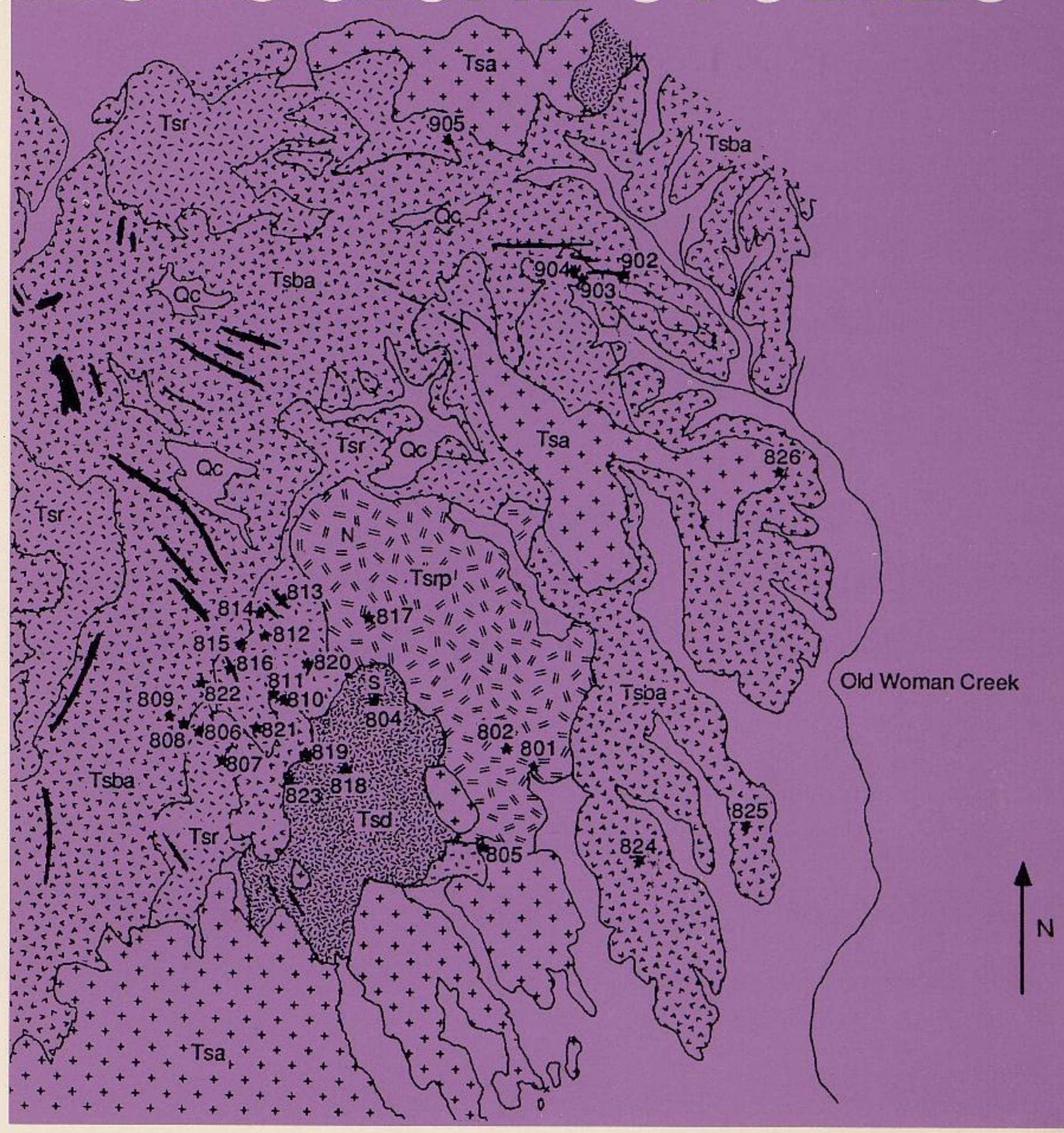
BAYLOR GEOLOGICAL STUDIES



FALL 1994

Bulletin No. 54

Thesis Abstracts

"Creative thinking is more important than elaborate equipment--"

FRANK CARNEY, PH.D.
PROFESSOR OF GEOLOGY
BAYLOR UNIVERSITY
1929-1934

Objectives of Geological Training at Baylor

The training of geologists in a university covers but a few years; their educations continue throughout their active lives. The purposes of training geologists at Baylor University are to provide a sound basis of understanding and to foster a truly geological point of view, both of which are essential for continued professional growth. The staff considers geology to be unique among sciences since it is primarily a field science. All geologic research including that done in laboratories must be firmly supported by field observations. The student is encouraged to develop an inquiring objective attitude and to examine critically all geological concepts and principles. The development of a mature and professional attitude toward geology and geological research is a principal concern of the department.

Cover: Geologic map and sample localities of Twin Mountains (after Lipman, 1976, Geologic Map of the Del Norte area, eastern San Juan Mountains, Colorado). (From Willman.)

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THESIS ABSTRACTS

These abstracts are taken from theses written in partial fulfillment of degree requirements at Baylor University. The original, unpublished versions of the theses, complete with appendices and bibliographies, can be found in the Jesse Jones Library, Baylor University, Waco, Texas.

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Department of Geology
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Baylor Geological Studies

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Runoff, Soil Water, and Groundwater Along a First Order Stream in the Austin Chalk

Norman L. Bingham

The shallow groundwater system in the Austin Chalk provides substantial baseflow to area streams. Along the Austin Chalk outcrop water flowing through the shallow groundwater system discharges at the surface allowing area streams to continue to provide valuable water to reservoirs long after the seasonal rains have ceased. Central Texas residents rely heavily on surface water for their basic needs. However, there is little understanding of the soil and groundwater flow within Austin Chalk watersheds. This study provides information that will help minimize impacts to the hydrologic system caused by urban development, non-point source pollutants, and toxic chemical spills along the Austin Chalk outcrop and similar environments.

The study area is located in McLennan County approximately nine miles south of Waco in Central Texas. It is two miles due west of the intersection of Farm Road 2113 and the Missouri Kansas Texas Railroad at a longitude of 97°14′ and a latitude of 31°26′. The site includes the headwaters of a first order ephemeral stream on the south branch of Bull Hide Creek. Regional topography is typified by gently rolling hills with slopes between one and eight percent. Physiographically the site is part of the White Rock Prairie, a subprovince of the larger Black Prairie and lithologically it is underlain by the Austin Chalk.

The monitoring of runoff plots and the analysis of rain simulation data indicate that infiltration occurs through the soil matrix during low-intensity storms and through macropores during more intense storms. The capacity of the soil matrix to infiltrate water is exceeded when 1.1 inches of rain falls at an intensity of 0.85 inches/hour causing runoff to occur.

The response of tensiometers in the lower soil profile before those in the upper profile indicates that water is flowing along macropores bypassing the upper profile and concentrating at the base of the soil horizon. Water then moves laterally along the soil rock interface (Fig. 1). Average linear pore velocities as high as 105 feet/hour were recorded in this zone. Depending on the storage available in the aquifer, water is either stored in the fractured rock or in the soil. When pre-storm aquifer levels are high, the additional water saturates the Fairlie Series Soil and begins to seep out at the surface. The conductivity of this seepage water is over 500 μS while that of the runoff is less than 200 μS.

The maximum depth of the active groundwater zone, where the majority of the groundwater flow takes place, varies between 6 and 12 feet across the basin (Fig. 2). This zone has its greatest depth in the valley region.

The flow of water within the shallow groundwater system can be divided into three flow periods over the monitoring year:

1. Groundwater levels 10 feet below the surface in the divides and valleys. Under this condition, the first major storm causes a rise only in valley wells and no long-term flow in the channel occurs. During this period, water only fills the lower portion of the active aquifer zone.
2. Valley aquifer levels higher than five feet below ground surface. A major storm at this time causes a rise in all wells basin-wide and continuous flow for several days in the channel. During this period, water cannot be drained completely by the subsurface soil and rock system and therefore flow begins on the surface. In addition, renewed groundwater flow begins in the divide regions.
3. Aquifer levels within six feet of the surface in both valley and divide. Aquifer properties are fairly uniform above six feet, thus shallow divide and valley wells respond in a similar fashion. Wells under these conditions all rose at least three feet and extended channel flow occurred during this period.

When a threshold rainfall volume and intensity are exceeded, quick flow is generated. If valley aquifer levels are high (five feet below the surface) additional water infiltrating and flowing laterally along the soil rock interface begins to seep out through the Fairlie Soil (Fig. 1). This baseflow continues for three to five days. The groundwater system in the valley region has the greatest hydraulic conductivity and responds to infiltrating water before the divide system; 6.9% of yearly precipitation is groundwater flow (Fig. 2). The majority of this passes through the soil and rock in the valley region. Nine percent of yearly precipitation runs off as quick flow and 84.1% was lost to evapotranspiration.

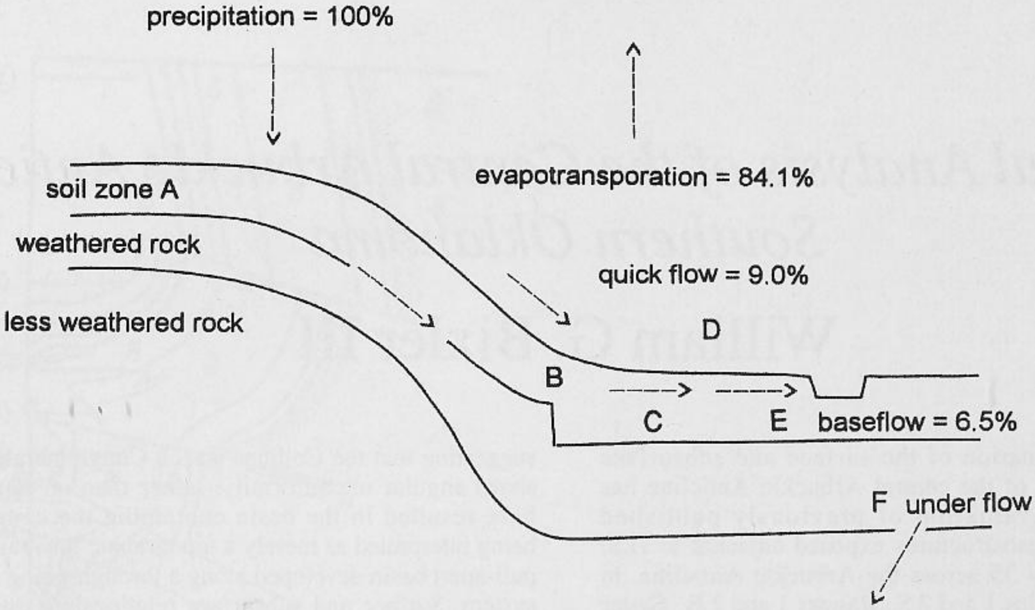

Fig. 1. Annual water budget and flow paths. Infiltrating water flows along the soil rock interface (A) at rates as high as 105 ft/hr (B). This water recharges the deeper Fairlie Series Soil (C) and seeps out at the surface as baseflow. Nine percent of precipitation runs off as quick flow (D), while 84.1% is lost as evapotranspiration. Baseflow (E) and under flow (F) combine to make 6.9% of yearly precipitation.

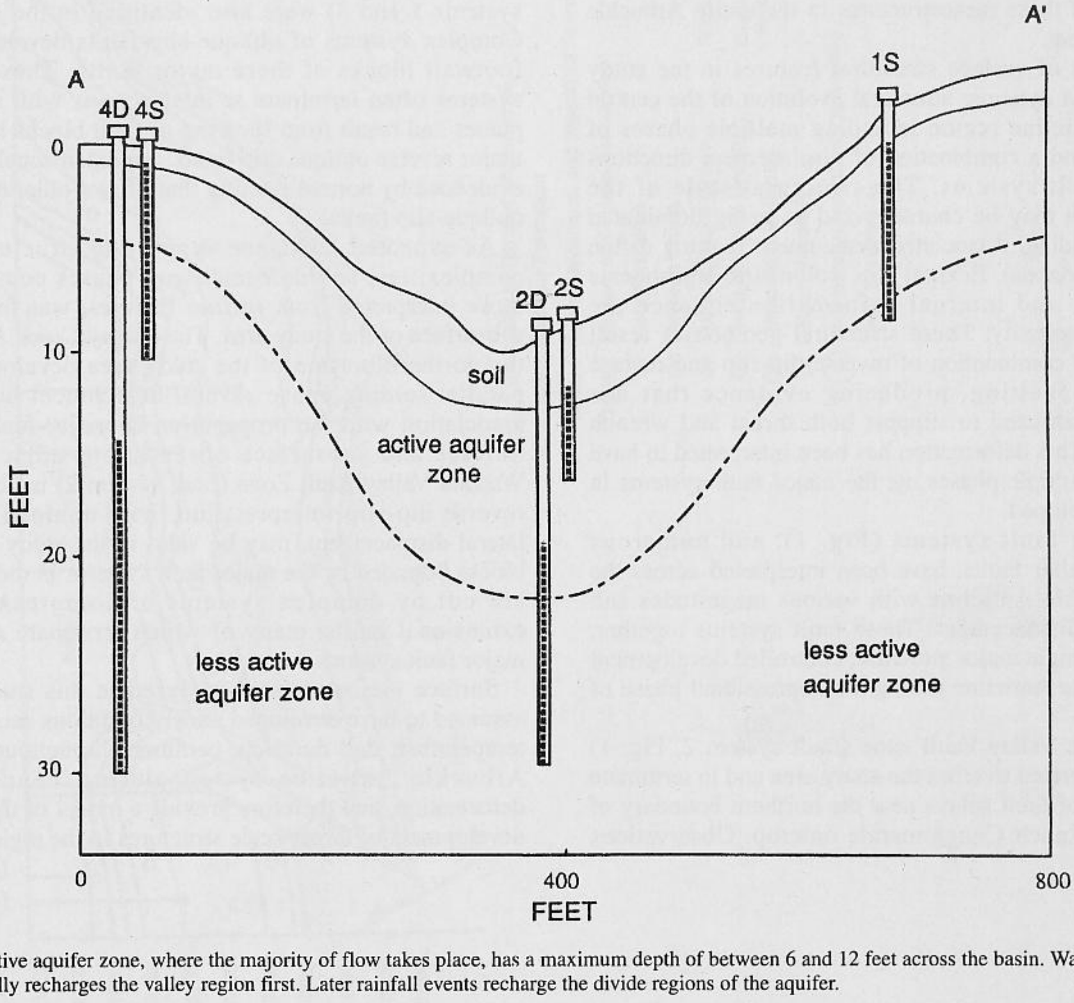

Fig. 2. The active aquifer zone, where the majority of flow takes place, has a maximum depth of between 6 and 12 feet across the basin. Water infiltrating and flowing laterally recharges the valley region first. Later rainfall events recharge the divide regions of the aquifer.

