 m^3/s equals 35.31 cubic feet per second (cfs) or 25,583 acre-ft/year). Numerous tributaries enter these rivers; many of these are ephemeral and many are ungaged. The Rio Grande and the lower reaches of many tributaries comprise the regional groundwater discharge zone.

In most parts of the basin, the water table is 0-60m below ground surface; on the Pajarito Plateau the water table is much deeper (up to 350m below the surface). Throughout much of the basin, the water table appears to intersect the surface at the Rio Grande. Perched waters exist on the Pajarito Plateau (Purtymun 1984), where the unsaturated zone is much thicker than in other parts of the basin.

The extent to which groundwater flows between Española Basin and adjacent basins is unclear. Unfortunately, water level data is sparse along all basin margins and thus it is impossible to definitively locate groundwater divides. McAda and Wasiolek (1988) estimated approximately 0.480 m^3/s water flowing south to the Albuquerque basin. No published estimates exist for fluxes from the San Luis and San Juan basins to the north; however, Santa Fe Group sediments are thin along the northern basin margins and hence fluxes into the basin from the north may be relatively small. To the east, it is reasonable to assume that the topographic divide along the Sangre de Cristo Mountains defines both a surface water and groundwater divide. Some groundwater, however, may enter the basin from the northeast in the Picuris embayment.

To the west, surface water divides in the Jemez Mountains may also define groundwater divides. Hydrologic separation of waters in the Valles caldera from those beneath the Pajarito Plateau has been postulated based on the presence of numerous large faults and on geochemical evidence (Blake et al. 1995). Insufficient

hydrologic data are available, however, to confirm this conceptual model. Until such data become available, we include the caldera in our analysis of the basin and make no *a priori* assumption about any hydrologic connection (or lack thereof) with groundwaters to the west.

Hydrologic characteristic of basin sediments

The regional aquifer in the basin lies almost entirely within the Santa Fe Group rocks, with the exception of some areas in the Jemez volcanic field where both volcanic and volcaniclastic rocks (Puye Formation, Cerros del Rio basalts) are saturated as well as portions of the Sangre de Cristo Mountains, where groundwater occurs in fractured Precambrian rocks, Although clays and gravels are common in the Santa Fe Group, aquifer sediments are predominantly sands and silts. In general, sediments are weakly consolidated but in some cases are cemented with secondary calcite. Based on hydraulic testing, estimates of hydraulic conductivity in the basin range from 0.1-9.8 m/day (Hearne 1980; Purtymun 1984; Hearne 1985; Frenzel 1995, Daniel B. Stephens, 1994). The Tesuque Formation, comprising the largest aquifer in the basin, is strongly anisotropic. This is due to numerous alternating sequences of siltstones and sandstones that dip to the west at angles ranging from 0 - 30° (Golombek, 1983). Estimated ratios of vertical to horizontal conductivity range from ,001 to .226 (Hearne 1980; Hearne 1985; Frenzel 1995).

Groundwater flow directions

In areas with significant regional topographic relief, such as the Española Basin, regional flow systems tend to develop with groundwater recharging in the highlands and discharging in the valleys (Toth 1963; Freeze and Witherspoon 1967). This general flow pattern is sometimes called "topography-" or "gravity-driven" flow (Person et al. 1996). Other factors such as heat flow, stratigraphy, structure, and local topographic features can cause a flow system to depart from the simple model of regional topographic-driven flow. In this section we summarize previous studies, present water level data at the basin-scale, and discuss the relative importance of these factors on flow in the Española Basin.

Several regions within the basin have been the focus of hydrologic studies, Coon and Kelley (1984) studied groundwater flow in the upper Rio Grande basin (north of the Colorado border to just south of Española). Based on water table contours and surface water flow measurements, they concluded that flow in this area is

predominately away from the basin margins and towards the Rio Grande. An exception to this gravity-driven flow system is the Embudo constriction, where structural features in the underlying rock force flow southwest, parallel to the Rio Grande. McAda and Wasioek (1988) and Frenzel (1995) have constructed numerical models of groundwater flow in the southern half of the basin. These models, which are consistent with measured heads in the Los Alamos and Santa Fe areas, suggest strong gradients towards the Rio Grande in the southern portion of the basin. Detailed studies of the groundwater system within the Pajarito Plateau (Purtymun 1984; Rogers et al. 1996b) depict a water table that slopes towards the Rio Grande; however, the magnitude and direction of deeper gradients are unclear. The picture that emerges from these three studies is one of gravity-driven flow, away from the highlands and towards the valley in all portions of the basin, except the Embudo district where a shallow bedrock constricts flow and forces it south.

Hydraulic gradients and heat flow

Figure 3 presents contours of water levels in wells throughout the basin. Data were compiled from many sources (Blake et al. 1995; Purtymun 1995; LANL 1997; U.S. Bureau of Indian Affairs 1997; U.S. Geological Survey 1997) This map, which includes water levels in wells drilled to a large range of depths, depicts the apparent horizontal component of hydraulic gradients; vertical hydraulic gradients that are known to exist within the basin are incorporated in these apparent gradients. These gradients suggest that the horizontal component of flow is predominately southerly in the northern portion of the basin and predominately east-west or west-east in the southern half of the basin. These trends are consistent with previously published water-level contour maps for the Pajarito Plateau (Rogers et al. 1996b; Purtymun, 1984) and with modeling studies in the southern half of the basin (McAda and Wasiolek 1988; Frenzel 1995).

Most wells in the basin are relatively shallow. Deep wells exist, but many of these have very long open intervals and thus it is difficult to assess vertical hydraulic gradients in the basin. *Figure 4* presents a vertical profile of head data extending from the Pajarito Plateau to the Santa Fe area. It is important to note that many wells do not lie strictly within the plane illustrated in this figure; however, since the transect is oriented approximately perpendicular to water level contours *(Figure 3)* this approximation is reasonable. This figure illustrates the upward flow near the Rio Grande, which causes artesian conditions in many wells along the river and in some wells in lower Los Alamos Canyon.

Limited heatflow data collected within the vicinity of the Española basin has been summarized by Reiter et al.(1975). All heatflow data collected within the basin (vicinity of Santa Fe and Buckman well fields) indicate relatively low heat flux (< 2 μcal/cm^2-sec or 0.084 Watts/m^2) in the basin. High heat flux values were measured in three locations along basin margins: the Jemez Mountains, the far northeastern corner of the basin near the town of Dixon (Reiter et al. 1975), and at Ojo Caliente (Trainer and Lyford 1979). Since these data are sparse, it is difficult to draw conclusions regarding the importance of heat flow on the regional groundwater flow patterns.

Geologic controls

Various theories concerning geologic control on groundwater flow have been hypothesized for the Española basin, including confining beds resulting in artesian conditions (Purtymun and Johansen 1974), buried high-permeability ancestral Rio Grande deposits causing flow parallel to the Rio Grande (Hawley, pers. comm., 1998), dipping Santa Fe Group beds, causing preferential flow parallel to the strike (Hearne 1985) and cemented fault zones acting as barriers to flow (Spiegel and Baldwin 1963; Blake et al. 1995). At the local scale, it is likely that any or all of these factors may play an important role. However, the influence of these factors on regional flow patterns is unclear. In this section, we summarize available data on geologic controls on groundwater flow in the basin.

The presence of vertical hydraulic gradients and flowing wells near the Rio Grande has led some to conclude that the aquifer is confined or semi-confined (Purtymun and Johansen 1974). In a study of the Buckman well field (shown in ***Figure 5***), Shomaker (1974) describes a "lower" aquifer and "upper" aquifer, separated by a confining bed. However, upward head gradients in the vicinity of a river will also produce flowing wells in an unconfined aquifer; this effect is often erroneously attributed to confining pressures (Freeze and Cherry 1979).

To further explore the possibility that a lower "confined" aquifer exists, we searched published reports and datasets for the following types of evidence: geologic evidence of low-permeability, laterally continuous confining beds, documentation of upward vertical hydraulic head gradients in areas far from groundwater discharge zones, and examination of pump test results. Regarding confining beds, clay and shale layers have been reported on well logs at numerous depths throughout the basin, yet these layers are rarely continuous for more than a mile or so (Galusha and Blick 1971; Hearne, 1985). Even within a small area such as the Buckman well field, "confining" layers cannot be correlated between wells (Shomaker, 1974). It is unclear whether these low-permeability layers have sufficient

continuity to produce confining conditions. Upward vertical gradients have been documented in the Santa Fe area (Wasiolek 1995) and near Tesuque (Hearne 1980), neither of Which are near a groundwater discharge zone. Wasiolek (1995) suggests that the higher heads at depth are caused by a connection between deeper beds in the Santa Fe Group with the primary recharge zone upgradient (the Sangre de Cristo Mountain front). This argument implies that bedding structures within the Santa Fe Group are sufficiently continuous and impermeable so as to prevent redistribution of water between adjacent beds. Hearne (1980) explains the observations on Tesuque pueblo land by a confined or semi-confined aquifer. The strongly anisotropic Santa Fe Group sediments may, in some portions of the aquifer, produce behavior similar to that of a "leaky" aquifer (Stoker et al. 1989).

Pump test results in wells throughout the basin indicate a wide range in aquifer characteristics. Tests conducted in water supply wells on the Pajarito Plateau indicate either "leaky aquifer" conditions (Stoker et al. 1989) or unconfined conditions (Purtymun et al. 1990). Storage parameters derived from pump tests within Santa Fe County, compiled by Daniel B. Stephens & Associates (1994) span a wide range of values, indicating both confined and unconfined conditions exist within the Santa Fe Group. Lack of hydraulic connection between pumping wells in one layer and nearby observations wells in deeper or shallower layers has been observed in many tests (Hearne 1980; Daniel B. Stephens & Associates 1994; Purtymun et al. 1995), a further indication that either local confining conditions or very low vertical permeability values are common in the basin.

Very little evidence exists regarding the possible influence of buried gravels deposited by the ancestral Rio Grande. Such deposits have been shown to be have higher hydraulic conductivities than most of the sediments in the Albuquerque basin to the south (Kernodle and Thorn 1995). Purtymun (1995) identified a relatively permeable trough, oriented north-south, in the upper Santa Fe Group beneath the Pajarito Plateau. Rogers (1996b) and others have suggested that this feature may cause groundwater to preferentially flow southward beneath the plateau. However, numerous wells are partially completed within this formation and proximal to this unit; water level gradients between these wells do not indicate a significant southerly component to flow.