Structural Analysis of the Central Arbuckle Anticline, Southern Oklahoma

William G. Bixler III

Detailed examination of the surface and subsurface structural geology of the central Arbuckle Anticline has resulted in a reevaluation of previously published interpretations. Mesostructures exposed adjacent to U.S. Interstate Highway 35 across the Arbuckle Anticline, in portions of Townships 1 and 2 S., Ranges 1 and 2 E., Carter and Murray Counties, southern Oklahoma, have been catalogued, photographed, and interpreted, in order to define the structural style and to understand the development of deformational features in the area. Additional local and regional subsurface observations have been incorporated into the study in order to demonstrate the relationship of these mesostructures to the entire Arbuckle Anticline system.

A synthesis of surface structural features in the study area suggests a dynamic structural evolution of the central Arbuckle Anticline region including multiple phases of deformation and a combination of displacement directions on major fault systems. The structural style of the mesostructures may be characterized as being dominated by parallel folding. Associated detachment faulting (often at multiple horizons), flexural slip, volumetric adjustments within folds, and internal deformation enhance the structural complexity. These structural geometries result mainly from a combination of reverse dip-slip and reverse oblique-slip faulting, producing evidence that has previously been used to support both thrust and wrench fault models. This deformation has been interpreted to have occurred in multiple phases, as the major fault systems in the region developed.

Five major fault systems (Fig. 1), and numerous associated smaller faults, have been interpreted across the central Arbuckle Anticline with various magnitudes and directions of displacement. These fault systems together, rather than a single major structure, controlled development of the Arbuckle Anticline during a compressional phase of deformation.

The Washita Valley Fault zone (fault system 2, Fig. 1) has been interpreted to cross the study area and to terminate into a system of fault splays near the northern boundary of the Collings Ranch Conglomerate outcrop. Observations suggesting that the Collings Ranch Conglomerate lies on a sharp angular unconformity, rather than on fault planes, have resulted in the basin containing the conglomerate being interpreted as merely a topographic low, rather than a pull-apart basin developed along a through-going strike-slip system. Surface and subsurface relationships suggest that the Washita Valley Fault is a reverse dip-slip fault, with minimal lateral offset on the central Arbuckle Anticline.

Two additional major, basement-involved, fault systems (fault system 4 and the Chapman Ranch Thrust) appear at the surface to the south of the Collings Ranch Conglomerate outcrop. Two major blind fault systems (fault systems 1 and 3) were also identified in the study area. Complex systems of oblique-slip faults developed in the footwall blocks of these major faults. These complex systems often terminate at intersections with major fault planes and result from shearing of fault blocks bounded by major reverse oblique-slip faults. An extensional phase was evidenced by normal faulting that offsets older dip-slip and oblique-slip faults.

As expected, evidence suggesting structural styles, complexities, and deformational phases comparable to those interpreted from surface features, was found in the subsurface of the study area. The Honey Creek Anticline in the northern portion of the study area developed due to parallel folding above several detachment horizons, in association with the propagation of major fault systems. Surface and subsurface observations adjacent to the Washita Valley Fault Zone (fault system 2) indicated that a reverse dip-slip interpretation, with minimal necessary lateral displacement, may be valid in the study area. Fault blocks bounded by the major fault systems in the study area are cut by complex systems of compressional and extensional faults, many of which terminate against the major fault systems.

Surface mesostructures analyzed in this study may be assumed to have developed under conditions such as stress, temperature, and duration, common throughout the entire Arbuckle Anticline system during Pennsylvanian deformation, and therefore provide a model of the style and development of larger-scale structures in the region.

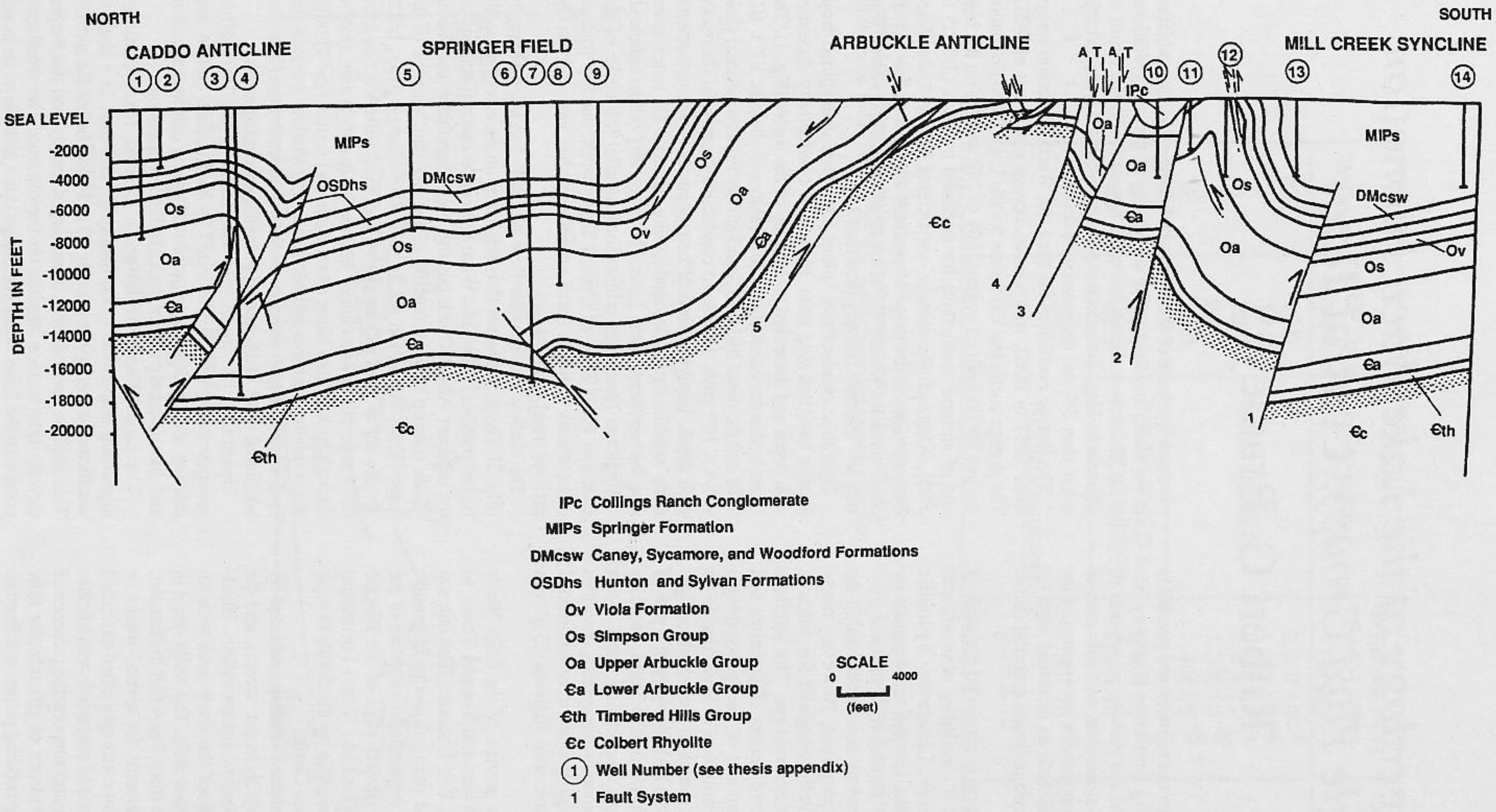

Fig. 1. Regional cross section drawn from Mill Creek Syncline, south to the Caddo Anticline. Fault systems are numbered 1 through 5 in the proposed order of development. Fault system 2 corresponds to the Washita Valley Fault zone. Fault system 5 corresponds to the Chapman Ranch Thrust.

The Hydrogeology of the Lake Waco Formation (Eagle Ford Group), Central Texas

Robert G. Bradley

The Lake Waco Formation crops out adjacent and parallel to the urban growth corridor along Interstate Highway 35. Approximately twenty percent of the present population of Texas is concentrated along this corridor, and this area is destined to be one of the most populated urban regions in the state. The effects of urbanization include an increased need for waste disposal sites and an increased pollution potential due to spills and leaks.

Because the Lake Waco Formation is composed primarily of shale, it is usually considered a confining unit without significant groundwater resources. Laboratory hydraulic conductivity (K) tests on the shale material meet regulatory requirements for a suitable waste disposal liner material ($K < 1.0 \times 10^{-7}$ cm/s). Landfills are presently constructed within the outcrop and others have been proposed. However, many shallow, hand-dug wells are located throughout the outcrop, indicating a shallow groundwater system. The shallow groundwater occurs in a weathered veneer of fractures and bedding plane separations. The amount of the hydrogeologic information on the Lake Waco Formation is small due to its nonaquifer status, resulting in a great need for hydrogeologic information. This study describes the hydrogeology of a part of the Lake Waco Formation outcrop in Central Texas, provides information about the hydrogeologic characteristics and in situ parameters of a weathered shale groundwater system, describes flow systems, and provides information to aid in environmental management decisions within the Interstate Highway 35 growth corridor.

The study area is located in a portion of the Lake Waco Formation outcrop belt between Waco and Temple, Texas, in southern McLennan and northern Bell Counties. The outcrop forms a thin band adjacent and parallel to the Interstate Highway 35 growth corridor. The westernmost exposure of the Lake Waco Formation forms the western edge of the Bosque Escarpment of Central Texas. The Lake Waco Formation overlies the Pepper Shale and is overlain by the South Bosque Shale which is overlain by the Austin Chalk.

Evidence of a shallow groundwater system is indicated by numerous hand-dug wells, baseflow to major streams, and the presence of phreatophytes along shallow stream valleys. Hand-dug wells are distributed throughout the study area with an overall density of .47 wells per square mile. The wells range in depth from 8.5 to 28.9 feet and are from 3 to 4 feet in diameter. Baseflow remains in the major streams for several weeks or months and is also found in minor streams after extended periods of precipitation. The presence of phreatophytes indicates abundant moisture near the surface and supports the presence of a shallow groundwater system. Evidence of groundwater flow also occurs as mineralization along bedding planes and fracture planes.

The effective porosity of the Lake Waco Formation is due to fractures and bedding plane separations primarily due to the overburden release of the overconsolidated shale, the oxidation of the shale, and the wetting and drying of the soil. Additional areas of interconnected fractures appear to be the result of adjacent faulting. The effective porosity estimated from outcrops is less than .5% and decreases with depth.

The storage coefficient for the Lake Waco Formation ranges from .0017 to .0063, with a mean storage coefficient of .0032. The storage coefficient (S) values for the Lake Waco Formation are low for unconfined systems (.01 to .30) and are on the high end of storage coefficients for confined systems (.00005 to .005). Although the storage coefficients for the Lake Waco Formation are in the range of confined systems, its low storage coefficients are actually the result of an unconfined system with very low effective porosity (less than 1%).

Hydraulic conductivity values also were determined from aquifer tests and slug tests. Hydraulic conductivity decreases with depth and decreased weathering of the shales (Fig. 1). The mean hydraulic conductivity for weathered shale is 1.7×10^{-4} cm/s and the mean hydraulic conductivity for unweathered shale is 1.4×10^{-7} cm/s. Hydraulic conductivity generally decreased with depth, implying vertical heterogeneity. The unweathered shale exhibits hydraulic conductivity values 1000 times lower than the weathered shale. Lateral heterogeneity is also indicated.

Regional lineament orientations parallel the strike of the Balcones Fault Zone of Central Texas while outcrop fracture orientations indicate local anisotropy that varies from the regional trend.

The shallow groundwater flow generally follows topography (Fig. 2). The groundwater flow originates on the topographically higher portions of the Lake Waco Formation outcrop belt and on the adjacent Austin Chalk outcrop belt. Regionally, the Austin Chalk outcrop and the outcrop of the Bouldin Member of the Lake Waco Formation act as regional recharge areas. Local divides act as local recharge areas and local valleys act as local discharge points. Recharge enters the groundwater system through bypass flow along macropores and matrix flow through the overlying Vertisols and Mollisols. The shallow groundwater is discharged into streams as baseflow and interflow and additional water is discharged through evapotranspiration.

Generally, the quality of the Lake Waco Formation groundwater system is dominated by bicarbonate, sulfate, and calcium ions which can be attributed to the calcareous matrix and disseminated pyrite of the Lake Waco Formation.

The Lake Waco Formation in Central Texas contains significant amounts of good quality groundwater in a shallow weathered veneer approximately 10 to 30 feet below the surface. This shallow groundwater is recharged locally and discharged directly into surface streams on the outcrop. The weathered portion of the Lake Waco Formation is not suitable as an in situ liner for landfills. The unweathered shale is marginally suitable as an in situ landfill liner but fractures associated with the Balcones Fault Zone may affect its integrity.

WELL #	DEPTH (feet)	DESCRIPTION	MEAN K FOR EACH WELL (cm/s)	MEAN K (cm/s)
2S	19.0	WEATHERED	2.8×10^{-4}	1.7×10^{-4}
5	14.2		1.9×10^{-4}	
6	26.7		3.5×10^{-4}	
7	25.0		1.3×10^{-5}	
9	25.0		1.4×10^{-5}	
3	43.5	WEATHERED & UNWEATHERED	2.3×10^{-4}	1.6×10^{-4}
4	43.5		2.9×10^{-3}	
2D	38.5	UNWEATHERED	1.3×10^{-8}	1.4×10^{-7}
11C	≈40		2.75×10^{-7}	

Fig. 1. Chart of in situ values of hydraulic conductivity. This chart shows the relationship of weathering to hydraulic conductivity. Generally, the more weathered the shale, the higher the hydrualic conductivity will be. Values for wells 2S, 3, 4, 5, and 6 were determined from two constant-rate aquifer tests and slug tests. Values for wells 2D, 7, 9, and 11C were determined by slug tests.

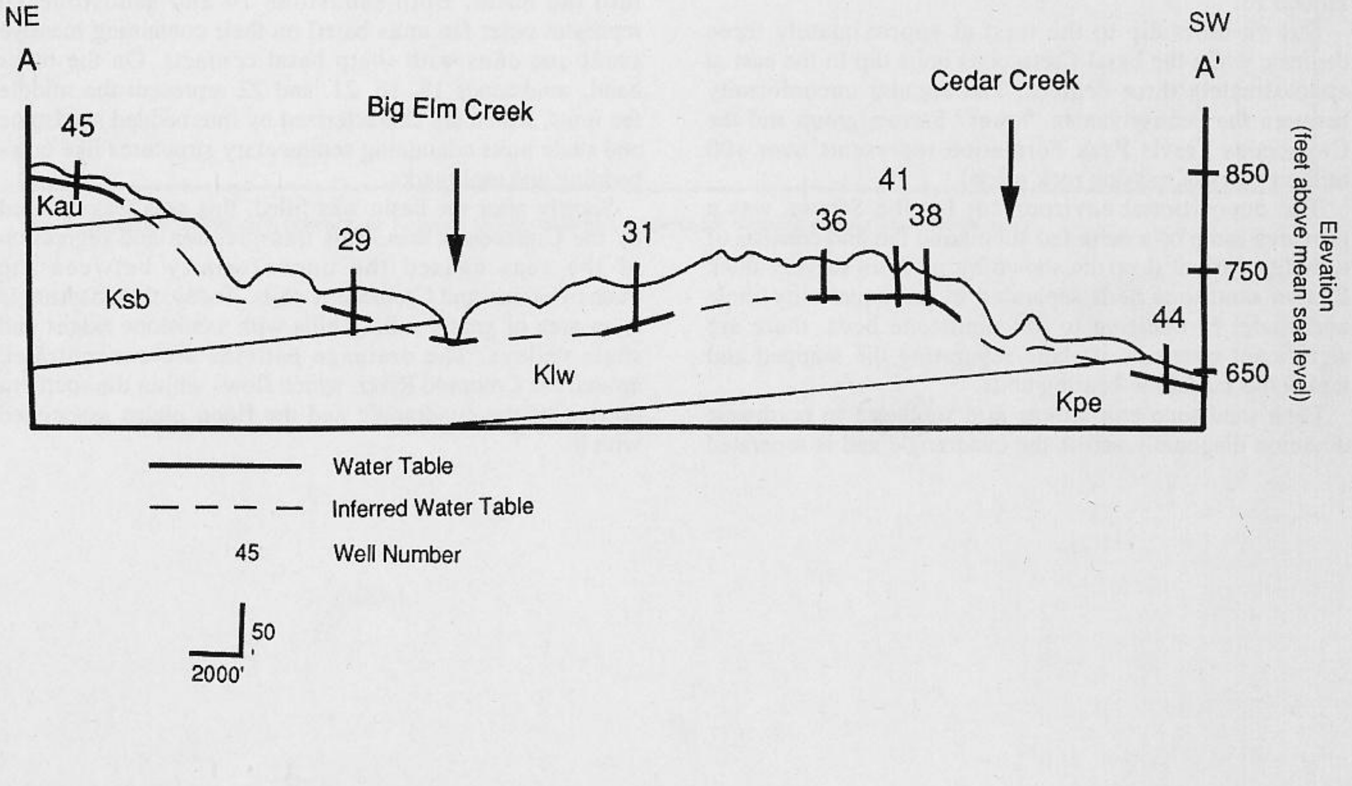

Fig. 2. Cross section showing a regional groundwater flow system in the Lake Waco Formation of Central Texas. In this cross section, the water table mimics the topography. This topographic expression is a result of the resistant nature of the Austin Chalk, which forms the White Rock Cuesta in the northeast, and the resistant nature of the limestone flags of the Lake Waco Formation.

Geology of the Regency Quadrangle, Central Texas

Paige H. Clark

The Regency quadrangle, a 7.5-minute quadrangle in the southern Fort Worth (foreland) basin, lies in the Strawn outcrop belt, 15 miles west of Goldthwaite, Texas, within Mills and San Saba Counties. The purpose of this study is to map the exposed rocks in the quadrangle and to provide detailed descriptions and information for broader regional studies. Exposed rocks include formations of the Pennsylvanian "lower" Strawn (sandstone units 14–22), the Cretaceous Travis Peak, and Quaternary alluvium along the Colorado River system.

Most of the Regency quadrangle is covered by gently rolling hills composed primarily of shale, capped with sandstone. Cretaceous deposits are located mostly in the north to northeastern portion of the area and consist of coarse-grained conglomerate. Paralleling the Colorado River are Quaternary alluvium deposits, which occupy flat floodplain areas.

Landscape evolution in the quadrangle was controlled by regional and local tectonism, mainly the Ouachita fold belt which created the Fort Worth basin. As the basin filled with lower Strawn shelf/slope sequences, shallowing basin conditions allowed for the progradation of fluvial/deltaic sequences.

Strawn units dip to the west at approximately three degrees, while the basal Cretaceous units dip to the east at approximately three degrees. The angular unconformity between the Pennsylvanian "lower" Strawn group and the Cretaceous Travis Peak Formation represents over 100 million years of missing rock record.

The depositional environment for the Strawn was a proximal ramp of a delta-fed submarine fan and consists of turbidity current deposits shown by medium to very thick Strawn sandstone beds separated by subequal silty-shale and shale. In addition to the sandstone beds, there are significant intervals of shale separating the mapped and numbered sandstone-bearing units.

Each sandstone unit strikes in a southeast to northwest direction diagonally across the quadrangle and is separated from the next higher sand interval by a shale interval that is generally underrepresented on the Brownwood sheet. Also on the Brownwood sheet, sandstone units are mapped as overlapping each other. In the limited field area, however, there are shale intervals between each sandstone unit. Another difference of treatment in this study compared to that on the Brownwood sheet is that sandstone 16 and 17 were combined and mapped as sandstone 16. There is no evidence of a shale interval between the two units, thus supporting the theory that these units should be treated as a single unit. In addition to this combination of units, Hanna Mountain is mapped as sandstone 20, but it is actually capped by a siltstone that is more characteristic of undifferentiated Strawn. Thus, this siltstone and thin sand interval is included in the undifferentiated Strawn that separates sandstone 16 and 20.

The Regency quadrangle depositional history of the lower Strawn units can be interpreted using the ramp fan model (Fig. 1). Standstone units were largely controlled by the prograding nature of the slope and by the filling and abandonment of fan lobes. As a lobe was abandoned, a new slope was created, further spreading the sediment fill out into the basin. Both sandstone 14 and sandstone 20 represent outer fan units based on their containing massive sandstone units with sharp basal contacts. On the other hand, sandstones 15, 16, 21, and 22 represent the middle fan units, which are characterized by interbedded sandstone and shale units containing sedimentary structures like cross bedding and tool marks.

Shortly after the basin was filled, this area was covered by the Cretaceous seas. This transgression and regression of the seas caused the unconformity between the Pennslyvanian and Cretaceous units. Today, the quadrangle is an area of gently rolling hills with sandstone ridges and shale valleys. The drainage patterns are concentrated around the Colorado River, which flows within the southern portion of the quadrangle and the flood plains associated with it.

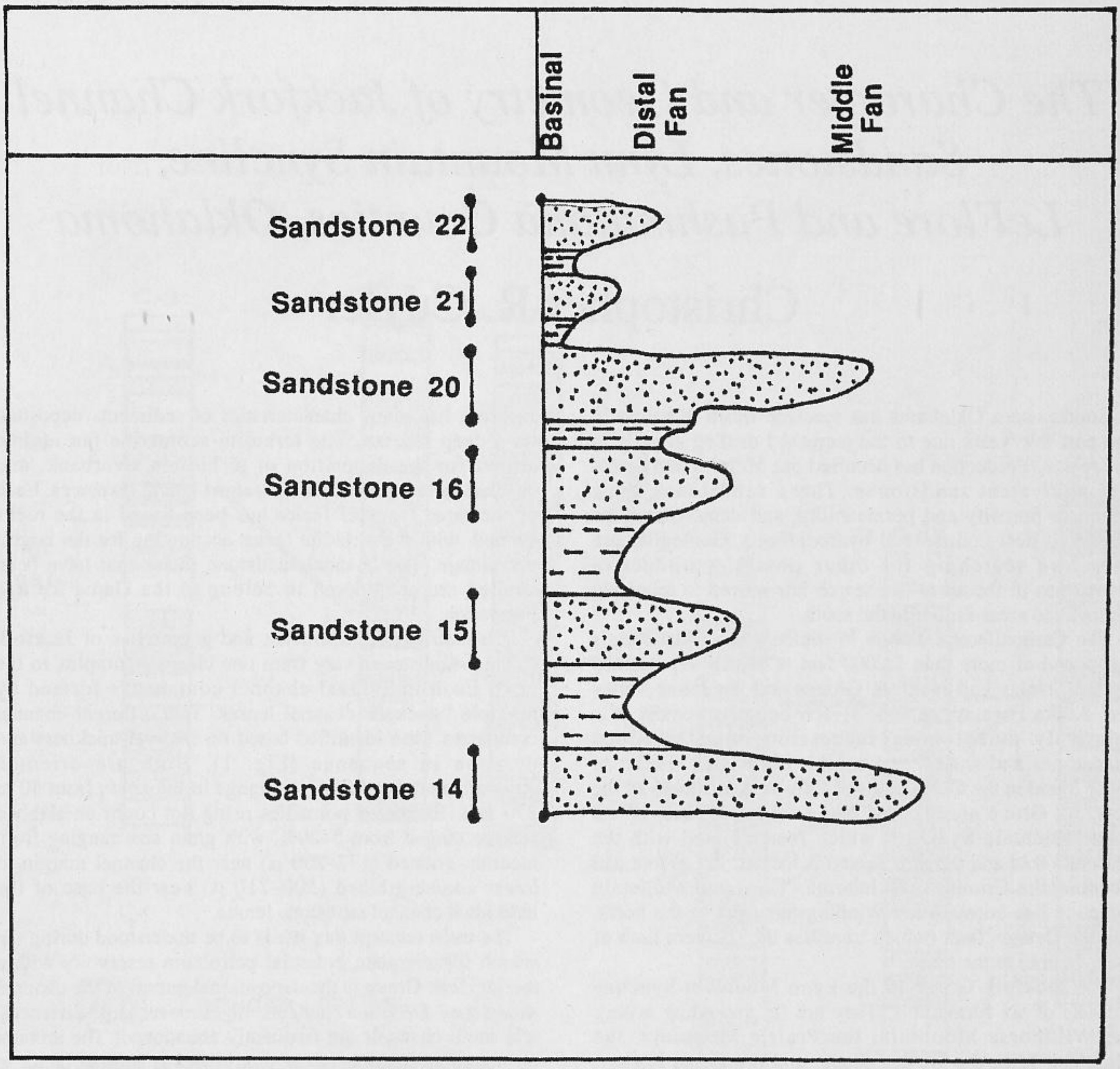

Fig. 1. Schematic correlation between the Regency Strawn units and the delta-fed submarine ramp depositional model.

The Character and Geometry of Jackfork Channel Sandstones, Lynn Mountain Syncline, LeFlore and Pushmataha Counties, Oklahoma

Christopher R. Cuyler

Southeastern Oklahoma has received much attention in the past few years due to the increased drilling activity in the region. Production has occurred out of Spiro and Spiro-age-equivalent sandstones. These sandstones show adequate porosity and permeability, and demonstrate the ability to host commercial hydrocarbons. Geologists are therefore searching for other possible productive formations in the area. The search has moved to relatively unexplored areas farther to the south.

The Carboniferous section in southeastern Oklahoma is composed of more than 15,000 feet of flysch, represented by the Stanley and Jackfork Groups and the Johns Valley and Atoka Formations. The flysch deposits consist of a relatively monotonous succession of interbedded sandstones and shales, and exhibit lithologies similar to those found in the Carpathians of Europe. Sandstones of the Jackfork Group are exposed along the north flank of the Lynn Mountain Syncline, which is associated with the Ouachita fold and thrust belt, and is located in LeFlore and Pushmataha Counties, Oklahoma. The Lynn Mountain Syncline lies between the Windingstair fault to the north, and the Octavia fault (which truncates the southern limb of the syncline) to the south.

The Jackfork Group in the Lynn Mountain Syncline consists of six formations. They are (in ascending order): the Wildhorse Mountain, the Prairie Mountain, the Markham Mill, the Wesley Shale, and the Game Refuge. The Jackfork Group lies above the Stanley Group, and below the Johns Valley Shale. The formations in the Jackfork Group contain significant amounts of sandstones, and the Jackfork is therefore the major ridge former in the area. The general trend is a decrease in shale and an increase in sandstone, as one moves younger stratigraphically. The sandstone range from fine- to coarse-grained sand and exhibit a fining upward character. The Jackfork has many characteristics of sediments deposited on a deep sea fan. The turbidite-submarine fan regime allows for the deposition of turbidites, overbank, and channel facies within relatively short lateral distances. Each of the three types of facies has been found in the rocks studied, with the turbidite facies accounting for the largest percentage. The channel sandstone facies that have been studied are considered to belong to the Game Refuge Formation.

The distribution, character, and goemetries of Jackfork channel sandstones vary from one channel complex to the next. Each individual channel complex is formed by multiple "stacked" channel lenses. Two different channel complexes were identified based on interval thickness and position in sequence (Fig. 1). Both are oriented approximately east-west, and range in thickness from 50 to 270 feet. Estimated porosities using dot count on slabbed surface ranged from 5–20%, with grain size ranging from medium-grained (177–250 μ) near the channel margin to lower coarse-grained (500–710 μ) near the base of the individual channel sandstone lenses.

The main concept that needs to be understood during the search for adequate potential petroleum reservoirs within the Jackfork Group is the depositional nature of the channel sandstones. Deep sea channels migrate over large distances, and small channels are frequently abandoned. The amount of interconnectedness of the sand lenses is another factor to be considered. In order for the Game Refuge channel sandstones to prove to be valuable to the petroleum industry, the lenses must be connected and fluids allowed to migrate throughout the channels. These channel sandstones, when found with the proper stratigraphy, structure, seal, source, maturation, and subsurface location, may prove to be excellent petroleum reservoirs.