Rifting has produced many faults within the basin, including the Pajarito Fault zone (to the west), the La Bajada fault zone (to the south), and a large number of north-south trending faults within basin-fill sediments (Kelley 1978). Numerous hypotheses have been suggested regarding the hydrologic significance of these faults; including "faults-as-barriers-to-flow" (the Pajarito fault zone (Goff and Sayer 1980), Guaje Canyon (Griggs and Hem 1964), and Santa Fe vicinity (Spiegel and Baldwin 1963)) and "faults-as-conduits for flow" (rising thermal

waters along the Rio Grande (Blake et al. 1995)). Direct evidence of faults serving as barriers to flow has been documented in the Santa Fe area where water level data are sufficiently dense and in the La Cienega (Spiegel and Baldwin 1963) and Ojo Caliente (Vuataz et al. 1984) areas where springs occur near faults. Elsewhere in the basin, direct hydrologic evidence of faults affecting flow is lacking.

In summary, regional trends in water levels *(Figures 3, 4, and 5)* are consistent with a simple conceptual model of topographic-driven flow. At local scales, however, the importance of faults as barriers to flow has been demonstrated (Spiegel and Baldwin 1963). There is no physical evidence of a widespread, laterally continuous confining unit within the Santa Fe Group and so the concept of a regionally extensive confined aquifer is untenable. More likely, fine-scale layering within the Tesuque Formation causes permeability to be strongly anisotropic with high horizontal to vertical permeability ratios. Artesian conditions reported in many wells near the Rio Grande may be caused by upward hydraulic gradients associated with groundwater discharge rather than by confining beds. Pump tests from some wells in the basin do indicate confined or leaky-aquifer behavior and so confining conditions are probably present in some locations. Documentation of vertical upward hydraulic gradients (Hearne 1980; Wasiolek 1995) in areas far from the regional discharge zone is additional evidence of local confining conditions.

Groundwater Fluxes

In this section, we discuss rates of aquifer recharge and discharge in the basin. In general, the available datasets (water level, streamflow, and springflow data) do not have sufficient resolution to be used to identify temporal trends that may be present due to climatic or water use trends. Rather, they are only sufficient for estimating long-term, average flux estimates. Since many of the data were collected either far from major municipal well fields or precede the recent rapid population growth in the southern portion of the basin, the flux estimates we present are approximations to natural conditions in the basin, before groundwater withdrawals substantially affected groundwater discharge to surface water courses, During this "pre-development" period, climatic fluctuations occured which may have caused an imbalance between aquifer recharge and aquifer discharge (to rivers) in any given year. However, over the coarse of decades we presume that, on average, aquifer discharge was approximately equal to aquifer recharge. Therefore, our estimates of discharge are approximations to average rates of aquifer recharge at the basin scale.

Aquifer Recharge

In semi-arid basins, recharge is thought to originate primarily in adjacent mountain ranges which receive relatively high rates of precipitation (Duffy and Al-Hassan 1988). In alluvial basins flanked by crystalline mountains, this recharge water enters the basin-fill sedimentary rocks in the subsurface along the mountain fronts (hence the term "mountain front recharge" (Wasiolek 1995)). In addition to mountain front recharge, some recharge may occur within the basin along stream channels (Anderholm, 1994), particularly in the basin highlands. Finally, diffuse areal recharge may also occur at very low rates.

Since it is impossible to measure groundwater recharge directly, various indirect methods are used to identify the location of recharge zones and to estimate rates. As a result, various studies may produce widely ranging estimates for the same region. The following sections describe studies in the Española Basin that have shed light on recharge rates, using chloride mass-balance, water budget, and groundwater flow modeling techniques.

Chloride Mass-Balance Method

This method is used to estimate the fraction of precipitation that escapes evapotranspiration to become net recharge, based on mass-balance calculations for chloride. Assuming chloride is conserved in the system, the relative concentration of chloride in precipitation (C_p) and in groundwater (C_r) can be used to estimate recharge rates by the following equation:

$$PC_p = RC_r,$$

where R = recharge rate (cm/yr)

P = precipitation rate (cm/yr)

C_p, P, and C_r can be directly measured thus, R can be estimated from this equation. Although simple in principle, application of the chloride mass-balance method is complicated by anthropogenic sources of chloride, such as septic system effluent and road salt. Appropriate selection of representative groundwater and precipitation data is critical to the success of the method.

Mountain-front recharge for the Santa Fe River, Rio Tesuque and Arroyo Hondo drainages on the western slope of the Sangre de Cristo Mountains was estimated by Anderholm (1994) using the chloride concentration of shallow ground water. In this study, mountain-front recharge is defined to include focussed stream-channel recharge and diffuse bedrock recharge upgradient from the sampling location. Selection of representative chloride

concentrations in groundwater was problematic, since concentrations were highly variable within the individual drainages (Anderholm, 1994, plate 1). This may be due to anthropogenic sources and/or transient recharge events. Anderholm selected the lowest observed concentrations to represent pre-development conditions. Corrected where necessary for surface runoff, the estimated recharge was 10, 7, and 10 percent of the total precipitation falling on the Santa Fe River, Rio Tesuque, and Arroyo Hondo drainage basins (see ***Table 1***).

Anderholm (1994) also estimated recharge rates in several arroyos and in low-elevation areas undissected by arroyos. Unsaturated-zone chloride and soil-moisture data from three boreholes in the undissected areas indicated recharge rates less than 0.1 mm/yr for the past 6,700 to 8,800 years and only slightly higher (1 to 2 mm/yr) recharge rates prior to these dates (Anderholm, 1994, fig. 7). The higher recharge rates indicated by the data from the deeper parts of the boreholes are presumed to represent pluvial conditions associated with the late Pleistocene. Similar unsaturated-zone data from beneath two arroyos produced lower-bound estimates of recharge in the arroyos of about 3 mm/yr.

To evaluate the relative importance of arroyo recharge and mountain-front recharge, Anderholm (1994, p. 35) examined chloride data from wells in the regional discharge area (Buckman well field). These waters had a chloride concentration of only about 5 mg/L, about an order of magnitude lower than the estimated concentration of arroyo recharge. Anderholm concluded that arroyo recharge is probably insignificant compared to overall recharge within the Española basin. However, ground water from the Buckman well field has the most depleted deuterium and oxygen-18 values in the basin, indicating it may have been recharged under a climate that was cooler (and wetter) than the present-day climate (Anderholm, 1994, p.45). Thus, even if substantial arroyo recharge were occurring under the present climate, its absence in ground water at the Buckman field might be explained by the fact that it has not yet reached the discharge area.

Applying this method to the Pajarito Plateau, we examined chloride concentrations and discharge measured at twenty springs in the west side of White Rock Canyon in the early 1960's reported by Purtymun (1980). These data result in a discharge-weighted chloride concentration for the springs of **4.2** mg/L. Using the chloride mass-balance equation and assuming an average precipitation for the Pajarito Plateau of 375 mm/yr, a volume-weighted average chloride concentration in precipitation (C_P) of 0.30 mg/L (Adams et al. 1995), and a chloride concentration for the recharge (C_r) of **4.2** mg/L gives a calculated areally-averaged recharge rate of **28** mm/yr or about **7%** of total precipitation. Although a recharge rate equal to *7%* of precipitation is a reasonable average value in the context

of recharge estimates that have been made for other parts of the Espanola Basin (see ***Table 1***), this average value does not indicate how recharge is distributed on the Plateau, Evidence provided by a number of studies suggests that most of the groundwater beneath the Plateau originates from infiltration at high elevations in the Sierra de los Valles immediately west of the Plateau or from infiltration along deep canyons that have been incised into the plateau, with very little net infiltration or recharge beneath mesas on the Plateau itself (Blake et al. 1995; Newman 1996; Rogers et al. 1996a; Rogers et al. 1996b; Newman et al. 1997)

We use a similar method to estimate average recharge within the Los Alamos Canyon watershed. We applied the chloride mass-balance method using a groundwater chloride concentration of 3.0 mg/L measured in a well (Test Well 3) just downgradient from the Los Alamos Canyon study area. This chloride value is 10 times greater than the average value in precipitation of approximately 0.3 mg/L (Adams et al., 1995), implying that recharge in the watershed averages 10 percent of total precipitation.

Water-Budget Methods

Water budgets were estimated for five drainage basins on the western slope of the Sangre de Cristo Mountains by Wasiolek (1995). The basins included the Rio Nambe, Rio en Medio, Tesuque Creek, Little Tesuque Creek and Santa Fe River drainages. Recharge in these basins was calculated as the residual between measured or estimated values of precipitation minus the sum of surface runoff plus evapotranspiration. Water budgets were calculated for winter, spring and summer/fall periods and summed to obtain annual totals. Effective precipitation was calculated for the winter months by adjusting the estimated precipitation to account for the sublimation of snow. Evapotranspiration was estimated with curves describing the seasonal relations between evapotranspiration and precipitation that had been established for the Rocky Mountain region; these curves also accounted for the effects of slope aspect on evapotranspiration. Like Anderholm (1994), Wasiolek (1995) concluded that most recharge in the mountain drainages occurred during winter months, which were considered by Wasiolek to extend from October through February. The estimated volumes of recharge and the percent of the total precipitation falling on the basins that becomes recharge are listed in ***Table*** 1.

The recharge estimates made by Wasiolek (1995) for the Rio Tesuque and Santa Fe drainage basins are high compared to the estimates made by Anderholm (1994) for the same basins, Some of the discrepancy between the recharge estimates may by due to the fact that the two studies considered slightly different basin areas and

precipitation amounts. More likely, the discrepancies for the recharge estimates for the two models arise from (1) the difficulty in choosing representative natural chloride concentrations when using the chloride-mass balance method and (2) the use of regional rather than site specific precipitation and evapotranspiration relations in the water-budget method.

Gray (1997) presented a detailed water budget analysis for the Los Alamos Canyon watershed on the Pajarito Plateau. Separate budgets were calculated for each of three water years (1993 – 1995). One significant different between this method and Wasiolek's is that Gray used site-specific measurements of evapotranspiration. For the three years, Gray estimates recharge rates to range from 10.2 to 18.5 cm/year, or 17 to *26%* of precipitation.