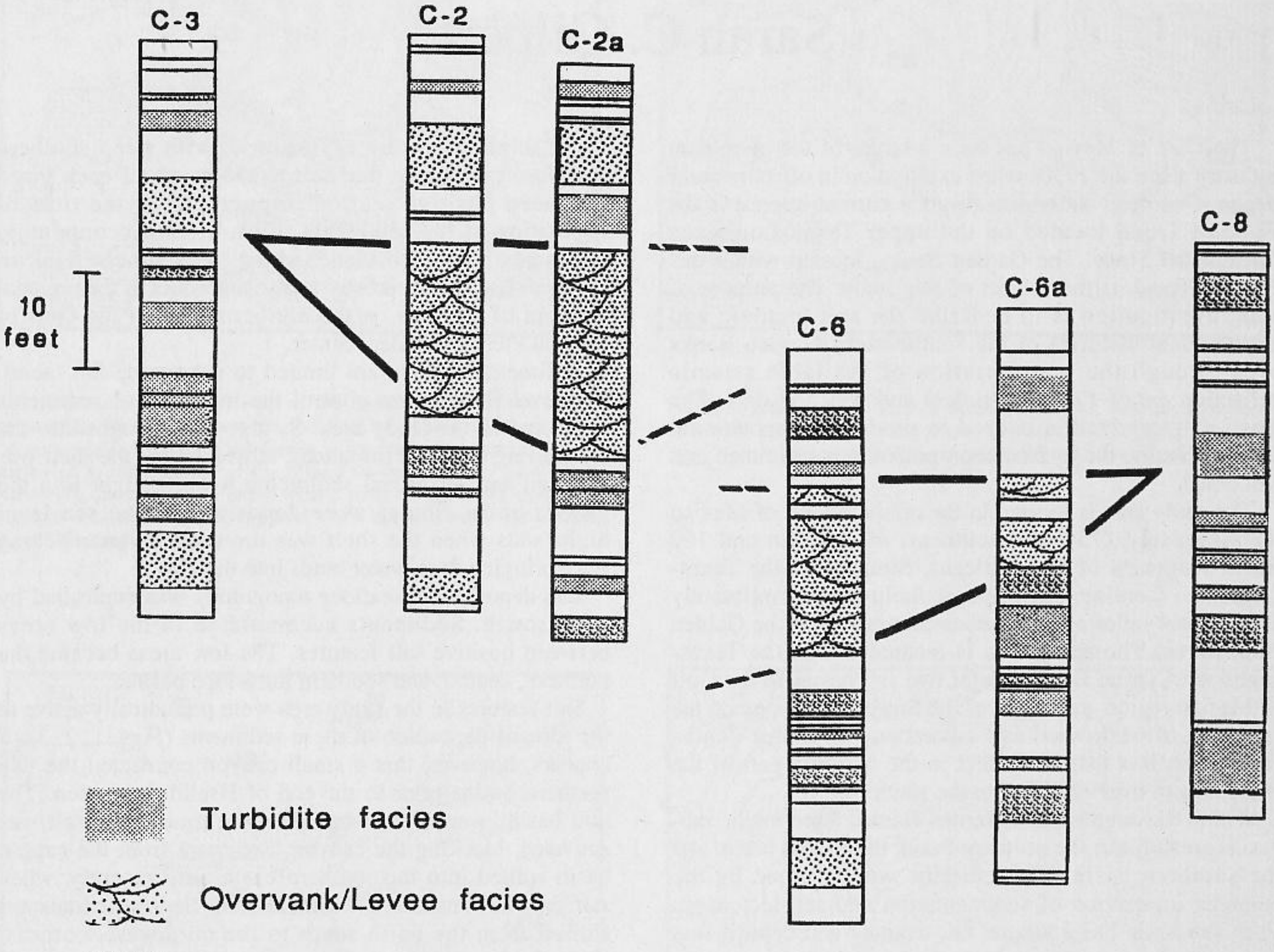

Fig. 1. Stratigraphic cross section showing the relationship and size of Jackfork channel sandstones within the surrounding submarine fan facies. Note the "stacking" of individual channel sandstone lenses within the channel complexes.

Salt Tectonism, Depositional History, and Hydrocarbon Potential of Southeastern Garden Banks, Offshore Gulf of Mexico

Sarah C. Gilbert

The Gulf of Mexico has been a target of the petroleum industry since the 1930s when exploration in offshore areas began. One deep-water area of great current interest is the Flexure Trend located on the upper Texas-Louisiana Continental Slope. The Garden Banks, located within this Flexure Trend, is the subject of this study. The purpose of this investigation is to describe the salt tectonic and depositional histories of the southeastern Garden Banks area through the interpretation of available seismic reflection data (~1700 line-miles) and well log data. The seismic interpretation is used to produce a depositional model. Finally, the hydrocarbon potential is examined and discussed.

The study area is located in the offshore Gulf of Mexico approximately 175 miles southeast of Houston and 160 miles southwest of New Orleans. Situated on the Texas-Louisiana Continental Slope, it includes approximately 1850 square miles of the southeastern corner of the Garden Banks area. The study area is located within the Texas-Louisiana Coastal Basin, one of five salt basins in the Gulf of Mexico region, and north of the Sigsbee Escarpment, the terminus of basinward salt advancement. Water depths range from less than 2000 feet in the northern part of the study area to over 7000 feet to the south.

Within the southeastern Garden Banks, three major sub-basins are defined: the northern basin, the central basin, and the southern basin. These basins were formed by the complex interaction of sedimentation and salt tectonism. Each sub-basin has a unique salt tectonic and depositional history. These histories are described with reference to two seismic horizons defined through seismic sequence analysis: (1) the Lenticulina (Lentic) horizon (~2.25 million years before present) and (2) the Helicosphaera Sellii (Hsellii) horizon (~1.27 million years before present). These horizons are based on paleontological abundances of marine microfossils and correspond to accepted global sequence boundaries.

Located north of the Sigsbee Escarpment, the area of study is characterized by isolated salt diapirs and narrow, linear salt walls (Figs. 1, 2). The area has been subject to high sedimentation rates. With increased sedimentation, the weight of overburden created instability in the system, which activated salt growth, both lateral and vertical.

All three basins are asymmetric, with steep southern margins, suggesting that salt to the south of each basin produced positive seafloor topography at the time of deposition of the sediments (Figs. 1, 2). Accumulating sediments cause subsidence along the southern flank of each sub-basin. Asymmetry is also a product of the regional location of the area on the northern edge of the Gulf of Mexico Pleistocene depocenter.

Sedimentary strata are limited to submarine fan facies. Sea level fluctuations control the lithology of sediments deposited in the study area. Sands were brought into the area during sea-level lowstands, at which time the shelf was exposed and terrestrial sediments were brought into the deeper basin. Shales were deposited during sea-level highstands when the shelf was drowned by encroaching seas, bringing deep water muds into the area.

The depositional seafloor topography was controlled by salt growth. Sediments accumulated in the low areas between positive salt features. The low areas became the northern, central, and southern intraslope basins.

Salt features in the study area were periodically active at the time of deposition of these sediments (Figs. 1, 2, 3). It appears, however, that a small canyon connected the two northern basins prior to the end of Hsellii deposition. The two basins were connected until the time when salt was activated, blocking the canyon. Sediment from the central basin spilled into the southern basin until recently, when salt growth constricted sedimentation. Sediment transport shifted from the north-south to the northwest-southeast (Figs. 3, 4).

The basins in the study area contain sediments known to produce hydrocarbons. Although the sediments are of Plio-Pleistocene age, which is considered young for hydrocarbon generation, they are located at considerable depth. At these depths, temperatures may have been high enough to develop mature hydrocarbons. Reservoirs are sea-level lowstand sands, and sources are the interbedded sea-level highstand shales. Salt acts as a seal to the hydrocarbons, and structural and stratigraphic traps exist in the area. Therefore, favorable conditions exist for hydrocarbon generation and accumulation within the area of study.

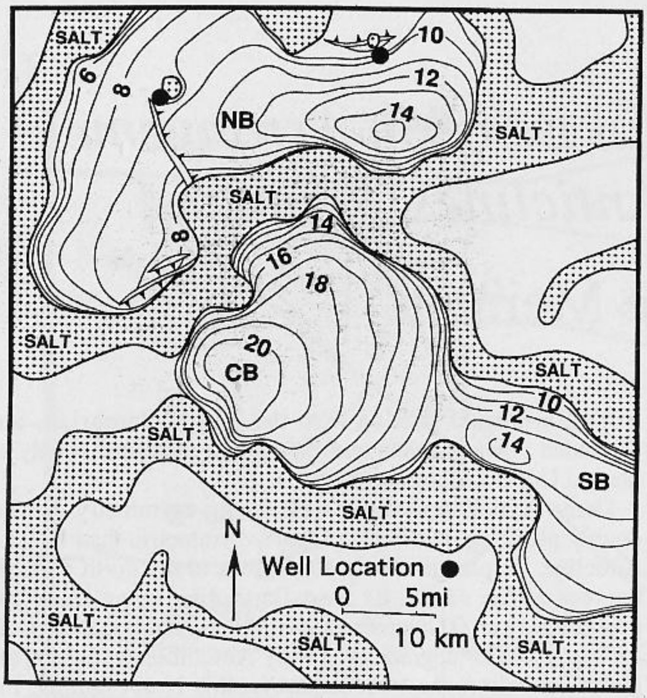

Fig. 1. Structural contour map of the Hsellii horizon. The area is characterized by narrow, linear salt walls and by isolated salt diapirs. Note the asymmetry of the three sub-basins to the south. This phenomenon is attributed to the location of the region on the northern flank of the Pleistocene depocenter and to the local activity of salt in the study area. Contours = x1000 feet below sea level. NB = northern basin, CB = central basin, SB = southern basin.

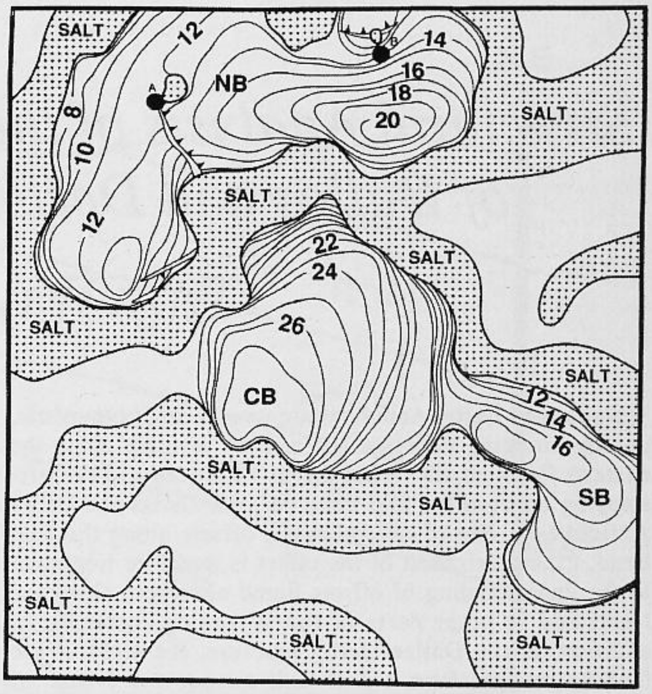

Fig. 2. Structural contour map of the Lentic horizon. Note the asymmetry of the three sub-basins to the south. Contours = x1000 feet below sea level. NB = northern basin, CB = central basin, SB = southern basin.

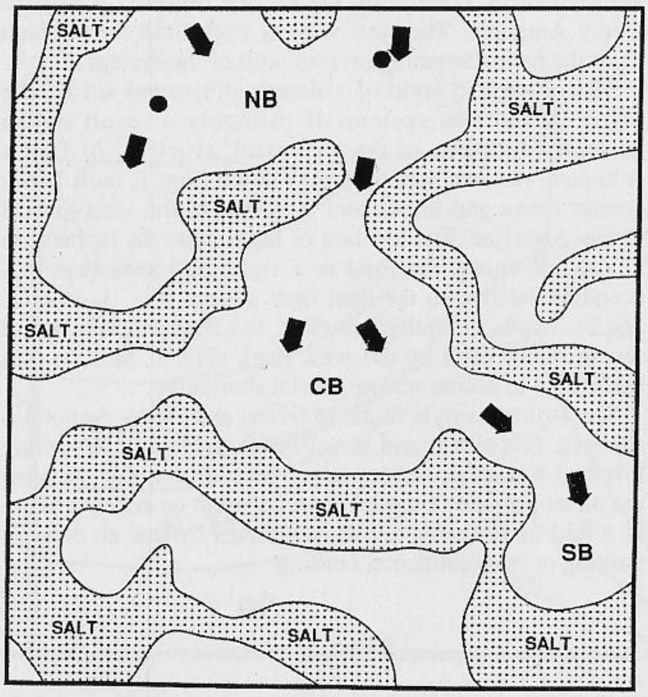

Fig. 3. Map of the study area showing the narrow canyon which connected the northern basin and central basin. This canyon was blocked by salt growth sometime during the Lentic-Hsellii interval. Note direction of sediment transport (arrows). Prior to canyon blockage, transport was mainly from the north to south. NB = northern basin, CB = central basin, SB = southern basin.

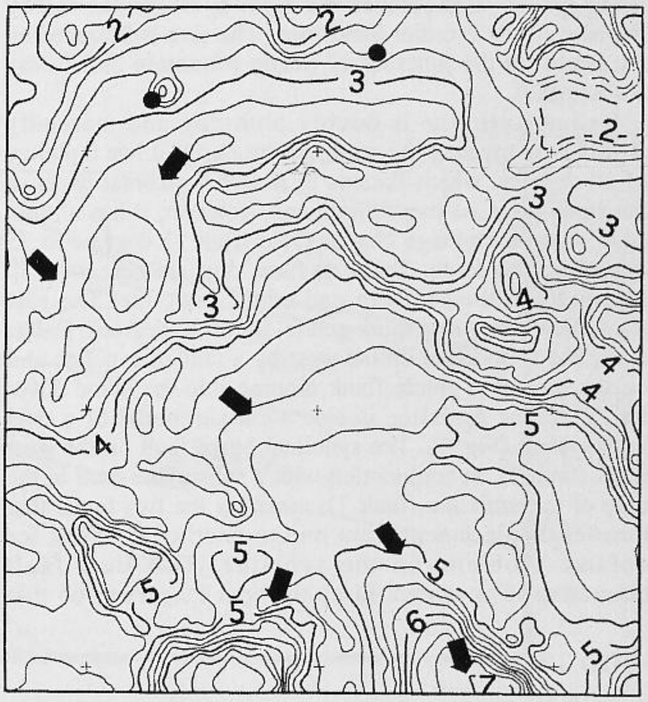

Fig. 4. Present-day bathymetry map of study area. Arrows show directions of present sediment transport. Transport has shifted from north to south (see Fig. 2) to generally northwest to southeast as shown in this figure. Contours = x1000 feet below sea level.

Structural Analysis of the En Echelon Arrangement of Dallas and Derby Anticlines, Wyoming

Troy Wayne Meinen

Dallas and Derby Anticlines are west-facing asymmetric, doubly plunging anticlines on a trend of folds along the eastern flank of the Wind River Mountains. The left-stepping offset of the fold trend between Dallas and Derby Anticlines is one of several small offsets along the fold trend. Characterization of the offset is therefore important to the understanding of offsets found elsewhere along the trend and in other parts of the Wyoming foreland. In addition, at the Dallas-Derby juncture, the style of the faulting changes from a dual fault system to a system of oppositely verging flank thrusts. Therefore the study of the geometries of Dallas and Derby Anticlines is important for the application of Laramide deformation models in the Wyoming foreland. This study describes the geometries of Dallas and Derby Anticlines, the offset between them, and relates their structural style to models of Wyoming foreland Laramide style deformation.

The study area is located in Fremont County, south of Lander, Wyoming. Its northern end is the middle of Dallas Anticline near the junction of state highway 28 and US highway 287. The area is then bounded on the west by a major syncline that parallels the highway and on the east by the base of the Frontier Sandstone. The area is bounded on the south by the juncture of Sheep Mountain and Derby Anticlines.

Dallas Anticline is doubly plunging and markedly asymmetric towards the west. To the north, it has a plunge of 10 degrees, which flattens to nearly horizontal north of the study area. At its southern end, however, it has a blunt end, where the plunge changes rate from 11 degrees to 35 degrees to the south. The west flank dips are near vertical, especially at the southern end of the anticline. The east flank, however, has a more gentle dip of 20 degrees. Dallas Anticline is bounded on the west by a faulted syncline and on the east by a gentle flank dipping into the Wind River Basin. Dallas Anticline is also a classic model of a dual fault system (Fig. 1). The synclinal hinge fault on the west flank (fault 2), in conjunction with a subsurface fault in the core of the anticline (fault 1), make up the two faults that transfer displacement from one to another to adjust for volume problems in the syncline. This dual fault characteristic is evidenced by the well Pure Fee #66 that drilled a normal section into the Upper Cambrian, an expanded Pennsylvanian and Permian section, and finally a normal Upper Cambrian section.

Derby Anticline also has west-facing asymmetry and is doubly plunging, but is more nearly symmetric than Dallas Anticline. Its plunge rate is 10 degrees to the north and 20 degrees to the south. Its west-flank dips range from 20 degrees to near 60 degrees, while its east flank has dips of 15 degrees to 20 degrees. At Derby Anticline, in contrast to Dallas Anticline, the west-flank syncline is not faulted. In addition, there is a reverse fault on the eastern flank that dies northward into a fold in the Cretaceous, and cuts across the south axis of Derby Anticline to die out in Lower Cretaceous rocks. In addition to these differences, Derby Anticline has a system of oppositely verging reverse faults as opposed to the dual fault system of Dallas Anticline (Fig. 2). The principal basement thrust (fault 1) is still southwest-vergent at Derby Anticline. The near symmetry of Derby Anticline, however, is caused by a northeast-vergent thrust (fault 3) that rises from bedding in the Permian out of the west-flank syncline and then repeats the Pennsylvanian, controlling the surface dips and plunge of Derby Anticline. The west-vergent back limb thrust (fault 4), is the next alternating reverse fault of the system.

The change in style of volume adjustment across the offset of the two systems is probably a result of the decreased tightness of the west flank syncline. At Dallas Anticline, the principal basement thrust (Fig. 1, fault 1) has greater heave and throw, and is farther to the west than at Derby Anticline. The location of fault 1 and the increase in heave and throw resulted in a tightened area that was accommodated with the dual fault system (Fig. 1, faults 1 and 2). South, at Derby Anticline, the tightness was not as severe so the fault on the west flank (Fig. 2, fault 3) was better able to accommodate crustal shortening.

The structural style found at Dallas and Derby Anticlines supports fold-thrust and thrust-uplift models of Wyoming foreland Laramide deformation. In addition, the study of the offset provides a model of an apparent en echelon offset of a fold trend that is not accomplished by true en echelon faulting or compartmental faulting.

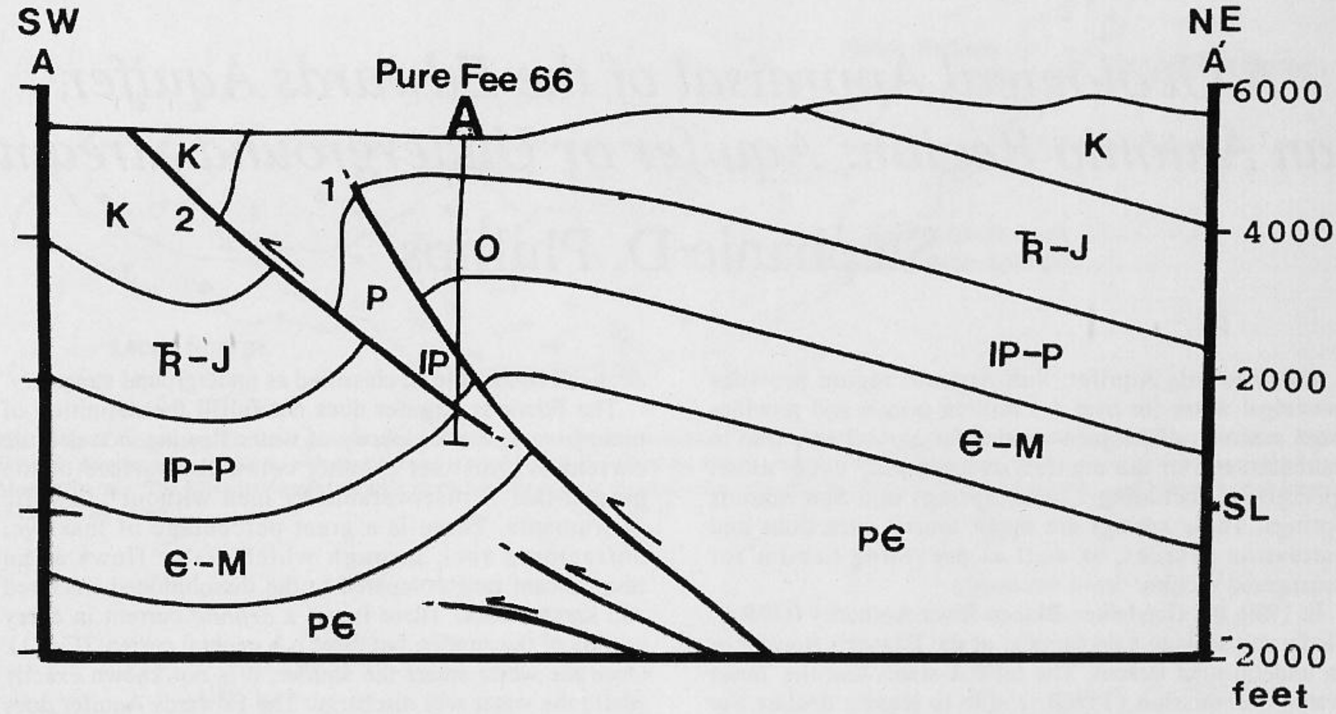

Fig. 1. Cross section A-A′ of Dallas Anticline through well Pure Fee #66 (sec. 13, T. 32 N., R. 99 W.) shows the anticline to fit the model of a fold-thrust structure. It has a controlling basement thrust (fault 1) and a synclinal hinge thrust (fault 2) that transfer displacement to accommodate excess volume in the syncline. There is no vertical exaggeration in the section and the elevations are relative to mean sea level (SL). Stratigraphic abbreviations are: K = Cretaceous, Ʀ-J = Triassic and Jurassic; IP-P = Pennsylvanian and Permian; Є-M = Cambrian, Ordovician, Devonian, Mississippian; PЄ = Precambrian crystalline basement.

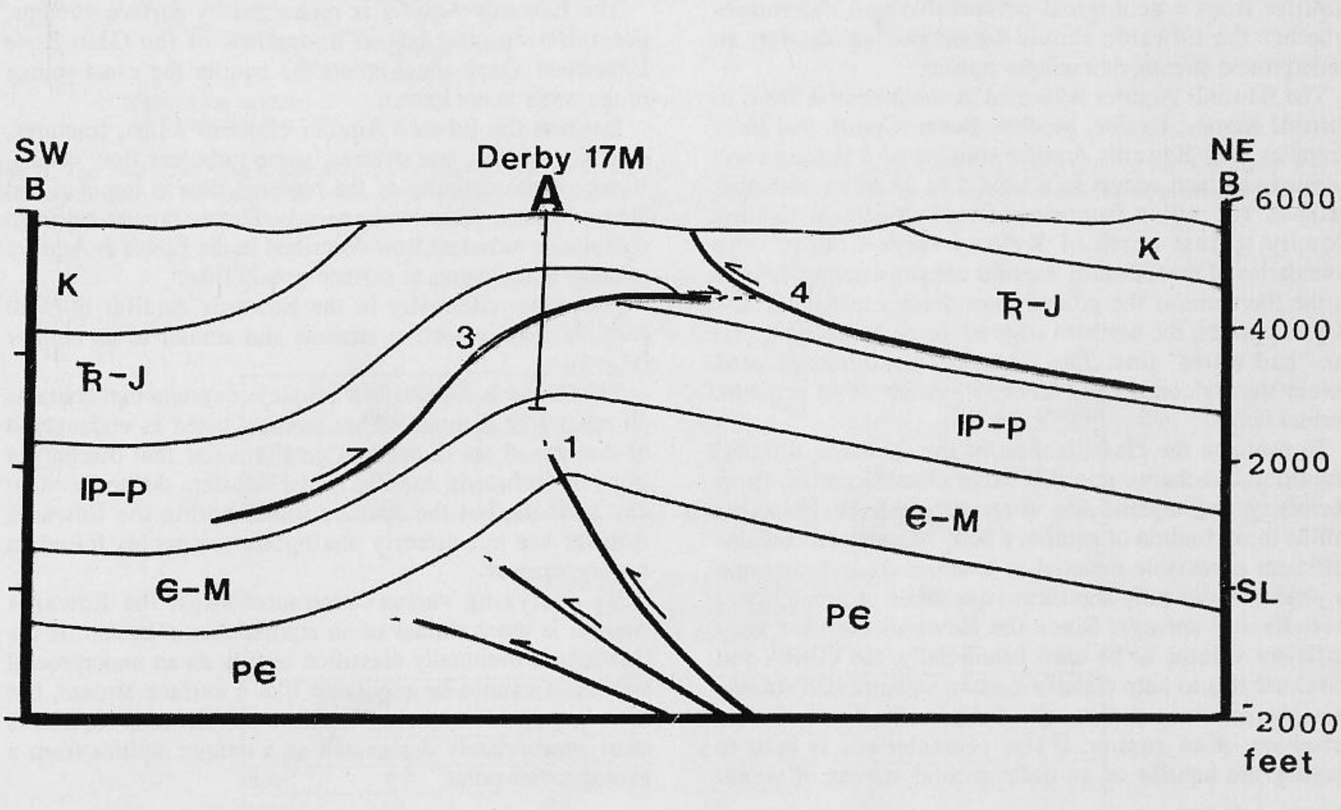

Fig. 2. Cross section B-B′ of Derby anticline through well Derby #17M (sec. 4, T. 31 N., R. 98 W.) shows the anticline to fit the thrust-uplift model. It has a system of oppositely verging thrusts for volumetric adjustments (faults 1, 3, and 4). There is no vertical exaggeration in the section and the elevations are relative to mean sea level (SL).

A Geological Appraisal of the Edwards Aquifer, San Antonio Region: Aquifer or Underground Stream?

Stephanie D. Phillips

The Edwards Aquifer, San Antonio region provides municipal water for over 1.3 million people and provides large amounts of irrigation water for agriculture. Due to increased use of the aquifer, overpumping could affect springflow, including Comal Springs and San Marcos Springs. These springs are major tourist attractions and recreation sources, as well as providing habitat for endangered species.

In 1989, the Guadalupe-Blanco River Authority (GBRA) filed a suit seeking a declaration of the Edwards Aquifer as an underground stream. The GBRA states that the Texas Water Commission (TWC) is able to legally declare the Edwards Aquifer an underground stream based on its similarities to surface stream characteristics. The main criteria used were: (1) well-defined and known boundaries, (2) definite source of supply, (3) current of water and known destination for flow of water, and (4) sufficient volume to be used beneficially.

This study evaluates the classification of the Edwards Aquifer from a geological perspective and determines whether the Edwards should be termed an aquifer, an underground stream, or a unique aquifer.

The Edwards Aquifer is located in south-central Texas in parts of Kinney, Uvalde, Medina, Bexar, Comal, and Hays Counties. The Edwards Aquifer consists of a recharge and artesian area and occurs in a band 5 to 30 miles wide that extends 175 miles from near Brackettville in Kinney County to just north of Kyle in Hays County. The boundaries of the Edwards Aquifer are consistently defined in the literature as the groundwater divides in Kinney and Hays Counties, the northern edge of the recharge zone, and the "bad water" line. The Edwards Aquifer is located within the Balcones Fault Zone, a system of en echelon, normal faults.

To evaluate the classification of the Edwards Aquifer, definitions, recharge and discharge characteristics, flow, chemistry, and aquatic life were analyzed. The Edwards fulfills the definition of aquifer, a body of rock that contains sufficient permeable material to conduct groundwater and to yield economically significant quantities of groundwater to wells and springs. Since the Edwards Aquifer is of sufficient volume to be used beneficially, the GBRA and TWC use this to help classify it as an underground stream. This characteristic is also one of the qualifications for the definition of an aquifer. If this characteristic is used to classify the aquifer as an underground stream, it would allow all aquifers to be classified as underground streams.

The Edwards Aquifer does not fulfill the definition of underground stream, a body of water flowing as a definite current in a distinct channel below the surface of the ground that is discoverable by men without scientific instruments. There is a great percentage of massive, unfractured rock through which water flows at an insignificant rate, compared to the dissolutioned fractured and karsted rock. There is not a *definite* current in every portion of the aquifer, but there is a general current (Fig. 1). Once the water enters the aquifer, it is not known exactly where the water will discharge. The Edwards Aquifer does not have a distinct channel. Since all channels are confined by banks and the boundary between a bank and floodplain is marked by a cessation of terrestrial vegetation, the Edwards Limestone does not have this type of boundary. The "well defined" boundaries of the Edwards Aquifer are neither discoverable without scientific instruments nor fixed in position (Fig. 2).

The Edwards Aquifer is recharged by surface streams, precipitation, and lateral underflow of the Glen Rose Limestone. Once water enters the aquifer the exact source of the water is not known.