Water budget components from all these methods are compiled in ***Table 1*** and presented in ***Figure 5.*** The average recharge rate is approximately 13% of total precipitation. There is no evidence that elevation controls the fraction of precipitation that becomes recharge. Other factors, such as geology, slope, aspect, and vegetation, might explain the variation shown in ***Figure 5.***

Ground-water flow modeling

Groundwater models can provide additional information about recharge because of the inherent mass-balance constraints and the relationship between recharge rates, permeability, and water levels that influences the model calibration process. By varying permeability values (within data-defined ranges) and recharge rates to seek the most accurate simulated water levels in wells, the model can provide bounds on plausible recharge rates. Most models also incorporate some type of independent estimates of recharge or discharge either formally in the calibration process or as a posterior check on model validity.

Ground-water flow in the southern part of the Española Basin was previously modeled by McAda and Wasiolek (1988) and by Frenzel (1995). The model of Frenzel was essentially a refinement of the model previously developed by McAda and Wasiolek (1988). Their model domains extended from near the city of Española on the north, La Cienaga on the south, the Pajarito Fault on the west and contact between the basin-fill materials and the Pre-Cambrian bedrock on the east. Most of the recharge to the aquifer was assumed to occur as mountain-front recharge and was applied as specified flux along eastern and western boundaries and to grid elements corresponding to stream reaches close to these boundaries. The total mountain-front recharge in the eastern basin is 0.75 m^3/s, and includes a larger area than that studied by Wasiolek (1995) where recharge totaled 0.58 m^3/s. Areal recharge was

estimated by McAda and Wasiolek (1988) to range from 0.127 to 1.02 cm per year (**0.38** cm/yr on the Pajarito Plateau). The nonzero recharge rates in the low elevation areas within the basin can be reconciled with the conclusion of Anderholm (1994) that recharge in the undissected areas is zero by noting that the low elevation areas are criss-crossed by numerous arroyos that probably contribute some recharge. Recharge rates used by Frenzel (1995) were somewhat smaller than recharge rates assumed by McAda and Wasiolek (1988). For example, Frenzel (1995) assumed that recharge on the Pajarito Plateau was only 0.05 cm/yr.

A very different approach was employed by Hearne (1985) in his model of the aquifer of the Pojoaque Basin. The specified mountain front recharge (along a slightly longer boundary than that defined by Frenzel) was 0.107 m^3/s, in addition to 0.441 m^3/s recharge along stream channels within the model domain, totaling 0.548 m^3/s recharge to the eastern basin. No areal recharge is specified in the model. Because of lower recharge rates, this model predicts much less baseflow to the Rio Grande than the models of Frenzel (1995) and McAda and Wasiolek (1988). In general, the McAda and Wasiolek model is most consistent with streamflow data north of Otowi Bridge and the Hearne model is most consistent with streamflow data south of Otowi Bridge. Further discussion of streamflow data will be presented below.

Aquifer discharge

Discharge to rivers

The Rio Grande and the low elevation reaches of its tributaries constitute the discharge areas for groundwater in the basin. It is very difficult, however, to accurately quantify the amount of baseflow the aquifer contributes to streamflow. A wide variety of techniques are commonly used, including indirect methods based analyses of streamflow records (U.S.Department of Justice, 1996; Spiegel and Baldwin, 1963) and water level measurements in wells (Halford, 2000) and direct methods such as seepage meters (Griggs, 1964; Purtymun, 1966). All of these methods are associated with some degree of error and the various techniques often produce quite disparate estimates (Halford, 2000). For the puposes of this paper, we present results from a very simple method which assumes that annual baseflow to a given stream reach can be approximated by subtracting the total January streamflow between two gages (Spiegel and Baldwin 1963; U.S. Department of Justice and New Mexico State Engineer Office 1996). In the Espanola Basin, monthly streamflow is at or near the annual minimum in January and

processes other than baseflow which affect streamflow gain or loss, such as evapotranspiration, surface runoff, and irrigation diversions are at a minimum. This method has also been used by Speigel and Baldwin (1963) to estimate evapotranspiration losses along the Santa Fe River by comparing low flows from July with those from January.

Table 2 provides the mean January flow for each gaging station in the vicinity of the Española basin; gage locations are shown in *Figure 6*. As shown in *Figure 7*, there is a strong correlation (r^2=0.98) between drainage area (reported by the U.S.G.S.) and mean January flow. Predicted mean January flow for each gaging station, based on drainage area using this regression analysis, is also provided in *Table* 2 for reference. We can also use this relationship to estimate the total baseflow discharge to all reaches within the basin, using the basin area defined by the polygon in *Figure* 3. This estimate is 3.50 m^3/s. Errors inherent in this estimate include departures from the drainage area-January flow relationship, streamflow measurement error, and processes other than baseflow contributing to stream flow gain/loss in January.

To determine the baseflow contribution to various reaches within the basin, we subtract the measured January flow between subsets of gages for each year that records exist. For reaches with headwaters within the basin, we assume that January flow equals total baseflow for that reach and no subtraction is necessary. *Table* 3 lists the eight reaches for which these analyses were performed. Reach 2 was initially subdivided into two reaches, separated by the gage above San Juan Pueblo (8281100, *Figure 6*). However, detailed analysis of the data at this gage and the resulting predictions of baseflow gains/losses above and below the gage lead us to believe that flow measurements at this gage were insufficiently accurate to be used in these analyses. Therefore, only calculations for total gain/loss from Rio Grande at Embudo and Rio Grande at Otowi (Reach 2) are presented below.

The gains and losses for these reaches exhibit substantial year-to-year variability *(Figure 8)*. We used several statistical tests to determine if this variability was related to factors such as precipitation or production from nearby supply wells or, alternatively, if the variability was random and thus reflective of measurement error. We expect measurement errors to be particularly significant along reaches where calculated baseflow is a very small percentage of total flow (e.g. 5% for Rio Chama). The tests conducted included (1) runs tests and autocorrelation tests to determine if the temporal patterns exist that depart from completely random behavior, and (2) cross-correlation tests to determine if the variability in stream-flow gains along a particular reach could be related to temporal variations in streamflow gains along other reaches or to other hydrologic variables.

There were several interesting results from these analyses. Regression analysis showed no evidence of linear temporal trends (increases or decreases) in annual gains/losses in any reach except for the lower Santa Fe River (see below). In addition, for most reaches correlations between annual gain and hydrologic variables (precipitation, pumping in nearby well fields, streamflow, and calculated gains/losses in nearby reaches) were statistically insignificant or very low, and thus most of the apparent interannual variability is presumably due to measurement error. Reach 1 proved to be an exception to this rule Cross-correlation analysis established that statistically significant correlations exist between annual streamflow gains along reach 1 and two factors: 1) pumping in the Santa Fe area in the previous 12 months ($r = -0.48$) and 2) annual precipitation at Espanola two years earlier ($r = 0.46$). This is evidence of weak but relatively rapid response of baseflow to pumping and precipitation.

Table 3 presents the mean baseflow gain calculated for these eight reaches. Baseflow gains are slightly larger in the northern portion of the basin (Chama River, Rio Embudo, and upper reach of Rio Grande). This portion of the basin, north of Española, represents half of the total basin area, yet represents almost 60% of the total groundwater discharge. This trend is probably due to climatic factors (increasing temperatures, decreasing rainfall to the south) and groundwater withdrawals in the south where population density is higher. Total baseflow gain for the basin (3.90 m^3/s) is approximately 16% higher than that calculated using basin area (*Table 2)*. This level of agreement, however, is acceptable given the uncertainties associated with these flux calculations.

Table 4 presents the results from *Table 3* in the context of similar estimates previously reported. For the various reaches along the Rio Grande, reported gains vary from 0.005 to 0.014 $m^3/s/km$. Two reaches of the Rio Grande have received the most study. For the Otowi Bridge to Cochiti reach, estimates range from 0.009 to **0.023** $m^3/s/km$. For the Embudo to Otowi Bridge reach, estimates range from 0.013 to 0.029 $m^3/s/km$. Possible causes of the variation include differences in methodology (seepage meters, streamflow data analysis, and groundwater flow modeling), differences in time periods considered, and criteria used to eliminate suspect data.

The Pojoaque River and its tributaries and the Santa Fe River are thought to include both upper reaches that recharge the groundwater system and lower reaches that receive groundwater discharge. Streamflow analysis for the Santa Fe River suggests that, on average for the past 30 years, this reach gains 0.29 m^3/s, with January flows showing a weak increasing trend with lime ($r = 0.57$). This mean value is greater than the flow of 0.18 m^3/s measured by Spiegel and Baldwin (1963, p. 188-192) along the Santa Fe River below Cienaga in January, 1952. The increase over time is presumably due to the population increase in Santa Fe and the greater amounts of water

discharged by the city sewage treatment plant into upgradient reaches of the Santa Fe River.. Regional models (Hearne, 1985 and McAda and Wasiolek , 1988) predict 0.12 and 0.18 m^3/s, respectively, pre-development baseflow to this reach.

Discharge to springs

Springs exist in a number of locations throughout the basin, including La Cienega, White Rock Canyon, numerous canyons on the Pajarito Plateau, and in the Jemez Mountains. However, very few quanititative springflow measurements have been reported in the literature. Flow rates have been reported for springs in White Rock Canyon and on the Pajarito Plateau (Purtymun 1966; Blake *et* al. 1995; U.S. Geological Survey 1997). Using these published datasets, we calculate the long-term average total discharge from springs on the Pajarito Plateau to be approximately .0096 m^3/s. For White Rock Canyon springs, total discharge is approximately .091 m^3/s. Flow rates of .18 m3/s were reported by Spiegel and Baldwin (1963) €or springs at La Cienega.

Discharge to wells

In the Española Basin, significant pumping began in the 1940's and has increased steadily to the present. As shown in ***Figure 9,*** total withdrawals from three major wellfields (Los Alamos, Buckman, and Santa Fe) have gradually increased to over 0.42 m^3/s. Municipal water usage by the city of Española is currently estimated to be 0.046 m^3/s and with irrigation and domestic wells, total groundwater pumping within the basin excluding the Chama River area is estimated to be 0.68 m^3/s (Duke Engineering and Services, 2000, table 5-6).