Because the Edward Aquifer contains joints, fractures, solution cavities, and caverns, some turbulent flow occurs. However, the majority of the regional flow is laminar, and Darcy's Law can be applied. There is not enough continuous turbulent flow described in the Edwards Aquifer to make it analogous to surface stream flow.

The water chemistry in the Edwards Aquifer is more constant than in surface streams and similar to an aquifer (Fig. 3).

The Edwards Aquifer is a unique ecosystem that contains 40 species of animals. Other animals listed as endangered or threatened are dependent on the water that discharges from the Edwards Aquifer. Most aquifers do not contain any animals, but the species found within the Edwards Aquifer are not directly analogous to species found in surface streams.

By analyzing various characteristics, the Edwards Aquifer is more similar to an aquifer than a stream. If the Edwards is eventually classified legally as an underground stream, it cannot be regulated like a surface stream, but must be treated uniquely. Therefore the Edwards Aquifer is most appropriately designated as a unique aquifer from a geologic viewpoint.

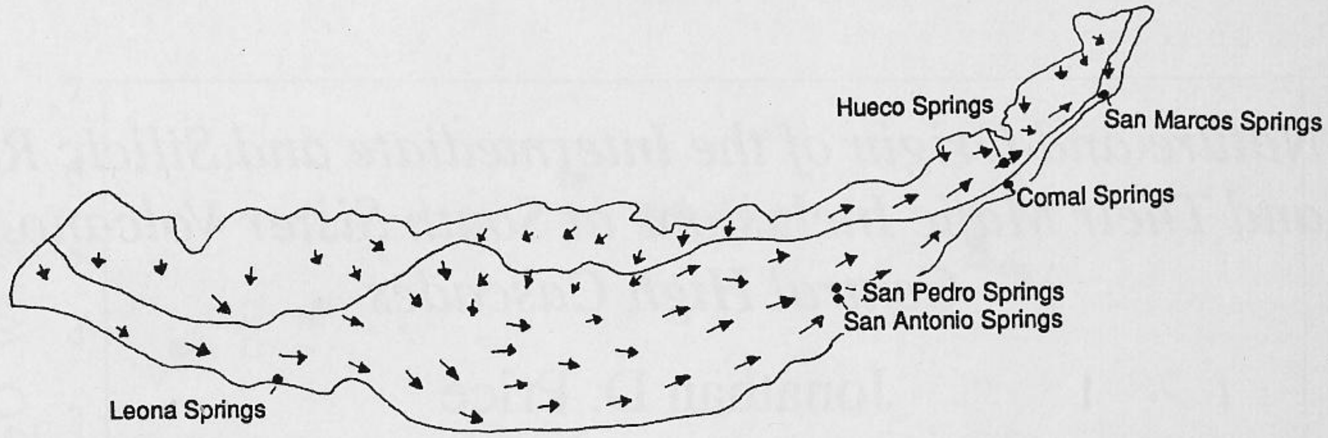

Fig. 1. The Edwards Aquifer has a general current flowing to the east and northeast. The six major springs that discharge from the aquifer are shown. If the Edwards Aquifer were similar to a surface stream, it would discharge all of its water at the "mouth" of the system which would be just northeast of San Marcos Springs. The Edwards Aquifer, though, discharges throughout the study area, including Leona Springs in the southwest, and Hueco Springs located in the recharge area.

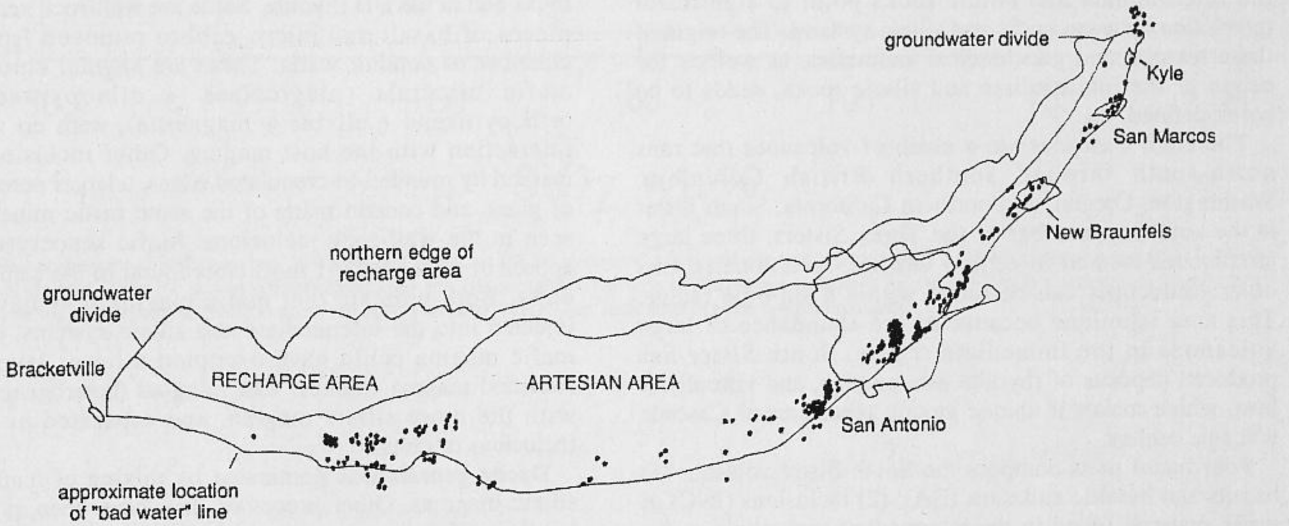

Fig. 2. The "bad water" line is only detected by the use of wells and chemical analyses. The well control shown within 5 miles of the "bad water" line is limited especially to the south. This line is also capable of shifting due to overpumping. The groundwater divides are not definite because these divides are able to shift due to climatic changes or overpumping stresses and these shifts can only be detected with scientific instruments.

Comal Springs, New Braunfels, Tx
Water Quality Data
(3-70 to 12-74)

	Temperature (deg C)	Calcium (mg/l)	Sulfate (mg/l)	Chloride (mg/l)	Dissolved Solids (mg/l)
range	0.8	7	3	3	12
max.	23.9	82	25	15	302
min.	23.1	75	22	12	290

Colorado River below Austin, Tx
Water Quality Data
(10-87 to 9-91)

	Temperature (deg C)	Calcium (mg/l)	Sulfate (mg/l)	Chloride (mg/l)	Dissolved Solids (mg/l)
range	19	19	62	97	233
max.	27	67	100	140	532
min.	8	48	38	43	299

Fig. 3. Comparison of water chemistry at Comal Springs to the Colorado River. The water discharging from Comal Springs is more constant and representative of groundwater when compared to the great variance in temperature and chemistry of surface water.

The Nature and Origin of the Intermediate and Silicic Rocks and Their Mafic Inclusions at South Sister Volcano, Central High Cascades

Jonathan D. Price

South Sister, a large stratocone located in the central Oregon High Cascades, exhibits geochemical variation that indicates complex magma genesis. Textural anomalies in the intermediate and silicic rocks point to significant interaction between mafic and silicic systems. The origin of these textural and geochemical anomalies, as well as the origin of the intermediate and silicic rocks, needs to be better defined.

The High Cascades are a chain of volcanoes that runs north-south through southern British Columbia, Washington, Oregon, and northern California. South Sister is the southern member of the Three Sisters, three large stratocones located in central Oregon. Additionally, two other stratocones can be found within a ten-mile radius. This area is unique because of the abundance of large volcanoes in the immediate region. South Sister has produced deposits of rhyolite ash, pumice, and vitrophyric lava, which makes it unique among other central Cascade volcanic centers.

Four major units compose the South Sister volcano: (1) basalts and basaltic andesites (BA); (2) inclusions (INC) of mafic material found in the intermediate and silicic rocks; (3) intermediate rocks, including andesites and dacites (AD); and (4) rhyolites from Rock Mesa (RM) and Devil's Hill (DH). Eruption of basalt occurred early in the history of the volcano, forming the large shield that underlies the volcanic edifice. Basaltic andesite constructed the early cone, and basaltic cinder cones formed to the south and southwest of the central vent locality. This was followed by andesite and dacite eruptions that constructed much of the main cone. The following main cone eruptions of basaltic andesite resulted in the construction of the two summits, while basalt activity included one cinder cone to the southwest. The most recent eruptions were the hemihyaline rhyolite flows and domes of Rock Mesa and the Devil's Hill Chain.

There is a prominent gap in silica (68–71% SiO_2) (Fig. 1), and a corresponding gap in rubidium (48–74 ppm Rb) found in the variation diagrams of this suite. Such gaps often indicate non-genetic relationships in the evolution of a system. Additional evidence is found in sodium compositions, where soda is enriched and then depleted. Such a trend is expected when fractionation of feldspar deplete soda concentrations. However, this system never developed sodic feldspar. Fractional crystallization may have played only a small role in the evolution of this system; magma mixing, crustal assimilation, and partial melting may have been major influences.

Mafic inclusions are found in many of the intermediate rocks and in the DH rhyolite. Some are wall-rock xenoliths, pieces of basalt and micro-gabbro removed from the chamber or conduit walls. These are angular clusters of mafic minerals (plagioclase + clinopyroxene ± orthopyroxene ± olivine ± magnetite), with no visible interaction with the host magma. Other inclusions are marked by rounded-to-crenulated edges, a larger percentage of glass, and contain many of the same mafic minerals as seen in the wall-rock inclusions. Mafic xenocrysts also appear in small (0.75–1 mm) clots found in the same rock units. Both indicate that mafic magma may have ben injected into the intermediate and silicic systems, or that mafic magma could have occupied a lower layer in a stratified magma chamber that mingled (heterogeneously) with the more silicic magma, and separated as small inclusions or clots.

Dacite genesis was dominated by mixing of mafic and silicic magmas. Other processes may have been at work, but their effect is not seen in the products, aside from mafic inclusions and reversely zoned crystal phases. Dacite consists of two geochemical types, based on variations in yttrium and niobium. Dacites with lower concentrations of these elements may have a closer genetic relationship to the mafic rocks, while other dacites were contaminated by sources enriched in these elements.

Assimilation-fractional crystallization modeling of the DH rhyolite suggests that these magmas were crustal melts slightly modified by fractional crystallization (Fig. 2). This model involved generating the DH magma from a dacite magma through partial assimilation of granitic crustal bodies similar to those found in the Klamath Mountains, and as xenoliths found in intermediate rocks at Crater Lake. What small amounts of fractional crystallization occurred left a pronounced effect on the heavy rare earth elements. These were fractionated into developing crystals of apatite and zircon.

Using major and trace element models, it can be determined that RM rhyolite evolved by 7% fractional crystallization of the DH rhyolite. It is possible that this resulted in a stratified chamber that first erupted the older, more evolved RM rhyolite, and then erupted the underlying DH magma.

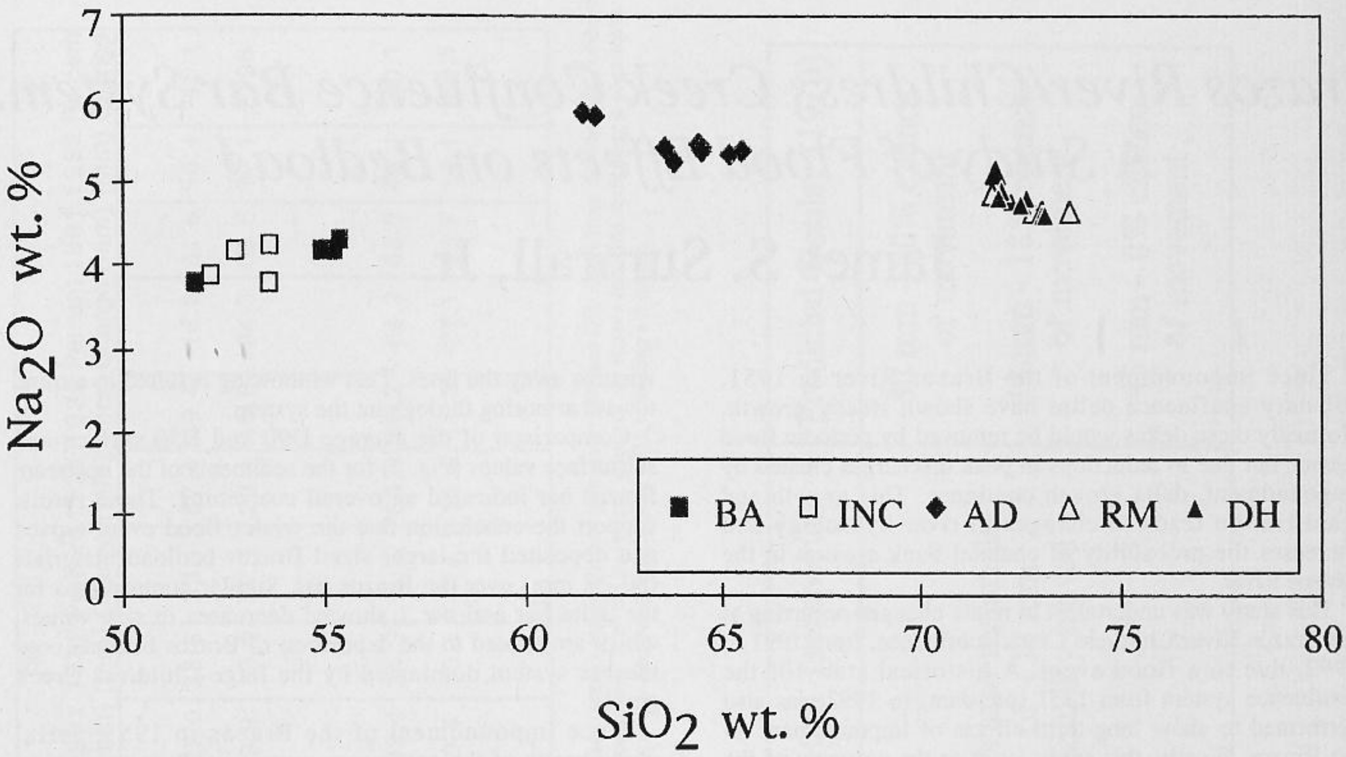

Fig. 1. Harker diagram of soda vs. silica. BA is basaltic andesite, INC is inclusion, AD is andesite and dacite, RM is Rock Mesa rhyolite, and DH is Devil's Hill Chain rhyolite. Note the gap in silica between AD and DH samples. This type of soda trend is usually the result of soda fractionation into feldspar; however, no sodic feldspar was found in these samples. This trend is therefore indicative of open system processes.

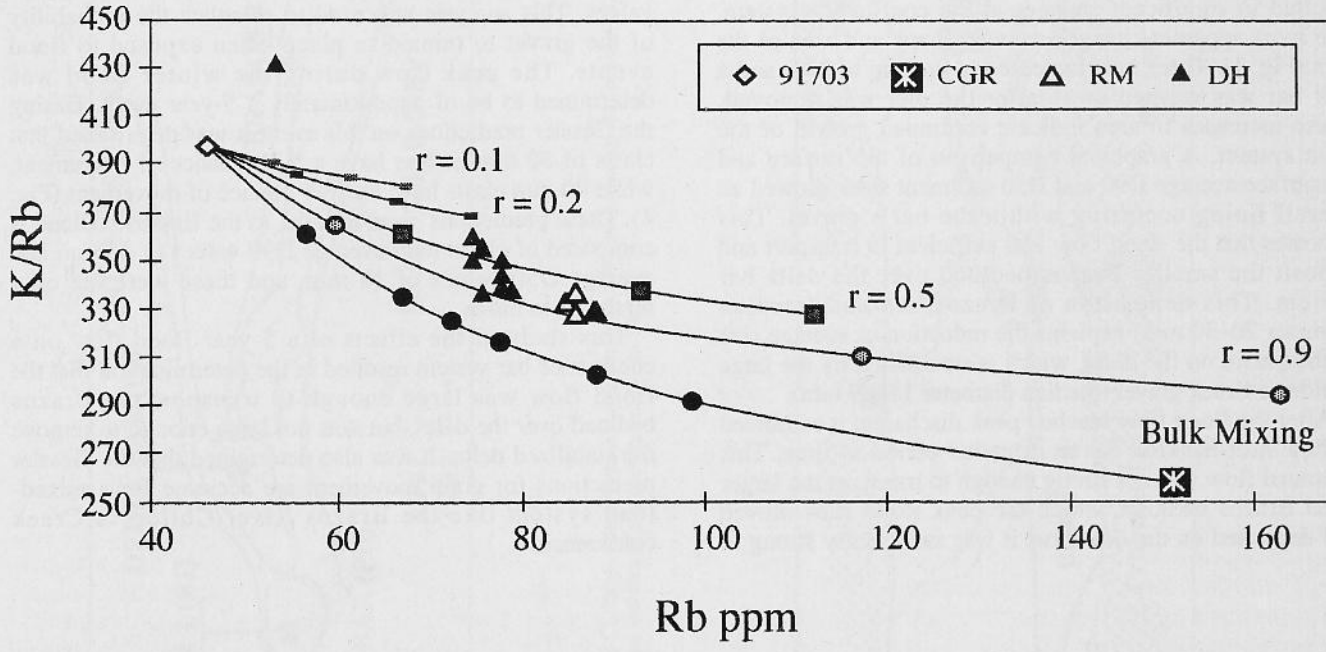

Fig. 2. Assimilation-Fractional Crystallization (AFC) diagram illustrating a possible model for the petrogenesis of Devil's Hill Chain (DH) rhyolite at South Sister. Crustal assimilant is CRG, a granite xenolith found at Crater Lake. Diagram shows that assimilation, coupled with minimal amounts of fractional crystallization, would result in magmas of DH composition.

Brazos River/Childress Creek Confluence Bar System: A Study of Flood Effects on Bedload

James S. Sumrall, Jr.

Since impoundment of the Brazos River in 1951, tributary confluence deltas have shown steady growth. Normally these deltas would be removed by periodic flood events, but due to reductions in peak discharges created by impoundment, delta growth continues. This growth and stabilization leads to changes in river hydrology and increases the probability of channel bank erosion in the Brazos River.

This study was undertaken to relate changes occurring at the Brazos River/Childress Creek confluence, from 1991 to 1992, due to a flood event. A historical study of the confluence system from 1951 (pre-dam) to 1992 was also performed to show long-term effects of impoundment on the Brazos. Finally, this study looks at the accuracy of the Gessler predictions for grain movement as related to bar materials and the winter flood event.

The study area includes the Brazos River/Childress Creek confluence. Childress Creek enters the Brazos River 22 miles downstream from Lake Whitney Dam and 16 miles upstream from Waco, Texas. Childress Creek's basin is bounded to the north by the Brazos basin and to the south by the North Bosque River basin.

The long period of continual near-bankfull flow on the Brazos River during the winter of 1991 and spring of 1992 resulted in significant changes at the confluence system. The most apparent changes were in shape and area of the bars (Fig. 1). Three bars increased in area up to 23% and a pier bar was washed away after the pier was removed. These increases in area indicate continued growth of the delta system. A graphical comparison of the surface and subsurface average D90 and D50 sediment sizes showed an overall fining occurring within the bar's gravel. This indicates that the flood flow was sufficient to transport and deposit the smaller Brazos bedload over the delta bar system. This deposition of Brazos bedload (median diameter 20–30 mm) explains the reduction in average size of the gravel on the delta, which is dominated by the large Childress Creek gravel (median diameter 15–89 mm).

After the flood flow reached peak discharge, it continued to flow near-bankfull for an extended period of time. This continual flow was not strong enough to transport the larger sized Brazos bedload, which the peak flood flow moved and deposited on the delta, but it was sufficiently strong to winnow away the fines. This winnowing resulted in a trend toward armoring throughout the system.

Comparison of the average D90 and D50 surface and subsurface values (Fig. 2) for the sediments of the upstream Brazos bar indicated an overall coarsening. These results support the conclusion that the winter flood event moved and deposited the larger sized Brazos bedload materials (30–38 mm) over the Brazos bar. Similar comparisons for the delta bar and bar 1 showed decreases in size values, which are related to the deposition of Brazos bedload onto the bar system dominated by the large Childress Creek gravel.

Since impoundment of the Brazos in 1951, aerial photography of the confluence system made it possible to track delta growth, and the resulting channel bank loss (Fig. 3). In 1964 the delta measured 64,000 sq. ft. and did not result in erosion of the Brazos River channel. In 1992, the delta measured 108,000 sq. ft. and resulted in Brazos channel bank loss of 17,200 sq. ft. Without a major flood discharge to remove the gravel, the trend of this delta system seems to be continued delta growth and Brazos channel bank loss.

A Gessler prediction for grain movement was performed on the confluence system using the average surface gravel values. This analysis was used to calculate the probability of the gravel to remain in place when exposed to flood events. The peak flow during the winter flood was determined to be of approximately a 5-year event. Basing the Gessler predictions on this event it was determined that clasts of 30 mm or less have a 50% chance of movement, while 40 mm clasts have an 18% chance of movement (Fig. 4). These predictions seem to hold, as the Brazos bedload is composed of clasts with average D90 values of 40 mm and average D50 values of 18 mm, and these were the only materials to move.

This study of the effects of a 5-year flood flow on a confluence bar system resulted in the determination that the flood flow was large enough to transport the Brazos bedload over the delta, but was not large enough to remove the stabilized delta. It was also determined that the Gessler predictions for grain movement are accurate for a mixed-load system like the Brazos River/Childress Creek confluence.

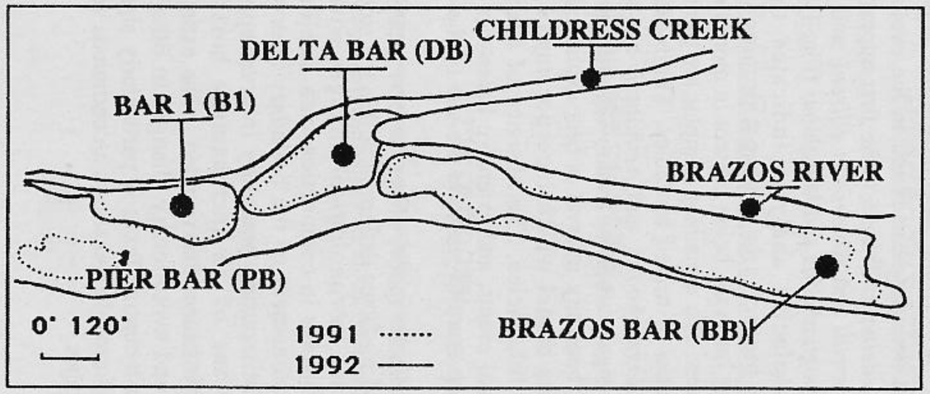

Fig. 1. Map of the Childress Creek/Brazos River confluence system showing the location of the bars and the changes in shape and area of the bars from 1991 to 1992.

		BB			DB			B1		
		1991 (mm)	1992 (mm)	Percent Change	1991 (mm)	1992 (mm)	Percent Change	1991 (mm)	1992 (mm)	Percent Change
ARMOR	D90	27.9	38.8	39.1	57.4	49.4	-16.2	81.8	56.4	-45.1
	D50	9.9	13.6	37.1	20.4	10.5	-94.3	50.6	17.1	-196
SUBARMOR	D90	33.1	34.3	3.7	66.1	40.2	-64.2	68.3	48.4	-40.1
	D50	9.4	10.1	7.4	14.4	11.3	-27.1	23.7	14.1	-68.1

Fig. 2. Average surface (armor) and subsurface (subarmor) D90 and D50 values from 1991 to 1992, of the three bars of the confluence system. Note the increasing values of the Brazos bar (BB), indicating a coarsening, and the decreasing values of the delta bar (DB) and bar 1 (B1), indicating a thinning.

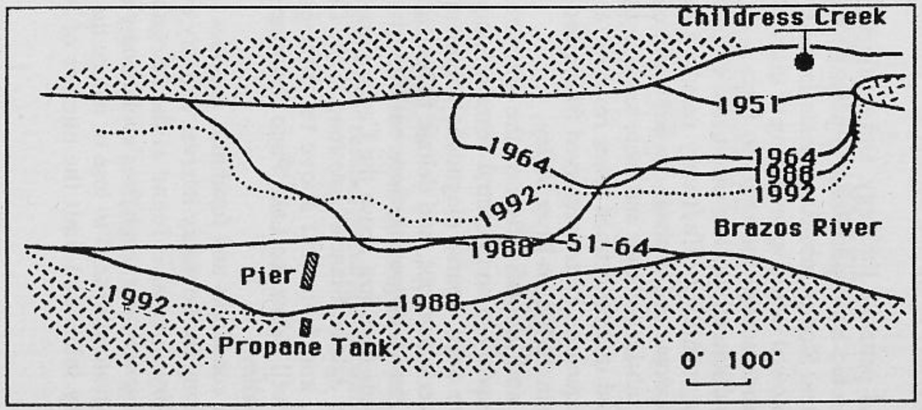

Fig. 3. Historical overview of the delta growth and bank loss at the confluence from 1951, pre-dam, to 1992.

Actual Transported	Predicted Gessler (1970)
Brazos materials of <30 mm actually moved	30 mm - 50% chance of movement 40 mm - 18% chance of movement 45 mm - 0% chance of movement

Fig. 4. Analysis of Gessler's (1970) predictions for grain movement as related to the Brazos River bedload.

Stratigraphy and Conodont Biostratigraphy of the Desmoinesian/Missourian (Pennsylvanian) Boundary Interval, Colorado and Brazos River Valleys, North-Central Texas

Jeffrey C. Turner

The placement and geologic interpretation of the Desmoinesian/Missourian boundary in Texas is controversial. Conodont biostratigraphy in conjunction with fusulinid data and facies analysis refine the location and placement of the boundary interval and provide data to test stratigraphic models.

The study area extends from the city of Mineral Wells on the northeastern edge of Palo Pinto County and terminates at the Colorado River near the township of Winchell in Brown County. The area of interest is no more than two-to-three miles wide, but is approximately 120 miles long.

The Desmoinesian/Missourian boundary in north-central Texas occurs within the Strawn Group, which consists of thick terrigenous clastic facies and relatively thin carbonate facies of late Atokan to early Missourian age. The Strawn Group crops out in two roughly triangular areas, the Brazos River Valley and the Colorado River Valley. These strata have been interpreted as fluvial, deltaic, and shallow marine facies.

Depositional patterns, lithology, and stratigraphy were primarily affected by depositional processes and basin subsidence rates. Slower subsidence rates in the Colorado River Valley allowed extensive reworking of distributary and mouth bar sands across the shelf to form a relatively condensed, multilateral section within the upper Strawn. The resulting turbidity effectively resulted in fewer diagnostic conodonts and limited, discontinuous carbonate units. Greater subsidence rates and more terrigenous clastic influx deposited in well-defined lobes resulted in thicker, multistoried sequences, better preserved facies, and a more marine section in the Brazos River Valley.

Definition and identification of the Desmoinesian/Missourian boundary in north-central Texas is hampered by few correlative units, sparse diagnostic conodonts, long distances between outcrops, and deltaic facies containing few fossils. The stratigraphic interval containing the boundary extends from just above the Capps Limestone to well below the Adams Branch Limestone in the Colorado River Valley, and from just above the Village Bend Limestone to well below the Lake Pinto Sandstone in the Brazos River Valley.

Use of both conodonts and fusulinids provides the best means for limiting the boundary interval. Primary evidence for the boundary interval is found at the *Idiognathodus delicatus* to *Idiognathodus magnificus* evolutionary change. Advanced idiognathodids tend to lose or reduce the size of their accessory lobes and limit the number of rows of transverse ridges. This is possibly due to the tendency to retain juvenile characteristics into adulthood in the Late Pennsylvanian. The transverse ridges of the advanced specimens are not developed as far anteriorly on the platform as those in older populations. Late Desmoinesian individuals rarely possess more than six transverse ridges. As the phylogeny develops, the specimens become smaller and less robust. However, the key to the sequence and recognition of the Missourian evolutionary event lies in the attitude of the rostral ridges and adcarinal grooves. Primitive specimens exhibit narrow adcarinal grooves and gently sloping rostral ridges. Advanced individuals of the late Desmoinesian show wide adcarinal grooves that flare anteriorly. Near the boundary, the rostral ridges become quite steep and terminate in a near vertical precipice. These changes result in specimens that resemble primitive neognathodids and may indicate a homeomorphic replacement of the extinct neognathodids. Supporting evidence for the boundary can found in the evolutionary change from *Beedeina* to *Triticites,* the first appearance of *Adetognathus merrilli* and *Hindeodus ellisoni,* and the last appearance of *Neognathodus* and *Mesolobus* (Fig. 1).

Conodont biofacies analysis indicates that the *Idiognathodus* biofacies is the most common of the conodont-bearing intervals, being present in more than one-half of the samples. The remaining samples are primarily of the *Adeotognathus* or mixed biofacies. The biofacies of *Hindeodus, Diplognathodus,* and *Aethotaxis* are strongly associated with algal carbonate banks. The lithofacies and biofacies of the boundary interval is best explained by the delta lithofacies model which incorporates conodont biofacies with lithofacies. The absence of deep water deposits, erosional events, and extensive limestone units in the Strawn Group seem to preclude the use of the eustatic sea level model.