In response to pumping water levels have declined significantly, up to several hundreds of feet in some areas. Water level declines in a given well will be influenced by many factors, including pumping rates in nearby wells, local and regional rates of aquifer recharge, and hydrologic characteristics of rocks in the vicinity of the well. Water levels have declined in some portions of the Los Alamos well fields by as much as 30 m; water levels in the Buckman wellfield have declined by over 90m. Although these declines are small compared to the total aquifer thickness in this portion of the basin, continued pumping and associated water level declines may enhance upward flow of deeper waters with high levels of dissolved solids and trace elements; hence, water quality may decline.

Basin ground water budget

In this section, we provide estimates for total recharge in the basin, assuming steady-state "pre-development" conditions. This estimate, which is associated with a fairly high degree of uncertainty, can nonetheless provide a useful benchmark for comparisons with other water budget elements such as groundwater withdrawals.

Based on two different methods, our estimates of the total groundwater discharge to rivers in the basin range from 3.5 to 3.9 m^3/s. Assuming a steady-state, "pre-development" groundwater system, aquifer recharge would be approximately equal to discharge. Using basin boundaries as defined in *Figure* 3 and the precipitation/elevation relationship shown in *Figure 2,* we calculate average annual precipitation to be 89.1 m^3/s. Therefore, we estimate total aquifer recharge to be approximately 3.9 to 4.3 % of total precipitation. This estimate includes mountain-front recharge, stream-focused recharge, and areal recharge. This rate varies spatially, with the highest rates occurring at high elevations and canyon bottoms, the lowest rates occurring on mesatops and low elevations.

It is interesting to compare these estimates with groundwater flow modeling studies. *Table* **5** presents a comparison of total aquifer recharge assumed by Frenzel (1995), McAda and Wasiolek (1988), and Hearne (1985). All three of these models assume that most recharge occurs outside the modeled area (Jemez and Sangre de Cristo mountains) and is applied to the model through lateral (east and west) boundaries. None of these models extend significantly north of Espanola. By summing our baseflow estimates for reaches within their model boundaries (reaches 1,2,4,7,8, and 9) we arrive at an estimate of 1.9 m^3/s, close to the estimates of Frenzel (1995) and McAda and Wasiolek (1988); nearly twice that of Hearne (1985).

Recent groundwater withdrawals in the Los Alamos and Santa Fe well fields have totaled approximately 0.43 m^3/s (see *Figure 9).* Total groundwater withdrawal in the basin, including other municipal water supply systems, pueblo water use, and irrigation withdrawals is substantially larger (0.68 m^3/s). The total withdrawals are significantly less than our estimates of total inflow to the aquifer, however, the withdrawals from the Los Alamos and Santa Fe areas are larger than estimated baseflow to the Rio Grande below Otowi (0.39 m^3/s). Therefore, it is likely that baseflow to this reach is being diminished by groundwater exploitation.

Summary

Groundwater flow in the Española Basin is generally topographically driven, although locally geologic features such as faults and complex stratigraphy may produce departures from this pattern. Precipitation in the basin is strongly elevation dependent and so aquifer recharge is presumably greatest at the highest elevations. Water balance studies in several sub-basins along the margins of the basin suggest that recharge at these upper elevations ranges from *7* to *26%* of total precipitation. Groundwater discharges to the Rio Grande and the lower reaches of many of its tributaries; hydraulic gradients are upward in these areas causing artesian conditions in many wells near the river.

We use long-term streamflow records to estimate baseflow to various reaches within the basin. These analyses assume that increases or decreases in streamflow along a reach, measured in January when surface runoff is minimal, are caused by aquifer discharge or recharge at the river. Our annual baseflow estimates exhibited considerable interannual variability; however, statistical analyses demonstrated that most of this variation is unrelated to other hydrologic variables such as precipitation (past or present) or calculated gains/losses at other reaches and hence we assume that most of the apparent variability is due to measurement error. This is a reasonable conclusion given that the magnitude of calculated gains/losses is frequently very small compared to total flow in the reaches. Therefore, we conclude for most reaches that streamflow data are insufficient for estimating temporal trends and should instead be used as indicators of long-term averages

Presumably before significant groundwater withdrawals began to occur the aquifer was at a quasi-steady-state, with total discharge approximately equal to total recharge. Therefore, total aquifer discharge can be used as one way to estimate total aquifer recharge. Using a derived January flow- drainage area relation, we have estimated the total recharge to the aquifer (long-term average rate), To compare these estimates with earlier studies of the southern portion of the basin, we scale our estimates by area and find that our predictions are comparable to those of Frenzel (1995) and McAda and Wasiolek (1988) but significantly higher than those suggested by Hearne (1985). We conclude that total recharge to the regional aquifer is much higher than documented groundwater withdrawals for municipal supplies (Los Alamos County and City of Santa Fe). However, the impact of continued pumping on baseflow to rivers may be significant, particularly along the lower reaches of the Rio Grande.

Table 1. Water budget studies

	Watershed area (km^2)	Area-weighted average elevation[a] (m)	Precipitation[b] (m^3/s)	ET"	Runoff	Recharge	Sublimation	(cm/yr)	Source
					fraction of total precipitation				
Santa Fe River	74.3	2741	1.46	0.69	0.19	0.11	0.01	6.9	Wasiolek
Santa Fe River	69.8	2741	1.19	0.71	0.19	0.1	---	5.1	Anderholm (1994, table 7)[d]
Little Tesuque Creek	19.9	2679	0.37	0.72	0.09	0.19	0	11.2	Wasiolek
Little Tesuque Creek	19.0	2679	0.34	0.88	0.05	0.07	---	4.1	Anderholm (994, table 7)[d]
Rio Nambe	88.5	2843	1.94	0.62	0.19	0.11	0.08	7.7	Wasiolek
Tesuque Creek	29.0	2804	0.61	0.65	0.2	0.1	0.05	6.2	Wasiolek
Tesuque Creek	30.7	2804	0.52	0.75	0.18	0.07	---	3.8	Anderholm (994, table 7)[d]
Combined Little Tesuque and Tesuque Creeks	49.7	2753	0.86	0.89	0.04	0.07		3.9	Anderhom (1994, table 7)
Rio en Medio	22.5	2818	0.47	0.64	0.15	0.14	0.07	9.5	Wasiolek
Arroyo Hondo	21.7		0.34	0.84	0.06	0.1	---	4.7	Anderholm (1994, table 7)[d]
LA canyon '93	26.2	2570	0.54	0.71	0.03	0.26	0	16.6	Gray
LA canyon '94	26.2	2570	0.49	0.83	0	0.17	0	10.2	Gray
LA canyon '95	26.2	2570	0.63	0.73	0.02	0.25	0	18.6	Gray
LA canyon	26.2	2570				0.10			This study"

[a] Average basin elevation was calculated by the authors for drainage basins as defined by Wasiolek (1995).No adjustments were made to the average drainage basin elevations to account for the small differences in drainage basin areas between the Anderholm (1994) and Wasiolek (1995) studies.

[b] Precipitation volumes reported by Wasiolek (1995) for winter and spring had already been adjusted to reflect the effects of sublimation of snow. The precipitation volume estimated to exist before before sublimation was determined using information provided by Wasiolek (1995, p. 18). Calculated fractions for evapotranspiration, runoff, recharge, and sublimation reported here for the Wasiolek study are relative to this pre-sublimation precipitation value.

[c] The authors estimated the evapotranspiration (ET) for Anderholm's study based on Anderholm's estimates for precipitation (P), runoff (RO), and recharge (R): $ET = P - RO - R$

[d] These estimates used the chloride-based recharge estimate corrected for runoff from the basin.

[e] Based on chloride mass balance method

Table 2. Stream flow at locations in the vicinity of the Española Basin.

Stream Gage	USGS StationID	DrainageArea (km^2)	Period of record	Number of years	Mean January flow (m^3/s)	Predicted flow
Embudo Creek at Dixon	08279000	789.6	1923-1998	68	0.83	0.43
Jemez River below East Fork	08321500	447.9	1958-1990	29	0.42	0.24
Little Tesuque C near Santa Fe NM	08304100	1.7	1962-1969	8	0.00	0.00
Los Alamos Canyon near Los Alamos	08313042	23.5	1991-1991	1	0.01	0.01
Middle Fork Tesuque C near Santa Fe NM	08302300	1.1	1961-1969	9	0.00	0.00
North Fork Tesuque C near Santa Fe NM	08302200	4.1	1962-1969	8	0.01	0.00
Pueblo Canyon near Los Alamos	08313060	18.0	1991-1994	4	0.03	0.01
Redando Creek at Jemez Springs	08319945	0.0	1981-1984	4	0.02	0.00
Rio Chama Bl Abiquiu Dam	08287000	5558.3	1961-1998	38	4.75	3.02
Rio Chama near Chamita	08290000	8139.4	1912-1998	80	3.99	4.43
Rio De Los Frijoles in Bandelier Ntl. Mon.	08313350	45.3	1982-1995	14	0.04	0.02
Rio En Medio near Santa Fe NM	08295200	1.6	1964-1973	10	0.01	0.00
Rio Grande above San Juan Pueblo	08281100	27312.7	1962-1986	25	14.08	14.85
Rio Grande at Cochiti	08314500	37797.7	1925-1970	46	18.43	20.55
Rio Grande at Embudo	08279500	26924.4	1888-1998	104	14.26	14.64
Rio Grande at Otowi Bridge	08313000	37021.0	1894-1998	99	19.52	20.13
Rio Grande below Cochiti Dam	08317400	38574.3	1970-1998	29	24.06	20.97
Rio Nambe at Nambe Falls	08294300	65.0	1962-1978	17	0.10	0.04
Rio Nambe below Nambe Falls Dam	08294210	88.3	1978-1998	21	0.06	0.05
Rio Ojo Caliente at La Madera	08289000	1084.7	1932-1998	67	0.53	0.59
Santa Clara Creek near Espanola	08292000	90.6	1936-1993	16	0.09	0.05
Santa Cruz River near Cundiyo	08291000	222.6	1932-1998	67	0.27	0.12
Santa Cruz River, Riverside N NM	08291500		1941-1950	10	0.17	0.00
Santa Fe River above Cochiti Lake	08317200	598.0	1969-1998	30	0.29	0.33
Santa Fe River near Santa Fe NM	08316000	47.1	1912-1998	87	0.07	0.03
Tesuque C near Santa Fe NM	08302400	1.2	1962-1969	8	0.00	0.00
Predicted flow corresponding to the area of the Española Basin		2483.4				3.5

Table 3. Estimated groundwater gain/loss to eight reaches within the Española Basin, in m^3/s.

	Reach	Mean	Standard Error	Median	Count	Definition	Years
Rio Grande (Otowi to Cochiti)	1	0.37	0.12	0.39	44	8314500 - (8313000+8313350[1])	1926-1969
Rio Grande (Embudo to Otowi)	2	0.83	0.11	0.67	74	8313000-(8290000 +8292000[1] +8291500[1]+8279500)- Reach 9	1912,1913, 1918,1921, 1922,1927, 1929-1997
Chama River (Abiquiu to Chamita)	3	1.09	0.09	1.00	37	8290000-8287000	1961-1997
Santa Fe River	4	0.29	0.01	0.27	29	8317200	1970-1998
Jemez River	5	0.42	0.03	0.40	27	8321500	1958-1989
Rio Embudo[2]	6	0.58	0.02	0.58	68	8279500	1923-1998
Santa Cruz	7	0.17	0.04	0.14	10	8291500	1941-1950
Santa Clara	8	0.09	0.01	0.09	17	8292000	1936-1940,1949 1983-1993
Pojoaque River	9			0.40		estimate by Rieland and Koopman, (1975).	
Total for basin				3.90			

[1] for these relatively small tributaries, average January flow was used in baseflow gain calculation for those years when flow data was not available.
[2] Baseflow to Rio Embudo within the basin (drainage area 217mi^2) was estimated using January flow measured at gage (drainage area of 315 mi^2), scaled by area ratio.

Table 4. Comparison of gain/loss calculations to other studies,

Reach	Source	Method	reach length (km)	total gain (m^3/s)	gain/mi (m3/s/km)
Rio Grande – (Otowi to Cochiti)	this report (Table 3)	streamflow analysis	41.8	0.39	0.009
	U.S. Department of Justice(1996)	streamflow analysis	41.8	0.40	0.009
	Hearne (1985)	numerical model	38.6	0.46	0.012
	McAda and Wasiolek (1988)	numerical model	27.4	0.63	0.023
	Purtymun (1966)	seepage runs	18.5	0.43	0.023
	Griggs (1964)	seepage runs	18.5	0.37	0.020
	Spiegel & Baldwin (1963)'	streamflow analysis	32.2	0.71	0.022
Rio Grande (Embudo to Otowi)	this report (Table 3)	streamflow analysis	51.0	0.67	0.013
	U.S. Department of Justice(1996)	streamflow analysis	51.0	1.50	0.029
Rio Grande (Espanola to Otowi)	Hearne (1985)	numerical model	33.8	0.20	0.006
	McAda and Wasiolek (1988)	numerical model	17.7	0.43	0.024
	USGS seepage runs	seepage runs	27.4	0.15	0.005
Rio Chama (Abiquiu to Chamita)	this report (Table 3)	streamflow analysis	39.9	1.00	0.025
	this report (Table 2)	area relation	39.9	1.19	0.030
	Hearne (1985)(lower reach only)	numerical model		0.06	
Santa Cruz River	Hearne (1985)	numerical model		0.04	
	this report (Table 2)	area relation		0.33	
	this report (Table 3)	streamflow analysis		0.14	
Pojoaque River	Reiland & Koopman (1975)	streamflow analysis		0.40	
	McAda and Wasiolek (1988)	numerical model		0.21	
	this report (Table 2)	area relation		0.27	
	Hearne (1985)	numerical model		0.09	
Santa Clara River	this report (Table 3)	streamflow analysis		0.09	
	this report (Table 2)	area relation		0.05	
Santa Fe River	this report (Table 3)	streamflow analysis		0.23	
	Spiegel and Baldwin (1963)	streamflow analysis		0.18	
	McAda and Wasiolek (1988)	numerical model		0.18	
	this report (Table 2)	area relation		0.32	
	Hearne (1985)	numerical model		0.12	
Rio Embudo	this report (Table 3)	streamflow analysis		0.58	
	this report (Table 2)	area relation		0.30	

'their approach includes only those years when the reach was deemed "gaining" ("losing))years were assumed to be caused by erroneous data and were deleted from the analysis)

Table 5. Comparison of estimates of total inflow to aquifer (predevelopment), south of Española, from several studies.

Study	Total inflow (m^3/s)
Hearne (1985)	1.0
McAda and Wasiolek (1988)	2.1
Frenzel (1995)	1.6
this study	1.9

References

Adams, A.I., Goff, F. and Counce, D. (1995) Chemical and isotopic variations of precipitation in the Los Alamos region, New Mexico. Los Alamos National Laboratory. LA-12895-MS

Aldrich, M.J., Jr and Dethier, D.P. (1990) Stratigraphic and Tectonic Evolution of the Northern Espanola Basin, Rio Grande Rift, New Mexico. GSA Bulletin 102(12): 1695-1705..

Anderholm, S.K. (1994) Ground-water recharge near Santa Fe, north-central New Mexico. U.S.Geological Survey Water Resources Investigations Report,94-4078.

Blake, W.D., Goff, F., Adams, A.I. and Counce, D. (1995) Environmental geochemistry for surface and subsurface waters in the Pajarito Plateau and outlying areas, New Mexico. Los Alamos National Laboratory Report LA-12912-MS.

Bredehoeft, J. (1997) Safe yield and the water budget myth. Ground Water 35(6): 929-929

Brian, C. and Wilson, P.E. (1997) Water Use by Categories in New Mexico Counties and River Basins, and Irrigated Acreage in 1995, New Mexico State Engineer Office Technical Report

Coon, L.M. and Kelly, T.E. (1984) Regional hydrogeology and the effect of structural control and the flow of ground water in the Rio Grande trough, Northern New Mexico, NM Geological Society Guidebook, 35th Field Conference, Rio Grande Rift: Northern NM., pp. 241-244.

Daniel B. Stephens & Associates, I. (1994) Santa Fe County water resource inventory. Daniel B. Stephens and Associates Inc. .

Duffy, C.J. and Al-Hassan, S. (1988) Groundwater circulation in a closed desert basin: topographic scaling and climatic forcing. Water Resources Research 24(10): 1675-1688.

Duke Engineering & Services (2000) Water supply study: Jemez y Sangre water planning region, New Mexico. Jemez y Sangre Water Planning Council, Sangre de Cristo Water Company .

Freeze, R.A. and Cherry, J.A. (1979) Groundwater. Prentice-Hall, Inc, Englewood Cliffs, New Jersey, 604 pp.

Freeze, R.A. and Witherspoon, P.A. (1967) Theoretical analysis of regional groundwater flow, 2, Effect of water-table configuration and subsurface permeability variation, Water Resources Research 3: 623-634.

Frenzel, P.F. (1995) Geohydrology and simulation of ground-water flow near Los Alamos, North Central New Mexico.U.S. Geological Survey Water Resources Investigations Report,.95-4091.

Galusha, T. and Blick, J.C. (1971) Stratigraphy of the Santa Fe Group, New Mexico. Bull. Amer. Mus. Nat. Hist. 144(1).

Goff, F. and Sayer, S. (1980) A geothermal investigation of spring and well waters of the Los Alamos region, New Mexico.LA-8326-MS.

Golombek, M.P., McGill, G.E. and Brown, L. (1983) Tectonic and geologic evolution of the Espanola Basin, Rio Grande Rift: structure, rate of extension, and relation to the state of stress in the western United States. Tectonophysics 94: 483-507.

Gray, R.N. (1997) Hydrologic budget analysis and numerical simulations of groundwater flow in Los Alamos Canyon near Los Alamos, New Mexico. M.S. Thesis, University of New Mexico, Albuquerque, NM, 202 pp.

Griggs, R.L. and Hem, J.D. (1964) Geology and groundwater resources of the Los Alamos Area, New Mexico. U.S. Geological Survey Water-Supply Paper, 1753.

Halford, K.J. and Mayer, G.C. (2000) Problems associated wtih estimating ground water discharge and recharge from steam-discharge records. Ground Water 38(3): 331-342

Hearne, G.A. (1980) Simulation of an aquifer test on the Tesuque Pueblo grant, New Mexico. U S. Geological Survey Open File Report,80-1022.

Hearne, G.A. (1985) Mathematical model of the Tesuque aquifer system underlying Pojoaque River basin and vicinity, New Mexico. U.S. Geological Survey Water-Supply Paper,2205.

Kelley, V.C. (1978) Geology of the Espanola Basin, New Mexico Bureau of Mines and Mineral Resources Map.

Kernodle, J.M., McAda, D.P. and Thorn, C.R. (1995) Simulation of ground-water flow in the Albuquerque basin, Central New Mexico, 1901-1994, with projections to 2020.USGS Water-Resources Investigation Report,.94-4251.

LANL (1997) Environmental surveillance and Compliance at Los Alamos during 1996. Los Alamos National Laboratory.

McAda, D.P. and Wasiolek, M. (1988) Simulation of the regional geohydrology of the Tesuque aquifer system near Santa Fe, New Mexico. U.S. Geological Survey Water Resources Investigations Report, 87-4056, 71pp.

Newman, B.D. (1996) Vadose zone water movement at area G, Los Alamos National Laboratory, TA-54, interpretations based on chloride and stable isotope profiles. Los Alamos National Laboratory Report . LAUR-96-4682.

Newman, B.D., Gilkeson, R.H. and Gallaher, B.M. (1997) Vadose zone water movement at TA-49, Los Alamos National Laboratory: Interpretations based on chloride and stable isotope profiles. Los Alamos National Laboratory Report, LAUR-97-3924.

Person, M., Raffensperger, J.P., Ge, S. and Garven, G. (1996) Basin-scale hydrogeologic modeling. Reviews of Geophysics 34(1): 61-87.

Purtymun, W.D. (1966) Geology and hydrology of White Rock Canyon from Otowi to the confluence of Frijoles Canyon, Los Alamos and Santa Fe Counties, New Mexico. U.S. Geological Survey Open-File Report.

Purtymun, W.D. (1984) Hydrologic Characteristics of the Main Aquifer in the Los Alamos Area: Development of Ground Water Supplies, Los Alamos National Laboratory Report,.LA-9957-MS.