The delta lithofacies model offers the best interpretation for boundary and biofacies definition within the study area. The evolutionary morphological changes within the *Idiognathodus* lineage in conjunction with fusulinid data offer the best refinement of the boundary interval. The Desmoinesian/Missourian boundary interval appears to record conditions of an increasingly hostile and deteriorating environment that resulted in the extinction of several genera and evolutionary changes in others. This boundary in north-central Texas is particularly significant because it is conformable and shows a continuous sequence of conodont lineages.

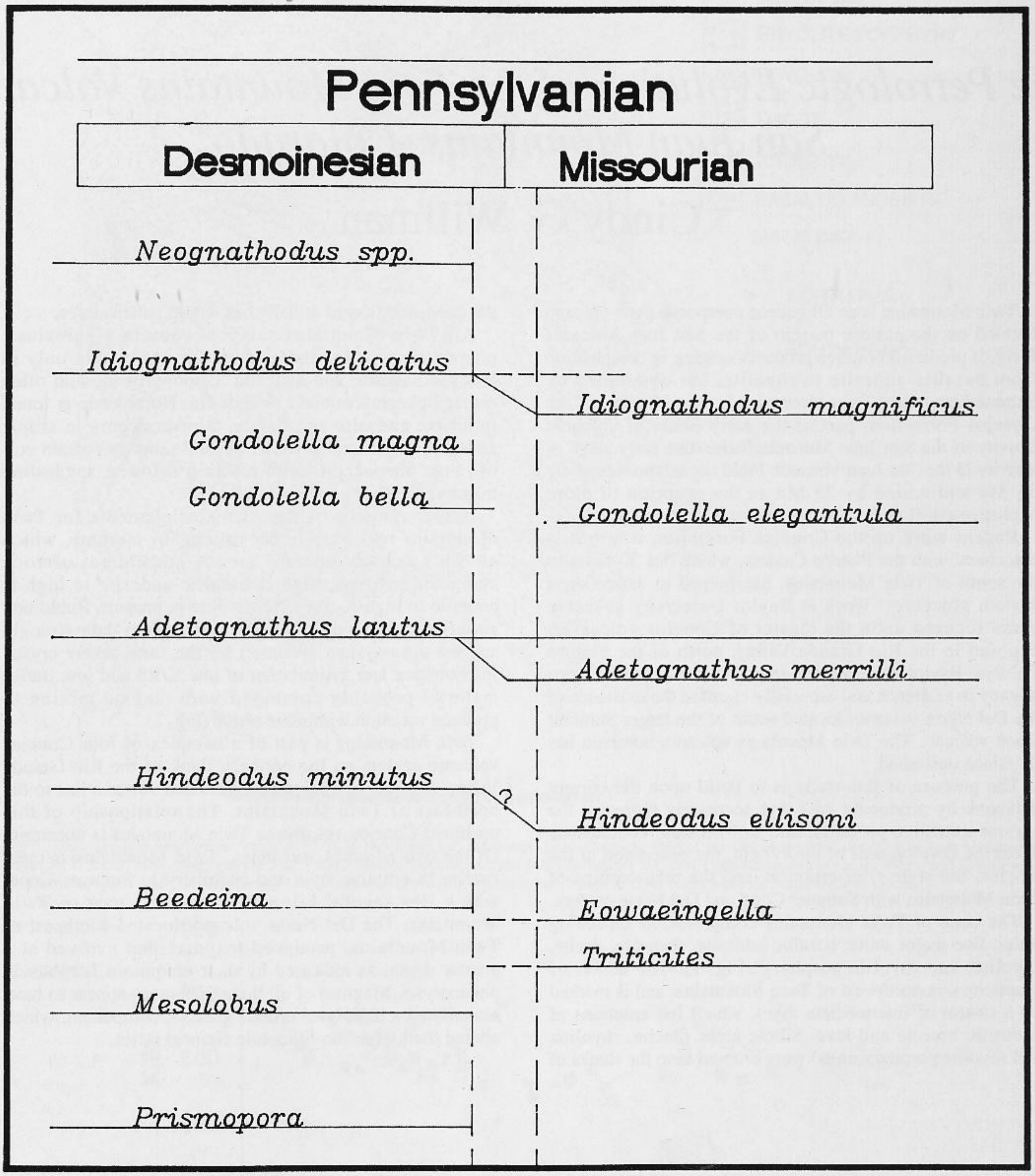

Fig. 1. Ranges and speciation events for taxa that have been used to locate the Desmoinesian/Missourian boundary interval in north-central Texas.

The Petrologic Evolution of the Twin Mountains Volcano, San Juan Mountains, Colorado

Cindy G. Willman

Twin Mountains is an Oligocene composite cone volcano located on the eastern margin of the San Juan Volcanic Field. It produced eruptive products ranging in composition from basaltic andesite to rhyolite, but dominated by intermediate rocks. Twin Mountains is representative of the Conejos Formation, part of the early phase of volcanic activity in the San Juan Volcanic Field. The early stage of activity in the San Juan Volcanic Field began approximately 35 Ma and ended by 29 Ma as the eruption of more explosive ash-flow tuffs began to dominate the region.

Recent work on the Conejos Formation where it is associated with the Platoro Caldera, which lies 30 miles to the south of Twin Mountains, has helped to define open system processes. Work at Baylor University in recent years focused upon the cluster of Conejos volcanism exposed in the Rio Grande Valley, north of the Platoro Caldera. Baylor studies have amplified the excellent maps already in existence and especially clarified the existence of the Del Norte volcano, located south of the larger Summer Coon volcano. The Twin Mountains volcano, however, has remained unstudied.

The purpose of this study is to build upon the current databank by producing data that accurately represent the various extrusive products, and to formulate conclusions about the development of the system, the generation of the magma, the style(s) of eruption, and the relationships of Twin Mountains with Summer Coon and Del Norte centers.

The cone of Twin Mountains comprises, in ascending order, five major units: basaltic andesite, andesite, dacite, rhyolite, and rhyolite porphyry (Fig. 1). The center of eruptions was northwest of Twin Mountains, and is marked by a swarm of intermediate dikes, which fed eruptions of andesitic breccia and lava. Silicic lavas (dacite, rhyolite, and rhyolite porphyry units) were erupted onto the flanks of the cone, and flowed radially out within paleovalleys.

All Twin Mountains samples contain plagioclase, magnetite, apatite, and zircon. Olivine occurs only in basaltic andesite and andesite. Clinopyroxene and often orthopyroxene are found in andesite. Hornblende is found in silicic andesite and dacite. Biotite occurs in silicic andesite, dacite, and rhyolite. Several samples contain out-of-range phenocrysts with reaction textures, suggesting magma mixing.

Harker plots of oxides of major elements for Twin Mountains rocks are linear except for alumina, which shows a convex-upward arc. A continuum of rock compositions from high-K basaltic andesite to high-K andesite to high-K dacite to rhyolite is present. Rubidium, strontium, zirconium, yttrium, and barium data strongly suggest open-system evolution for the suite, where crystal fractionation and assimilation of low K/Rb and low Ba/Rb material probably combined with magma mixing to produce variation within the series (Fig. 2).

Twin Mountains is part of a complex of four Conejos volcanic centers on the northern flank of the Rio Grande Valley. A large, unstudied and unnamed volcano lies to the northeast of Twin Mountains. The relationship of this unnamed Conejos volcano to Twin Mountains is uncertain. Of the two adjacent volcanoes, Twin Mountains is most similar in eruptive style and chemistry to Summer Coon, which lies several kilometers directly east of Twin Mountains. The Del Norte volcano, located southeast of Twin Mountains, produced magmas that evolved at a greater depth, as indicated by their ubiquitous hornblende phenocrysts. Magmas of all three volcanoes appear to have assimilated a high-level crustal gneissic component, which altered their large-ion-lithophile element ratios.

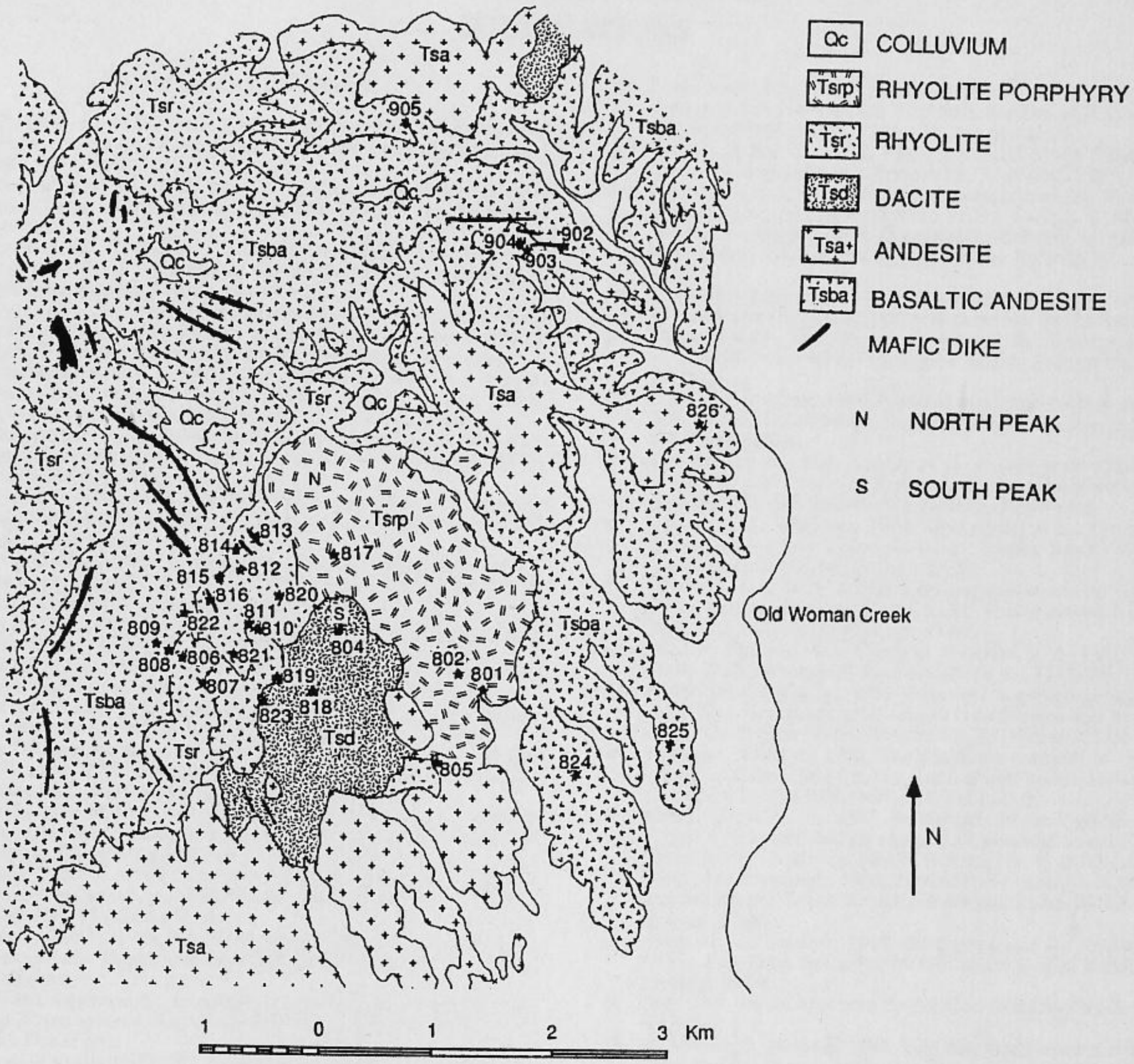

Fig. 1. Geologic map and sample localities of Twin Mountains (after Lipman, 1976, Geologic Map of the Del Norte area, eastern San Juan Mountains, Colorado: U.S.G.S. Misc. Investigation, Map I-952). Sample localities marked with filled stars ("92" prefix omitted to simplify map). North and South Peaks are identified by "N" and "S" respectively. Old Woman Creek roughly separates the Twin Mountains center from the Summer Coon center.

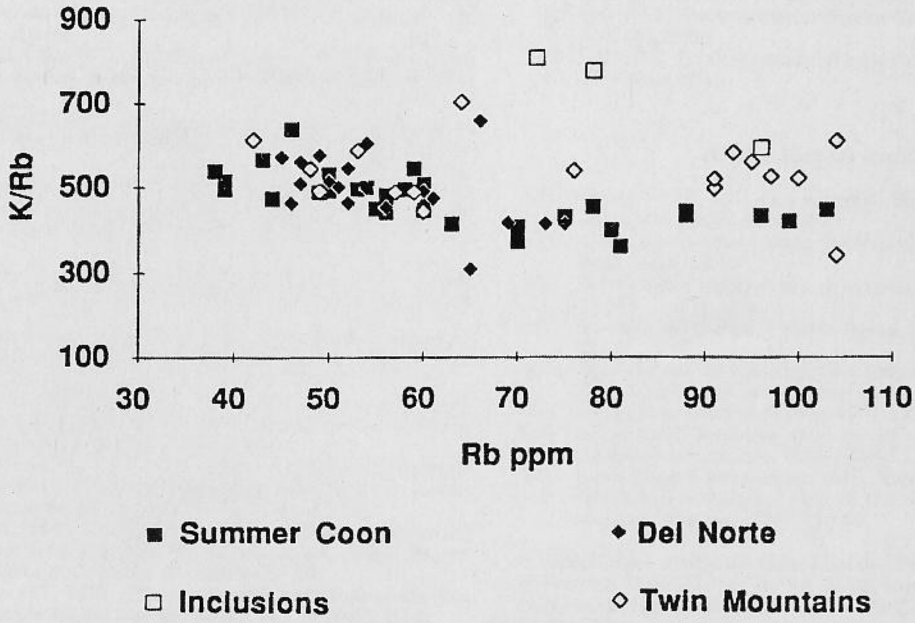

Fig. 2. A Harker plot of K/Rb versus Rb for Summer Coon, Del Norte, and Twin Mountains. If fractionation were the only process at work in the system then this trend would be a flat, horizontal line. Deviance from a flat line implies that other processes, such as assimilation and magma mixing, must also be at work in the system. The degrees to which fractionation, assimilation, and mixing were actively affecting the system can only be estimated from this plot, but fractionation still appears to be the dominant influence.

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