Purtymun, W.D. (1995) Geologic and hydrologic records of observation wells, test holes, test wells, supply wells, springs, and surface water stations in the Los Alamos area. Los Alamos National Laboratory Report,LA- 12883-MS.

Purtymun, W.D. and Johansen, S. (1974) General geohydrology of the Pajarito Plateau. New Mexico Geologic Society Guidebook, 25th Field Conference.

Purtymun, W.D., McLin, S. and Stoker, A. (1990) Water supply at Los Alamos during 1990. Los Alamos National Laboratory Report, LA- 12471-PR.

Purtymun, W.D., Stoker, A.K. and McLin, S.G. (1995) Water supply at Los Alamos during 1993. Los Alamos National Laboratory Report, LA- 12951-PR.

Reiland, L.J. and Koopman, F.C. (1975) Estimated availability of surface and ground water in the Pojoaque River drainage basin, Santa Fe County, New Mexico. U.S.G.S. Open-File Report 74-151.21 pages.

Reiter, M., Edwards, C.L., Hartman, H. and Weidman, C. (1975) Terrestrial heat flow along the Rio Grande Rift, New Mexico and Southern Colorado. Geological Society of America Bulletin 86: 811-818.

Rogers, D.B., Gallaher, B.M. and Vold, E.L. (1996a) Vadose zone infiltration beneath the Pajarito Plateau at Los Alamos National Laboratory. In: B.S.K. F. Goff, M.A. Rogers, L.D. McFadden, and J.N Gardner (Editor), New Mexico Geological Society Guidebook, 47th Field Conference, Jemez Mountain Region, pp. 413-420.

Rogers, D.B., Stoker, A.K., McLin, S.G. and Gallaher, B.M. (1996b) Recharge to the Pajarito Plateau regional aquifer system, NM Geol.Soc.Guidebook, 47th Field conference, Jemez Mountins Region.,.

Shomaker, J.W. (1974) Interpretation of well logs in Buckman well field, and general limits of Tesuque formation aquifer *(prepared for Black and Veatch Consulting Engineers)* .

Spiegel, Z. and Baldwin, B. (1963) Geology and water resources of the Santa Fe area. U.S. Geological Survey Water-Supply Paper,. 1525.

Stoker, A.K., McLin, S.G., Purtymun, W.D. and Maes, M.N. (1989) Water supply at Los Alamos during 1989. LANL .LA-12276-PR.

Toth, J. (1963) A theoretical analysis of groundwater flow in small drainage basins in central Alberta. J. Geophysical Research(68): 4795-4812.

Trainer, F.W. and Lyford, F.P. (1979) Geothermal hydrology in the Rio Grande Rift, north-central New Mexico, New Mexico Geologic Society Guidebook, 30th field Conference, Santa Fe Country.

U.S. Bureau of Indian Affairs, (1997) Unpublished data.

U.S. Department of Justice and New Mexico State Engineer Office (1996) Selection of a hydrogeologic model for water-rights administration in the Pojoaque River Basin Santa Fe County, New Mexico.

U.S. Geological Survey, (1997) WATSTORE database.

Vuataz, F.D., Stix, J., Goff, F. and Pearson, C.F. (1984) Low-temperature geothermal potential of the Ojo Caliente warm springs area, northern New Mexico. Los Alamos National Laboratory .LA-10105-OBES.

Wasiolek, M. (1995) Subsurface recharge to the Tesuque Aquifer system from selected drainage basins along the western side of the Sangre de Cristo mountains near Santa Fe, New Mexico. U.S. Geological Survey Water-Resources Investigations Report, 94-4072.

List of Figures

Figure 1. Geologic map of the Española Basin, modified from Aldrich (1990).

Figure 2. Precipitation data, Basin data is taken from Spiegel and Baldwin (1963); Pajarito Plateau data is taken from stations at Los Alamos National Laboratory, the Soil Conservation Service, and Bandelier National Monument.

Figure 3. Water levels in wells. Contour interval = 100m. Dots are well locations; polygon shows approximate location of basin boundary. Heads along cross-sections AA' are shown in Figures 4.

Figure 4. Water levels in wells along cross-section AA'. Only wells with relatively short screens are included in the figure.

Figure 5. Water budget components (expressed as a fraction of total precipitation) for studies summarized in Table 1

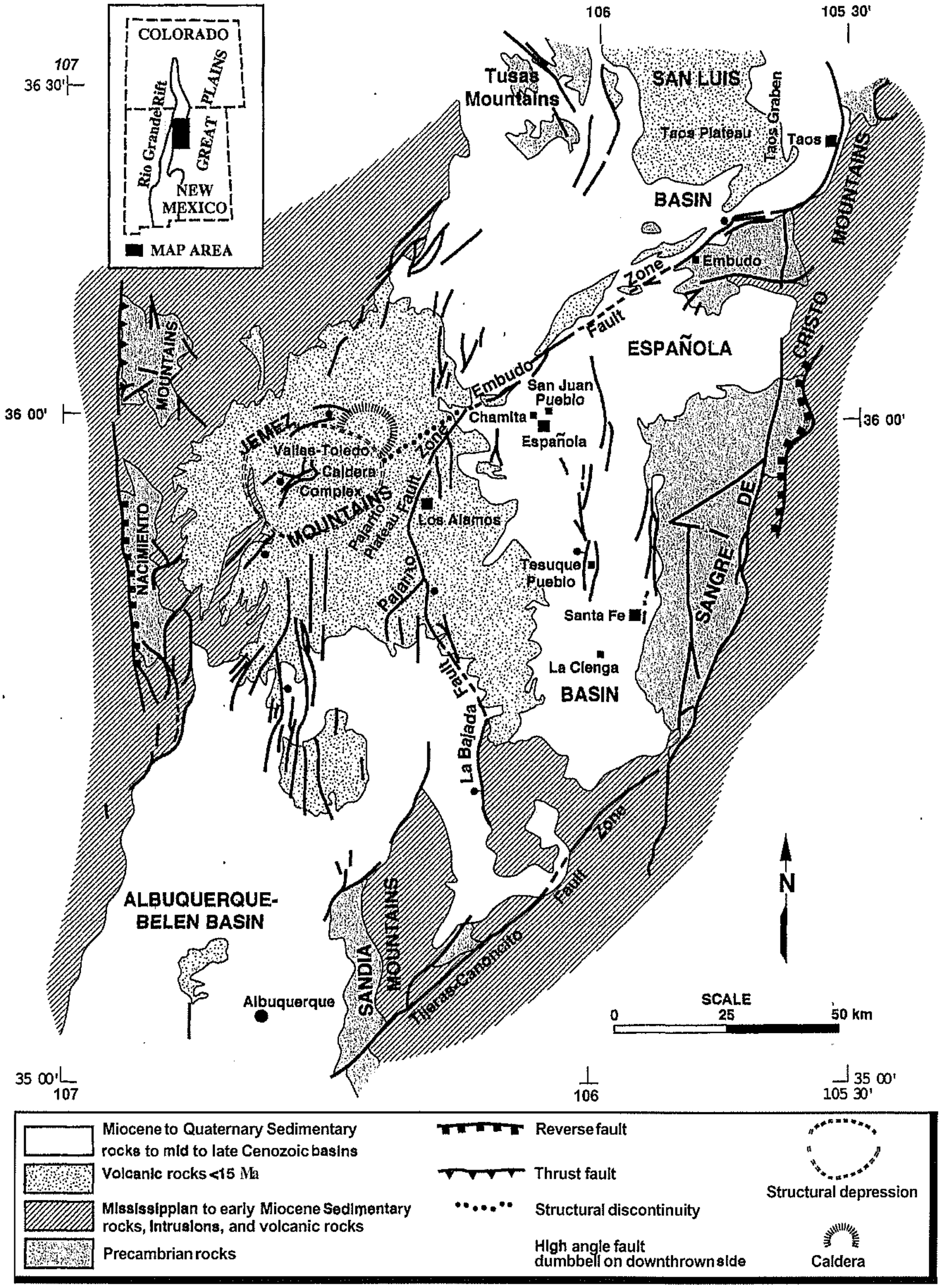
Figure 6. Location of stream gages in the vicinity of the Espanola Basin

Figure 7. Relation between drainage area and mean January flow

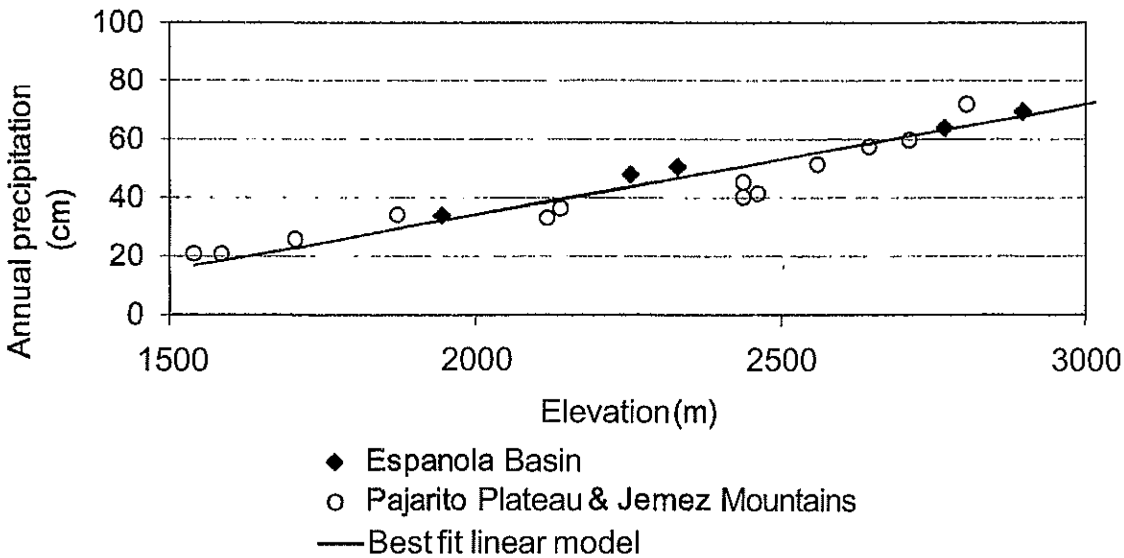
Figure 8. Yearly baseflow gains along various reaches of streams and rivers in the Espanola Basin

Figure 9. Withdrawals from municipal well fields in southern portion of the basin.

Figure 1. Generalized geologic map of the Espanola Basin, modified from Aldrich et al. (1990)



COLORADO
Rio Grande Rift
GREAT PLAINS
NEW MEXICO
MAP AREA
107
36 30'
106
105 30'
Tusas Mountains
SAN LUIS
Taos Plateau
Taos Graben
Taos
BASIN
Embudo
Embudo Fault Zone
ESPAÑOLA
SANGRE DE CRISTO MOUNTAINS
NACIMIENTO MOUNTAINS
36 00'
JEMEZ MOUNTAINS
Valles-Toledo Caldera Complex
Pajarito Plateau Fault Zone
San Juan Pueblo
Chamita
Española
Los Alamos
Tesuque Pueblo
Pajarito Fault
Santa Fe
La Cienga
BASIN
La Bajada Fault
ALBUQUERQUE-BELEN BASIN
SANDIA MOUNTAINS
Tijeras-Canoncito Fault Zone
Albuquerque
N
SCALE
0
25
50 km
35 00'
107
106
105 30'
Miocene to Quaternary Sedimentary rocks to mid to late Cenozoic basins
Volcanic rocks <15 Ma
Mississippian to early Miocene Sedimentary rocks, Intrusions, and volcanic rocks
Precambrian rocks
Reverse fault
Thrust fault
Structural discontinuity
High angle fault dumbbell on downthrown side
Structural depression
Caldera



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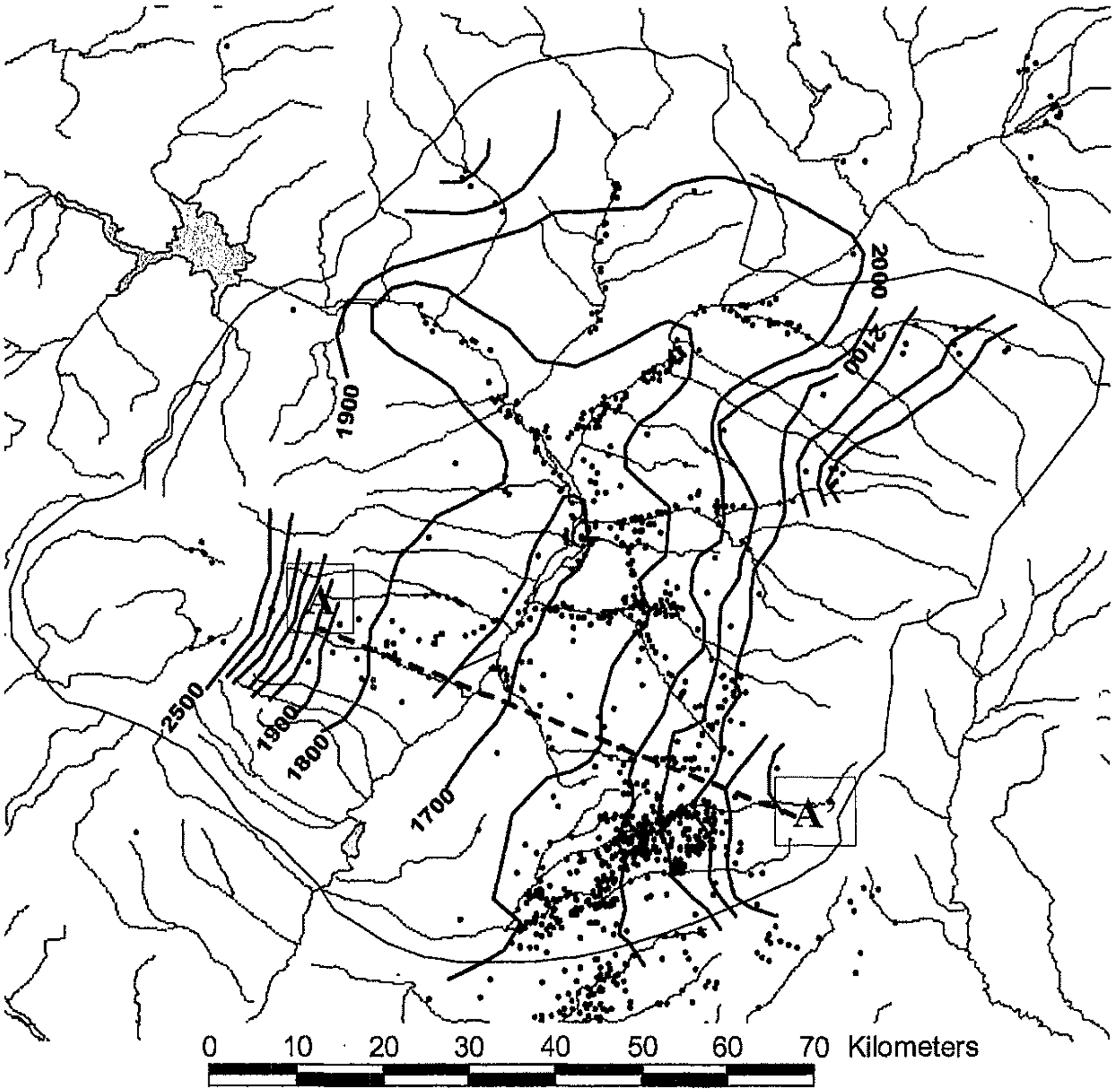

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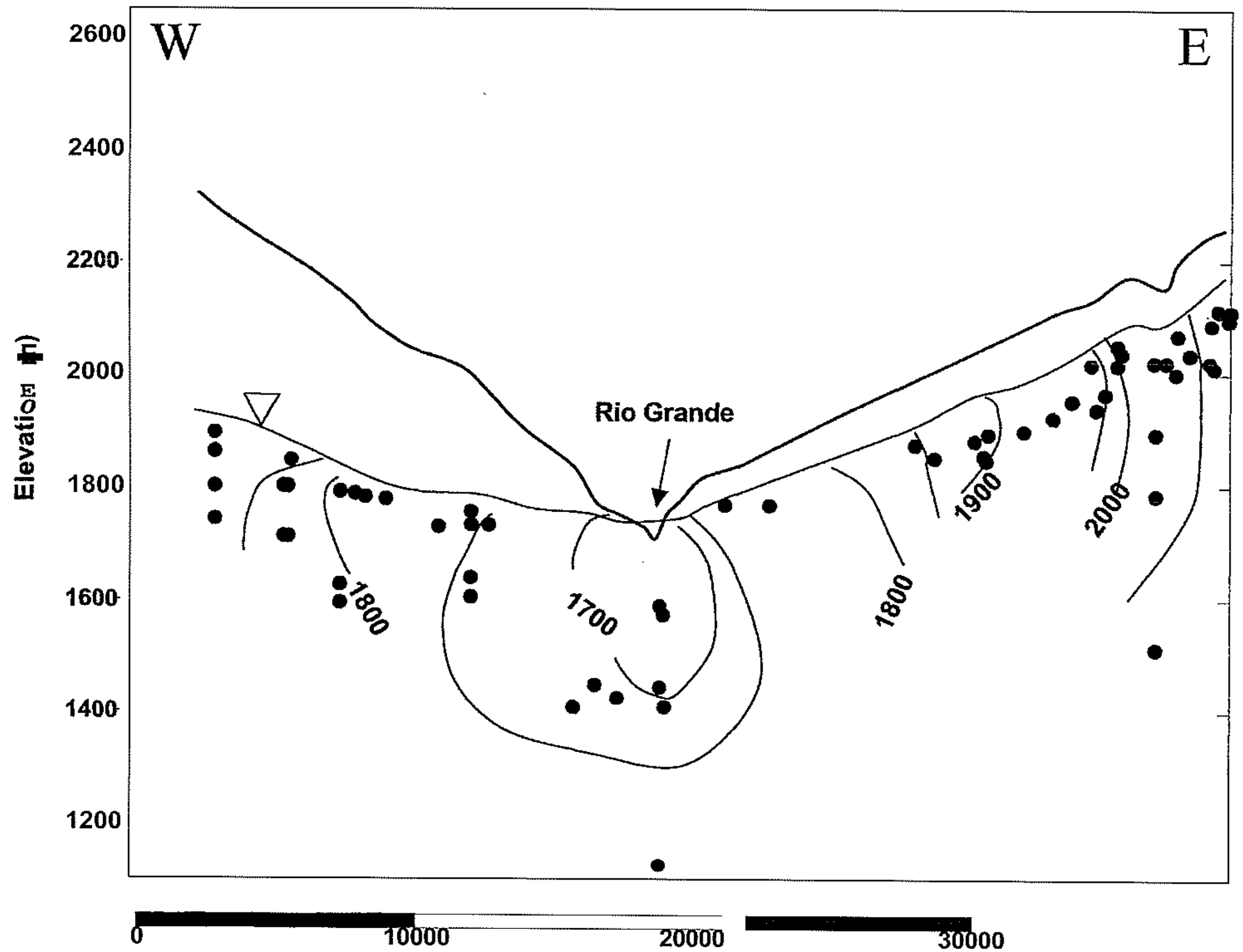
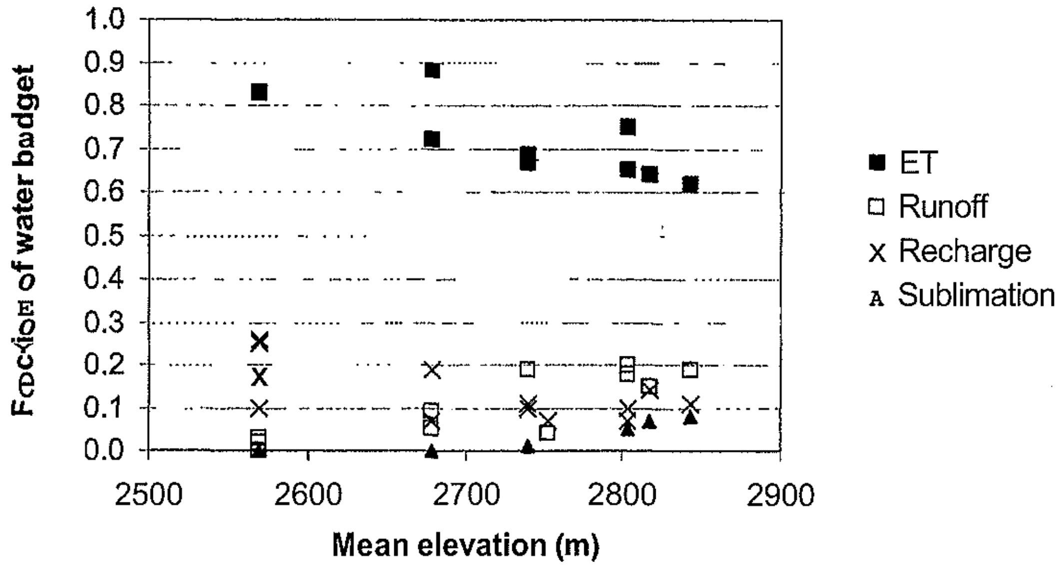


Figure 4. Water levels in wells along cross-section AA'. Approximate ground surface, piezometric surface, and head contours (50m interval) are shown. Vertical exaggeration= 20.



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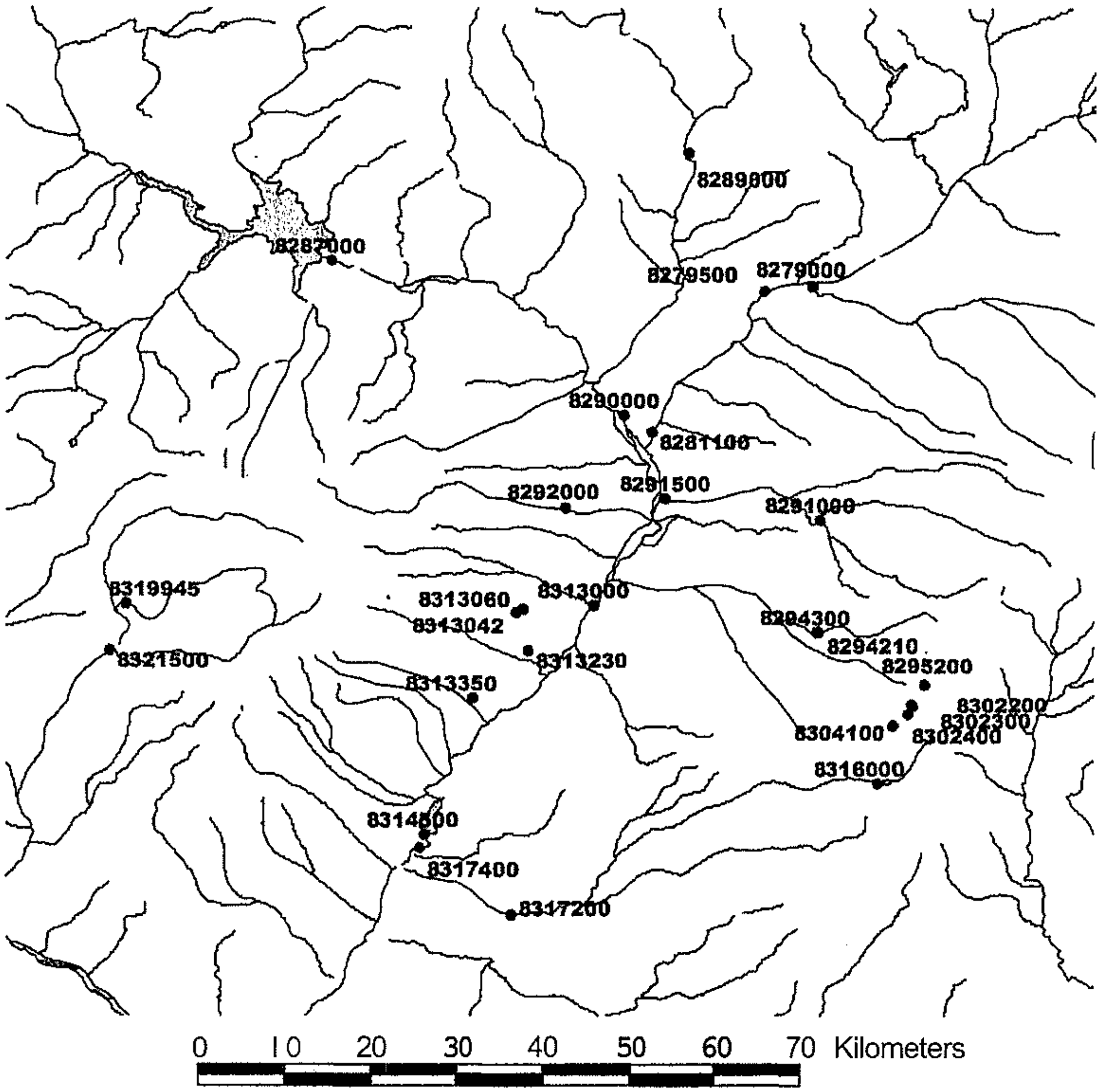

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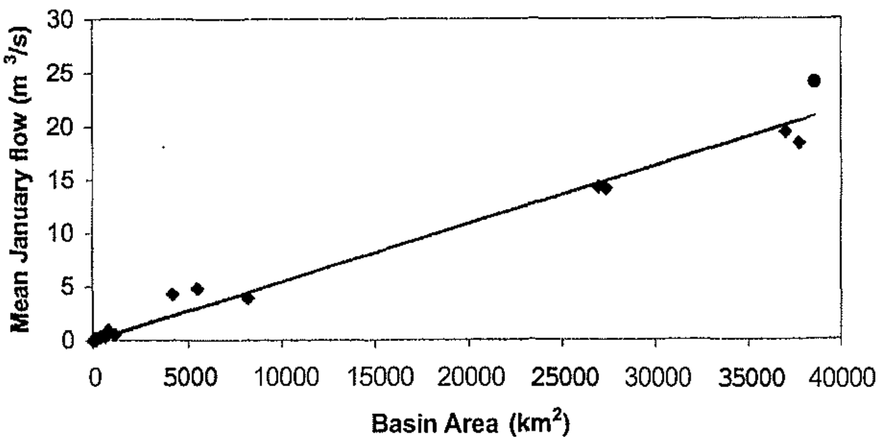

Figure 7. Relation between drainage area and mean January flow

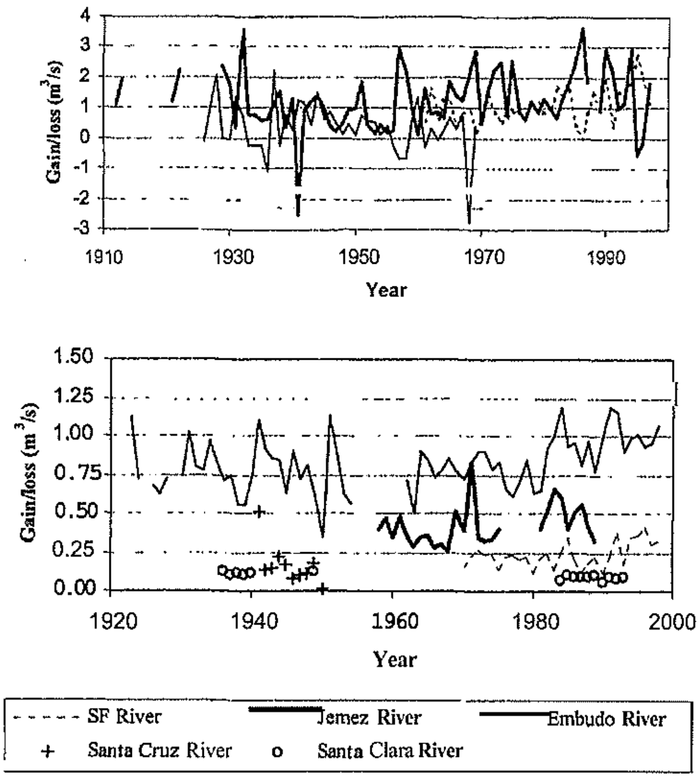

Figure 8. Yearly baseflow gains along various reaches of streams and rivers in the Espanola Basin

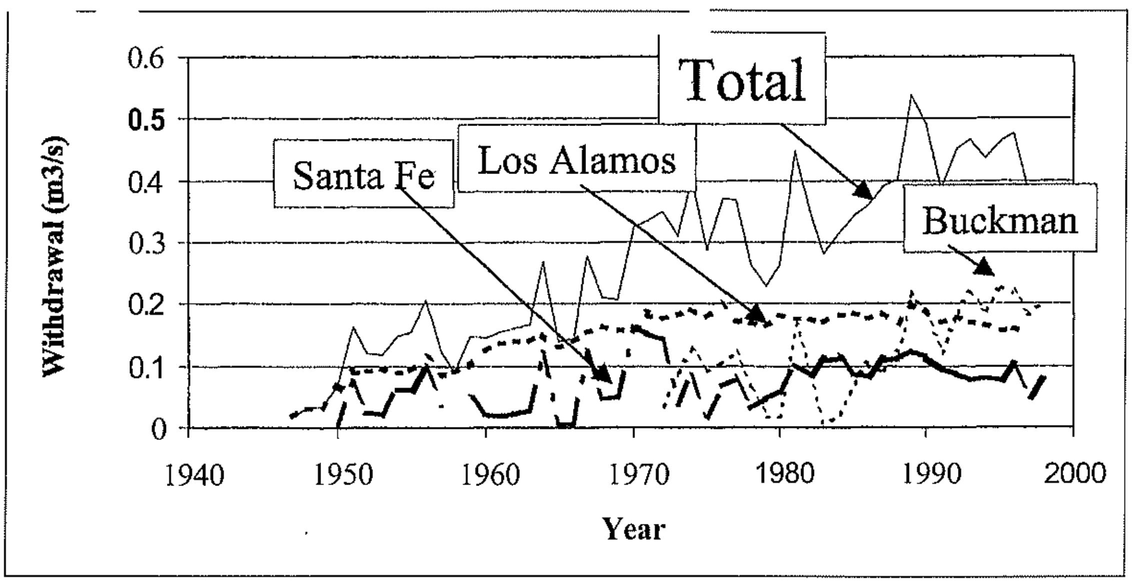

Figure 9. Annual production from municipal well fields